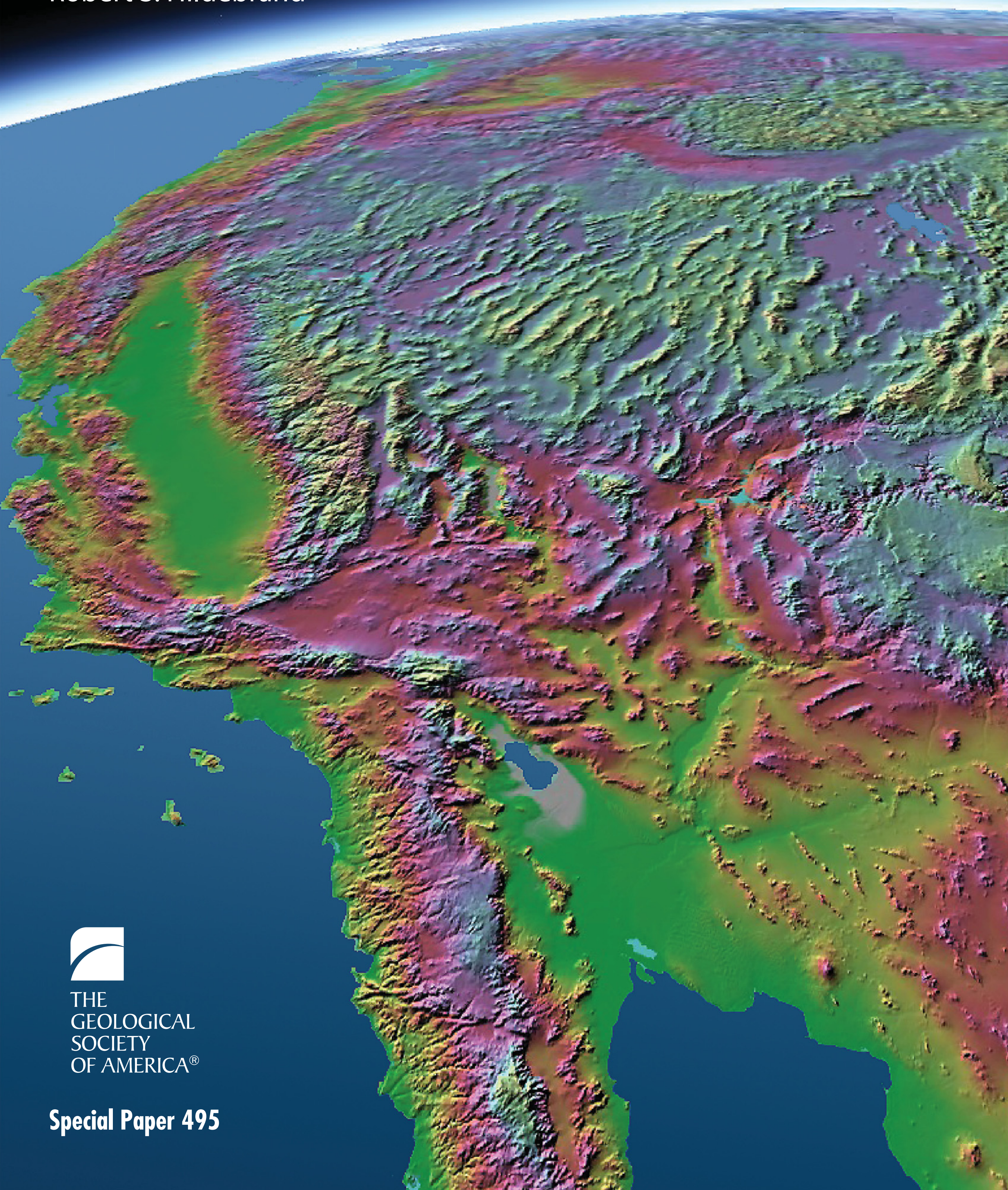
Mesozoic Assembly of the North American Cordillera

Robert S. Hildebrand



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Special Paper 495

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The North Face of Half Dome, Yosemite National Park, California, was carved from Half Dome granodiorite, a 92 Ma pluton of the Sierran batholith, one of the great Cretaceous Cordilleran-type plutonic belts of the western Americas.

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Robert S. Hildebrand
Department of Geology
University of California
One Shields Avenue
Davis, California 95616-8605
USA

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Cover: Three-dimensional shaded relief map of the southwestern United States and northern Mexico. The illustration was produced using shuttle radar data processed by Ray Sterner at The Johns Hopkins University Applied Physics Laboratory. It was then wrapped on the appropriate area in Google Earth®.

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Mesozoic Assembly of the North American Cordillera

Robert S. Hildebrand
Department of Geology, University of California, One Shields Avenue, Davis, California 95616-8605, USA

ABSTRACT

The broadly accepted hypothesis for the development of the segmented Cordilleran orogen above a long-lived eastwardly dipping subduction zone is at odds with many critical observations. Therefore, I explore an alternative collisional model in which the western edge of North America was partially subducted to the west beneath the Rubian ribbon continent. The collision of the two initially led to the localized Sevier fold-thrust belt and later to the more extensive Laramide deformational event.

The Rubian ribbon continent was assembled piece by piece, but at 160 Ma, two previously assembled blocks, Sierrita and Proto-Rubia, collided. Proto-Rubia formed during the Mississippian by collision of the Roberts Mountain allochthon with the Antler margin, a Neoproterozoic–Paleozoic passive margin of unknown provenance. Additions at 260–250 Ma included Yukon-Tanana–Slide Mountain terranes and the Golconda allochthon. Sierrita formed during the Middle Jurassic between ~170 and 160 Ma when several east-facing arcs, including the Smartville, Slate Creek–Lake Combie, and Hayfork, were amalgamated on the western side of the Sierran–Black Rock arc just prior to and at about the same time as it collided with the western margin of Proto-Rubia to the east. Consequent slab failure generated an arc-parallel suite of postcollisional intrusions, including the Independence dike swarm and the bimodal, alkaline Ko Vaya suite. New eastward subduction beneath the western margin of Rubia started sometime between 159 Ma and ~130 Ma.

The Sevier phase of the Cordilleran orogeny began at ~125 Ma when a promontory located in the Great Basin segment of the North American craton was pulled into the westward-dipping subduction zone that existed on the Panthalassic side of the Rubian superterrane. The entry of the margin into the trench formed the Sevier fold-thrust belt and led to accretion of exotic megathrust sheets to western North America. During(?) and after the collision, most of the Rubian superterrane migrated southward relative to North America. To the west at ~100 Ma, a dextral transpressional collision led to the closure of the Gravina-Nutzotin-Dezadeash-Gambier basin(s) in Canada and Alaska, the accretion of the Alisitis arc in Baja California, and the closure of a now cryptic basin within the Sierra Nevada. Post-collisional plutonic suites, such as the La Posta and the Sierran Crest may have been caused by slab failure.

At around 80 Ma, North America started to migrate southward, and this led to the collision of the entire Rubian ribbon continent, which extended from the Alaskan sector at least to northern South America, with the outboard margin of North America during the Laramide phase of the Cordilleran orogeny. The resultant shutdown of

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subduction along both margins led to (1) termination of Cordilleran-type magmatism, (2) exhumation of Franciscan blueschists within accretionary complexes along the western side of Rubia, (3) emplacement of a linear belt of slab-failure magmatism within the Sonora-Mojave region and the then adjacent Coast plutonic belt, and (4) oblique northward migration of Rubia. The oblique convergence between Rubia and North America created a region of thick-skinned deformation within the Great Basin segment south of the Orofino fault and thin-skinned folding and thrusting in the Alaskan, Canadian, and Sonoran sectors. Prior to their northward migration, the previously amalgamated terranes presently located within the Canadian Cordillera were located several thousand km farther south and joined at their south end with the northern end of the Sonoran sector.

The collision was followed by linear, orogen-parallel regions of extensional collapse and exhumation. Final collapse of the thickened collision zone to form the Basin and Range province occurred in the Great Basin and Sonoran sectors where the North American craton had been pulled beneath the Rubian superterrane. New easterly directed subduction started beneath the amalgamated collision zone at ~53 Ma and has continued to the present.

INTRODUCTION

Cordilleran orogens are peculiar in that they are interpreted to be collisions between two opposites: dense, low-lying, lower-plate oceanic lithosphere and less dense, buoyant upper-plate continental lithosphere. Ideas about their development started forty years ago when Hamilton (1969a, 1969b) proposed the modern volcanic Andes as an actualistic model for the great batholiths of North America. Soon the rush was on as Hsü (1971) realized that the Franciscan mélange represented an ancient subduction complex; Moores (1969, 1970) understood that ophiolites are pieces of oceanic crust and subjacent mantle; Ernst (1970) suggested that the Great Valley–Franciscan contact marked a Mesozoic Benioff zone; Dickinson (1970) connected arc magmatism, batholiths, and fore-arc sedimentation to subduction; and Dewey and Bird (1970) coined the term, Cordilleran orogen, to describe such belts. Ever since, most geologists have assumed that the western edges of the American continents formed the upper plates above east-dipping oceanic plates, either starting during the Late Devonian (Burchfiel and Davis, 1972, 1975; Price, 1981; Monger and Price, 2002; Dickinson, 2000, 2004, 2006; Colpron et al., 2007) or the Triassic (Schweickert and Cowan, 1975; Ingersoll, 2008). In both interpretations it is entirely the convergence and interactions between lower oceanic and upper continental plates that creates Cordilleran orogens with their voluminous Cordilleran-type batholiths, local areas of doubly-thickened crust, hinterland metamorphism, and retro-arc fold-thrust belts (Armstrong, 1974; Coney and Reynolds, 1977; Dickinson and Snyder, 1978; Saleeby, 2003; DeCelles, 2004; DeCelles et al., 2009). This overall Cordilleran model, which I loosely refer to as the back-arc model because the majority of deformation is inferred to take place in a retro-arc setting, has been extensively applied to both modern and ancient continental margins with slight regional variations.

Within the existing framework, the geology of California played a defining role because the triad of Franciscan complex–Great Valley fore-arc–Sierra Nevada batholith—interpreted to represent a Mesozoic subduction complex–fore-arc basin–continental arc, respectively (Dickinson, 1981)—are well exposed in a region of good weather and, for the most part are readily accessible, so that they became core lithotectonic elements of the Cordilleran model (McPhee, 1993; Moores et al., 1999). Despite the long history of investigation and the thousands of boots on the ground, there is no consensus on many issues, and so major problems still abound. For example, the abrupt appearance of a strongly accretionary phase within the Franciscan complex at ~123 Ma is poorly explained in the current model because it provides no obvious source for the voluminous sediments nor why there should be such a change at that time. A 100 Ma deformational event within the Sierran and Peninsular Ranges Cordilleran batholithic terranes is poorly known and understood, as are the post-collisional plutonic suites. Also, there is still no consensus on how the contact of the Franciscan complex and Coast Range ophiolite—Tehama-Colusa serpentinite mélange—Great Valley Group developed, how the Franciscan high-grade rocks were exhumed, or even when it was initiated. Why does there appear to be no Laramide deformation in the Sierra Nevada and White-Inyo Mountains, yet deformation of that age occurs up and down the coast from northern South America to Alaska? Why was Sevier thrusting confined to the Great Basin region and why did magmatism stop there at 80 Ma, yet continue elsewhere? These, and countless other long-standing problems might be resolved with a different, and more dynamic, approach.

While there have been challenges to the standard Cordilleran model, they have been rare and not widely accepted. For example, Moores has long argued (1970, 1998; Moores and Day, 1984; Moores et al., 2002) that western North America was partially subducted to the west beneath an arc. Mattauer et al. (1983) recognized similarities between the deformation and metamorphism of the Alps and the Canadian Shuswap terrane, which led them to suggest a subduction model for the western edge of North America. The ideas of Chamberlain and Lambert (1985; Lambert and Chamberlain, 1988) were remarkably prescient because they developed a model for the Canadian Cordillera in which the majority of exotic terranes were assembled offshore then migrated northward to collide *en masse* with western North America. More recently, the standard Cordilleran model was challenged anew with a dynamic collisional model in which west-dipping subduction of North America led to Cretaceous–early Tertiary collision with an amalgamated, arc-bearing ribbon continent (Johnston, 2008; Hildebrand, 2009). In the collisional model, the suture between the two is located today within the Sevier–Rocky Mountain thrust belt just west of the North American shelf edge.

In this contribution I will demonstrate how the collisional model can provide a unifying rationale for wide-ranging, and seemingly unrelated, observations, including those in western California. Most of these relations were entirely unforeseen in the earlier presentations and demonstrate the predictive power of the collisional model. In order to accomplish this goal, I'll first present an overview of the pertinent geology, then utilize aspects of the collisional model to explain several previously unexplained features within the California sector, and end by offering a model that integrates key components of the orogen. The descriptions that follow are geared toward providing the reader with sufficient understanding to follow the basic arguments. It is not a comprehensive document to justify the collisional model, for such reviews, while perhaps slightly dated, are already available and contain the basic arguments (Johnston, 2008; Hildebrand, 2009). In my earlier contribution I focused more heavily on cratonic North America and its interactions with the Rubian superterrane. Here the focus is more on the amalgamation of the superterrane itself. Even readers who cannot accept the collisional model should still find value in the testable scenario for the assembly of the Cordillera presented here. Some repetition of basic ideas presented in Hildebrand (2009) is necessary to place the material within its proper context.

Rubia is a long, linear megaterrane, or ribbon continent, made up of nearly all the exotic terranes and superterranes of the North American Cordillera. It grew incrementally through time by the addition of various terranes to a nucleus of Neoproterozoic and Paleozoic terranes. Its greatest period of growth occurred during the Middle and Upper Jurassic when a variety of arcs, both oceanic and continental, were accreted to two different blocks, named Sierrita and Proto-Rubia, which in turn, collided at 160 Ma to form the Rubian ribbon continent.

PROBLEMS WITH THE EXISTING BACK-ARC MODEL

For those who have not yet read the earlier syntheses, it is worthwhile to review a few of the major weaknesses inherent in the currently accepted Cordilleran, or back-arc, model for North America. These are not all the problems extant in that model—and more are detailed in earlier contributions (Johnston, 2008; Hildebrand, 2009)—but are some major issues that, in my opinion, should be addressed in any model for it to be successful.

1. In current interpretations of the Cordilleran orogeny of North America, the passive margin, or miogeocline, is considered to extend from the craton westward to central Nevada and eastern California (Stewart, 1970, 1972, 1976; Stewart and Poole, 1974; Armin and Mayer, 1983). The main problem with such an interpretation is that the westerly-facing platform edge of the passive-margin terrace occurs today within the Sevier fold-thrust belt at or near the Wasatch front (Armstrong and Oriel, 1965; Peterson, 1977; Rose, 1977; Doelling, 1980; Palmer and Hintze, 1992); whereas there is another platform (Fig. 1), with its platform-margin edge located mainly in central Nevada and eastern California (Kepper, 1981; McCollum and McCollum, 1984; Heck and Speed, 1987; Montañez and Osleger, 1996; Morrow and Sandberg, 2008; Sheehan, 1986; Harris and Sheehan, 1998; Stevens et al., 1998; Stevens and Stone, 2007). The North American, or Rocky Mountain, margin developed in the Early Cambrian and consists of terrigenous clastics overlain by a Middle Cambrian carbonate platform, whereas the western platform, which Hildebrand (2009) termed the Antler platform, and is known farther north as the Cassiar platform, initially developed during the Neoproterozoic and consists of terrigenous clastics overlain by Lower Cambrian Archeocyathid-bearing reefs (Oriel and Armstrong, 1971; Stewart, 1972; Fritz, 1975; Read, 1980; Pope and Sears, 1997). Both Johnston (2008) and Hildebrand (2009) recognized that the two platforms were separated by deeper-water sedimentary facies and had different tectonic, sedimentological, and magmatic histories; so suggested that they sat on different plates, which were united during the Cretaceous–Tertiary Cordilleran orogeny.

2. Another feature difficult to explain in the back-arc model is the paucity of latest Neoproterozoic–Early Cambrian rift basins and associated volcanic rift deposits on the North American margin. Except for local areas, most modern rifts and rifted margins (Fig. 2) are characterized by abundant volcanic deposits (Ebinger, 1989; Ebinger and Casey, 2001; Menzies et al., 2002; Sawyer et al., 2007). It is, of course, possible that parts of the rifted margin were hyperextended, nonvolcanic, and highly asymmetrical (Lister et al., 1991) such that rifted crust was predominantly on one margin like the present-day North Atlantic margin (Keen and Dehler, 1997); but since that reduces the width of the rifted margin on the other side, it would be likely that palinspastically-restored units west of the Sevier fold-thrust belt would not have been floored by North American crust. And based on today's margins it seems highly unlikely that the entire margin from Alaska to Mexico would be amagmatic. It is the overall scarcity of latest Precambrian–Cambrian volcanic rift deposits erupted on extended Laurentian crust over the entire length of the orogen that is entirely unaccounted for in the currently-accepted model.

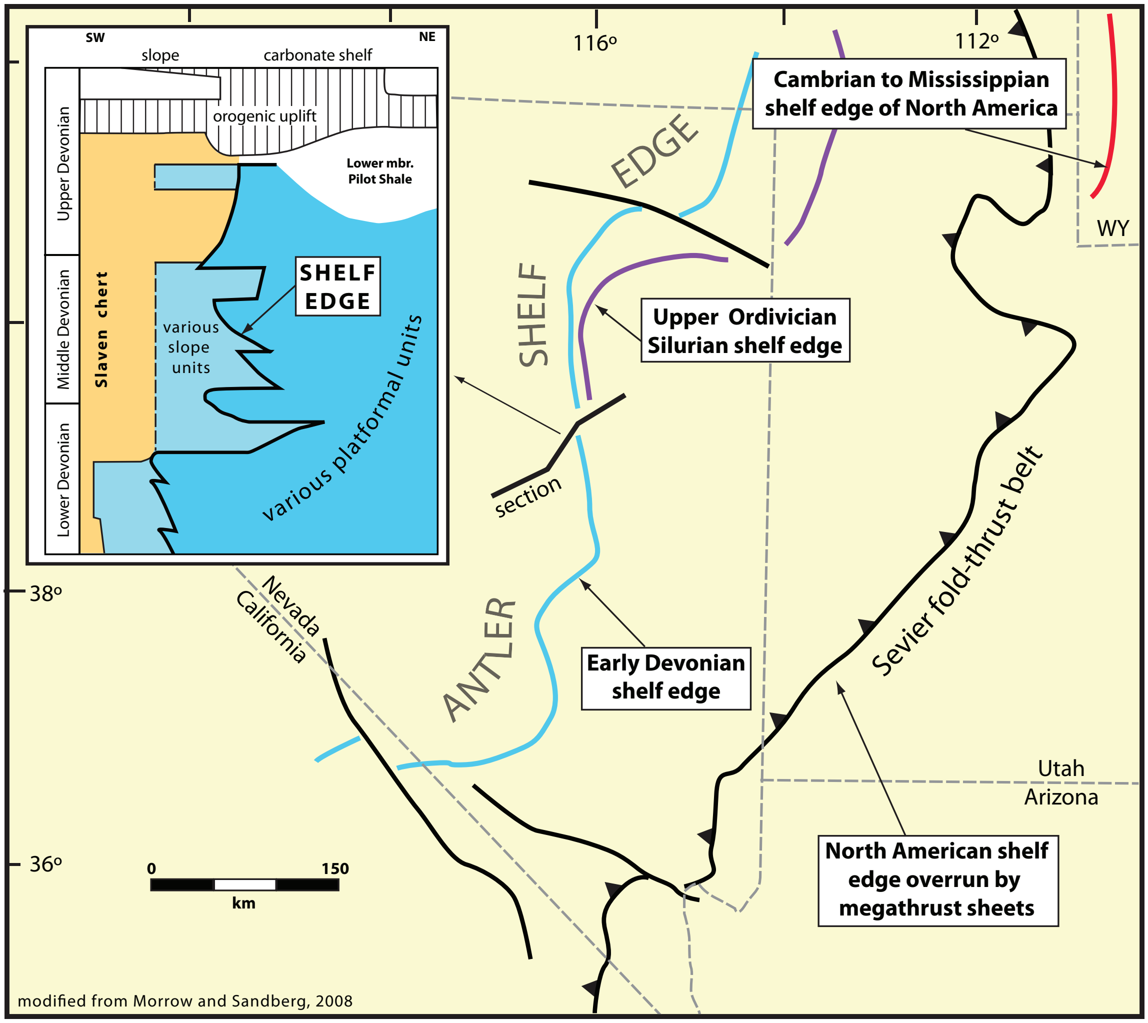


Figure 1. A sketch map showing the locations of the two Paleozoic shelf edges located within the Great Basin segment of the orogen. The Antler shelf edge is located some 200–400 km west of the North American margin. Exotic rocks riding on eastward-vergent thrust faults overrode the North American shelf edge nearly everywhere, except in the Wyoming salient, located in SW Wyoming, and in the Main Ranges of the southern Canadian Cordillera, where the shelf edge is preserved and known as the Kicking Horse Rim (not shown). Modified from Morrow and Sandberg (2008).

Some workers (Stewart, 1972; Burchfiel and Davis, 1975; Lund, 2008) argued that rocks of the Windermere Supergroup, and equivalents, or even older rocks (Dehler et al., 2010), represent rift deposits on the western margin of North America, but as they are some 85–100 Myr older (Lund et al., 2003; Fanning and Link, 2004) than the development of the passive margin, the margin wouldn't have retained enough heat to match the rate of early Paleozoic subsidence (Bond and Kominz, 1984; Devlin and Bond, 1988). Additionally, the Windermere doesn't contain the requisite tracts of volcanic rocks.

3. The presence of persistent mafic magmatism throughout much of the Paleozoic in rocks commonly considered to represent drift facies of the passive margin (Fig. 3), such as the Selwyn Basin and Kechika Trough (Goodfellow et al., 1995; Cecile, 2010) is difficult to reconcile with a passive margin setting, as is the recent recognition of a suite of 664–486 Ma alkaline plutons (Fig. 4) intruding rocks of the Belt Supergroup and its miogeoclinal Paleozoic cover in central Idaho (Lund et al., 2010; Gillerman et al., 2008). Furthermore, at least one of the plutons was likely deroofed during the Upper Cambrian (Link and Thomas, 2009; Link and Janecke, 2009), a peculiar occurrence for the outer part of a miogeocline.

4. In general, there is no evidence for collision, in the form of deformation or exotically-derived sedimentation, on the North American shelf from the Cambrian to the Cretaceous. For example, extensive areas of western Nevada were supposedly incorporated as part of North America after the Mississippian, but the North American shelf saw no deformation or sedimentation related to major 160 Ma deformation, including 7–14 km of crustal thickening and major thrusting in the Black Rock Desert

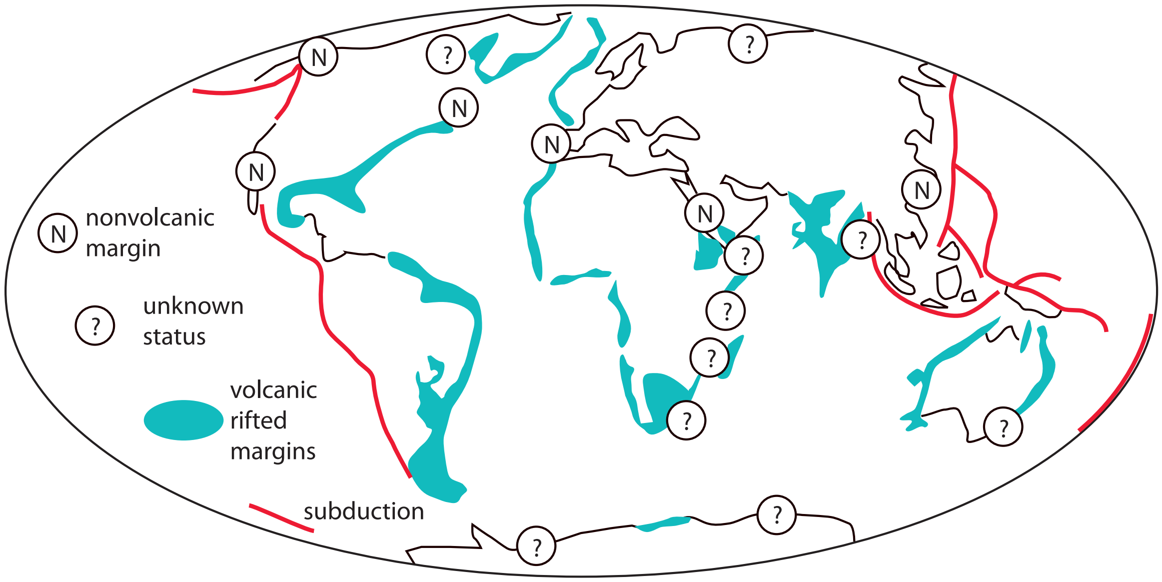


Figure 2. Distribution of passive margin types on today's continental margins, modified from Menzies et al. (2002). Note that nonvolcanic rifted margins are of extremely limited distribution and area compared to volcanic margins. This indicates that the absence of rift-facies volcanic rocks in the North American Cordillera is peculiar and requires an explanation. Hildebrand and Bowring (1999) suggested that most of the rift-facies rocks were subducted due to slab break-off.

and other nearby terranes (Wyld et al., 2001; Wyld, 2002). There was also intense deformation and metamorphism throughout the hinterland belt during the Jurassic in which the crust was doubled in thickness with the development of westerly-vergent recumbent thrust nappes (for example: Camilleri et al., 1997), yet there were no deformational effects on the shelf, located less than 80–100 km away after restoration of Basin and Range normal faulting (Hildebrand, 2009).

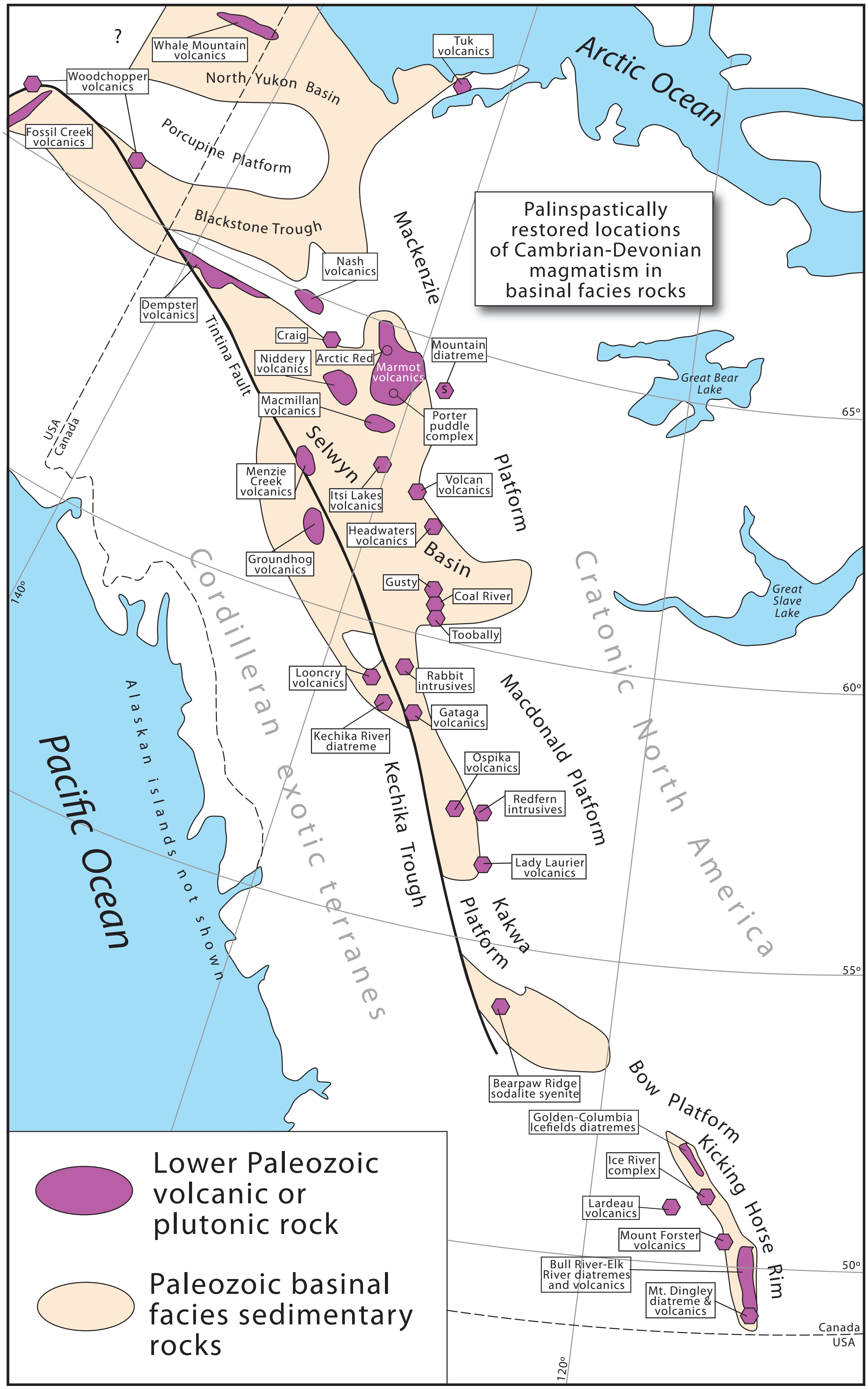
Similar relations exist in the Canadian Rockies. In the Selkirk fan structure, located on the eastern flank of the Monashee complex (Fig. 5, on insert accompanying this volume[1]), which is an erosional window that exposes probably duplexed North American basement and cover rocks beneath the Kootenay terrane, 187–173 Ma plutons were intruded before and/or during deformation and before a 173–168 Ma period of rapid exhumation of rocks from 7 kb to 3 kb (Colpron et al., 1996). The Scrip nappe, a west-verging isoclinal structure located just to the north, has an overturned limb as broad as 50–60 km across strike and probably formed at about the same time (Raeside and Simony, 1983). Similar structures farther south were documented and reported by Höy (1977) and, while poorly constrained, are likely to be between 178 and 164 Ma (Read and Wheeler, 1975). Colpron et al. (1996, 1998) argued that these events took place in strata of the outer and proximal North American miogeocline (Colpron and Price, 1995), but there is simply no record, either deformational or sedimentological, of major plutonism, folding, thickening, and exhumation at this time on the North American cratonic terrace (Fig. 6). In fact, sedimentary rocks of this age deposited upon the North American platform are phosphorites, with up to 30% P_20_5, and which typically form along the eastern margins of open oceans due to upwelling of cold, nutrient-rich waters (Poulton and Aitken, 1989; Parrish and Curtis, 1982).

Jurassic sedimentary rocks of the Morrison (United States) and Fernie formations and Kootenay group (Canada) are the only Jurassic units of the passive margin sequence known to contain westerly-derived sediment, but they contain no plutono-metamorphic debris and are at least 25–40 Myr older than the initiation of foredeep sedimentation in the Western Interior basin. Additionally, the Morrison doesn't thicken westward, but instead thins westward from depocenters located some distance from the platform edge along the Utah-Colorado border (Heller et al., 1986; DeCelles, 2004).

5. Recent fieldwork and U-Pb geochronology have shown that many of the Canadian exotic terranes, including the Slide Mountain oceanic tract, were amalgamated by the Triassic, not the Jurassic as required in the back-arc model (Beranek and Mortensen, 2007; Beranek et al., 2010a). Furthermore, robust paleomagnetic data indicate that this block and other terranes of the Cordillera located west of the cratonic terrace did not completely dock with the craton until ~70–60 Ma (Enkin, 2006; Enkin et al., 2006a, Kent and Irving, 2010).

6. Rocks of the Belt-Purcell supergroups are generally interpreted to have been deposited on Laurentian crust, but 1.2–1.0 Ga metamorphism and deformation found in Belt-Purcell metasedimentary rocks and intrusions (Anderson and Davis, 1995; Nesheim et al., 2009; Zirakparvar et al., 2010) are

[1]Figure 5 is also available as GSA Data Repository Item 2013125, online at www.geosociety.org/pubs/ft2013.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.



Palinspastically restored locations of Cambrian-Devonian magmatism in basinal facies rocks
Arctic Ocean
Pacific Ocean
Whale Mountain volcanics
Tuk volcanics
Woodchopper volcanics
Fossil Creek volcanics
North Yukon Basin
Porcupine Platform
Blackstone Trough
Mackenzie
Nash volcanics
Dempster volcanics
Tintina Fault
Craig
Arctic Red
Niddery volcanics
Marmot volcanics
Mountain diatreme
Great Bear Lake
Macmillan volcanics
Porter puddle complex
Selwyn
Platform
USA
Canada
Menzie Creek volcanics
Itsi Lakes volcanics
Volcan volcanics
Headwaters volcanics
Groundhog volcanics
Basin
Cordilleran exotic terranes
Cratonic North America
Gusty
Coal River
Toobally
Great Slave Lake
140°
Looncry volcanics
Rabbit intrusives
Alaskan islands not shown
Kechika River diatreme
Gataga volcanics
Macdonald Platform
60°
65°
Kechika Trough
Ospika volcanics
Redfern intrusives
Lady Laurier volcanics
Kakwa Platform
55°
Bearpaw Ridge sodalite syenite
Bow Platform
Golden-Columbia Icefields diatremes
Kicking Horse Rim
Ice River complex
Lardeau volcanics
Mount Forster volcanics
Bull River-Elk River diatremes and volcanics
Mt. Dingley diatreme & volcanics
50°
Canada
USA
120°
Lower Paleozoic volcanic or plutonic rock
Paleozoic basinal facies sedimentary rocks

Figure 3. Palinspastically restored locations of Cambrian to Devonian magmatism in basinal facies rocks of the Canadian Cordillera modified after Goodfellow et al. (1995). In the traditional model these rocks are generally considered to be part of the miogeocline, but such basinal facies magmatism is more typical of marginal basins than drift phase passive margins. The suture between Rubia and North America is interpreted to lie between the basinal facies rocks and the North American Paleozoic carbonate terrace (Johnston, 2008; Hildebrand, 2009), which, on this figure, is most easily located in the south by the swarm of alkaline plutons and diatremes located just to the west of the Kicking Horse Rim.

unknown in cratonic northwestern North America and so suggest long-distance transport.

7. The back-arc model inadequately explains the intense shortening, high-grade metamorphism (>9 kb and ~800 °C), convergent temperature-time paths, and extensional collapse of the Sevier hinterland, where a minimum of 70 km of shortening and as many as 30 km of crustal thickening occurred during the Late Cretaceous (Camilleri et al., 1997). This means that the crust was doubled to twice normal cratonic thickness. It is unclear how the compressive stresses required could be generated in a back-arc environment and how the thinned continental crust of a back-arc basin could produce the observed pressures during deformation, given that the deformation was generally thin-skinned. Experimental and seismic data show that similar zones in other orogens are readily interpreted as well-developed erosional thrust duplexes made of lower-plate rocks with a klippe of exotic rocks to the foreland side (Malavieille, 2010; Schmid et al., 2004).

8. In the most recent variant of the back-arc model (DeCelles et al., 2009), ~400 km of North American cratonic crust must have been subducted to the west beneath the Sierra Nevada to balance the upper-crustal shortening within the cover. How 400 km of cratonic crust might be subducted to sufficient depth to be melted, without being attached to oceanic lithosphere to pull it down into the mantle, is problematic.

9. There are strong mismatches in the timing of deformation between that known from the North America platform and that known from rocks just to the west (Fig. 7). For example, within the Canadian Rockies, rocks of the platform were not deformed until the Santonian–Campanian, but rocks immediately to the

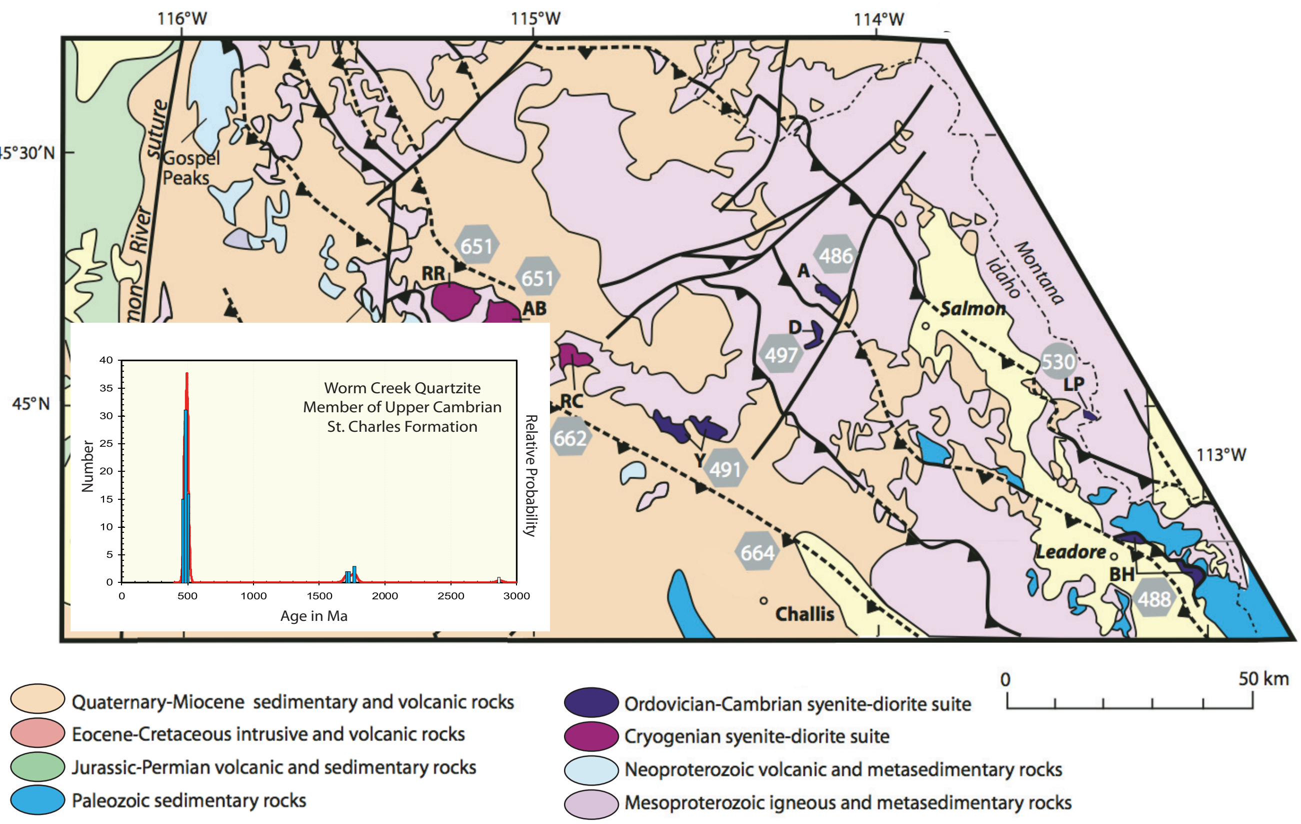


Figure 4. Distribution of recently discovered Cryogenian to Ordovician plutonic rocks in the Belt-Windermere allochthons after Lund et al. (2010). Also shown are the results of U-Pb analyses of detrital zircons in the Worm Creek member of the Upper Cambrian St. Charles Formation of south-central Idaho, which suggest, based on the dominance of 498 Ma detritus, that the 497 Ma Deep Creek pluton (D) was deroofed soon after emplacement (Link and Thomas, 2009; Link and Janecke, 2009). Other abbreviations refer to individual plutons not discussed here.

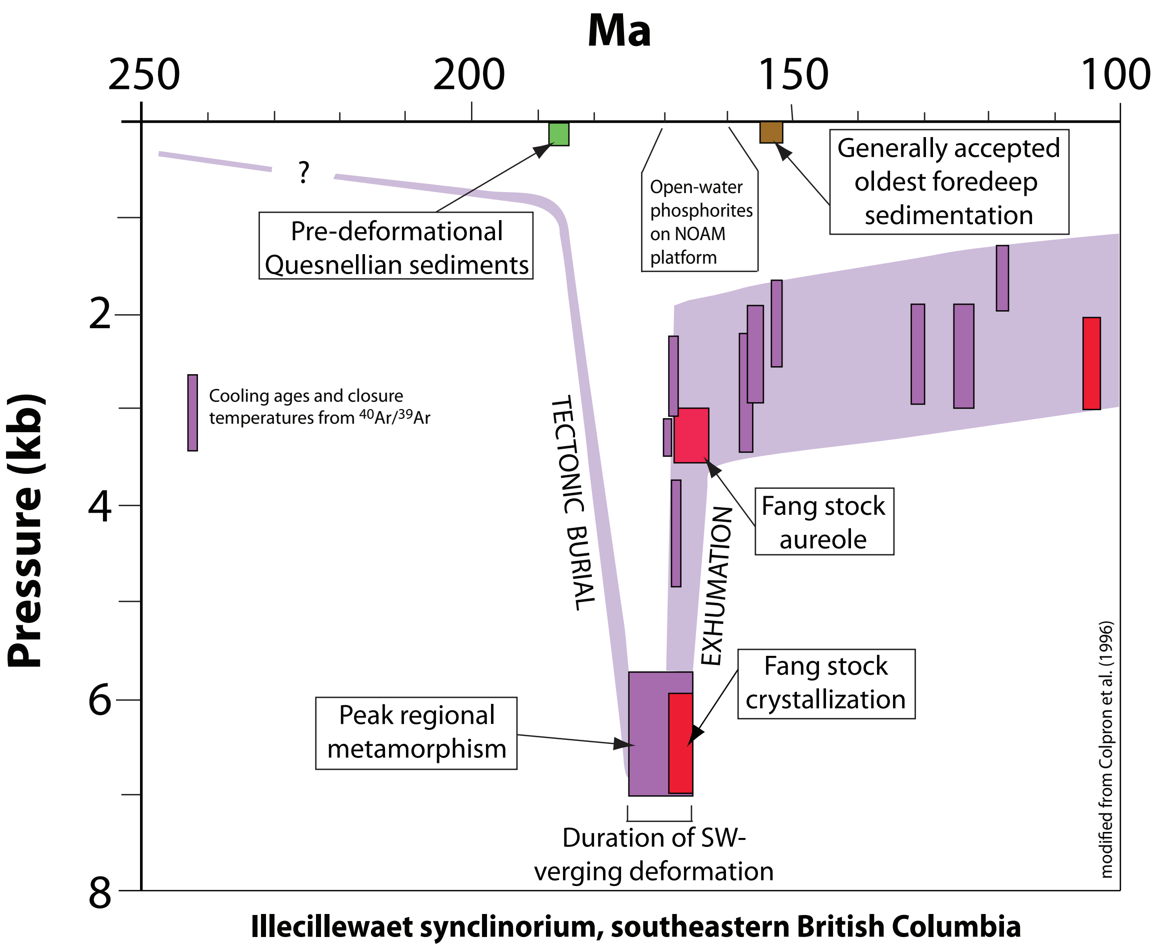


Figure 6. Pressure-time plot for the Illecillewaet synclinorium, which occurs along the eastern side of the Monashee complex in the Canadian Rockies (Fig. 5), modified from Colpron et al. (1996) showing the rapid exhumation of rocks within the synclinorium between 170 and 160 Ma and the simultaneous deposition of open-water, North American (NOAM) phosphorite deposits, which suggest that the two regions were not proximate at that time. In this paper I argue that the Illecillewaet synclinorium was formed during widespread collisional events within the Rubian superterrane prior to its collision with North America.

west on the Windermere high were involved in major thrusting earlier than 108 Ma (Larson et al., 2006). Similarly, rocks of the hinterland to the Sevier fold-thrust belt contain evidence of two intense periods of deformation, Jurassic and Upper Cretaceous (Camilleri et al., 1997), while rocks of the North American platform terrace show no evidence of Jurassic deformation but had major Aptian–Cenomanian eastward-directed thin-skinned thrusting (DeCelles, 2004).

The flaws in the current back-arc model demonstrate a need to re-examine the concept of Cordilleran orogenesis. Today we understand so much more about how plate tectonics creates the geology that we see on the surface than we did 40 years ago when the Cordilleran model was developed, so it is not surprising that a fresh look is warranted. What follows is not a definitive document—for no doubt parts of it are incorrect and will require revision, or even abandonment, as new facts are revealed and concepts developed—but rather a transient attempt to integrate current knowledge into an actualistic model that can be tested and refined as new data are collected and analyzed.

A SEGMENTED OROGEN

A principal feature of the Cordilleran orogen of North America is its segmented nature (King, 1966). Based on both contrasting geology and varied development, Hildebrand (2009) divided part of the orogen into three segments: Canadian, Great Basin, and Sonoran (Fig. 5) from north to south. The two northern elements are approximately divided by the Orofino fault, whereas the Great Basin and Sonoran sectors are separated by the Phoenix fault, an apparent sinistral transform fault. Detrital zircon profiles from the basal foredeep (Leier and Gehrels, 2011) clearly show the break in the vicinity of the Orofino fault and further serve to delineate it as a significant segment boundary. A major dextral fault, along which the Sierra Nevada and

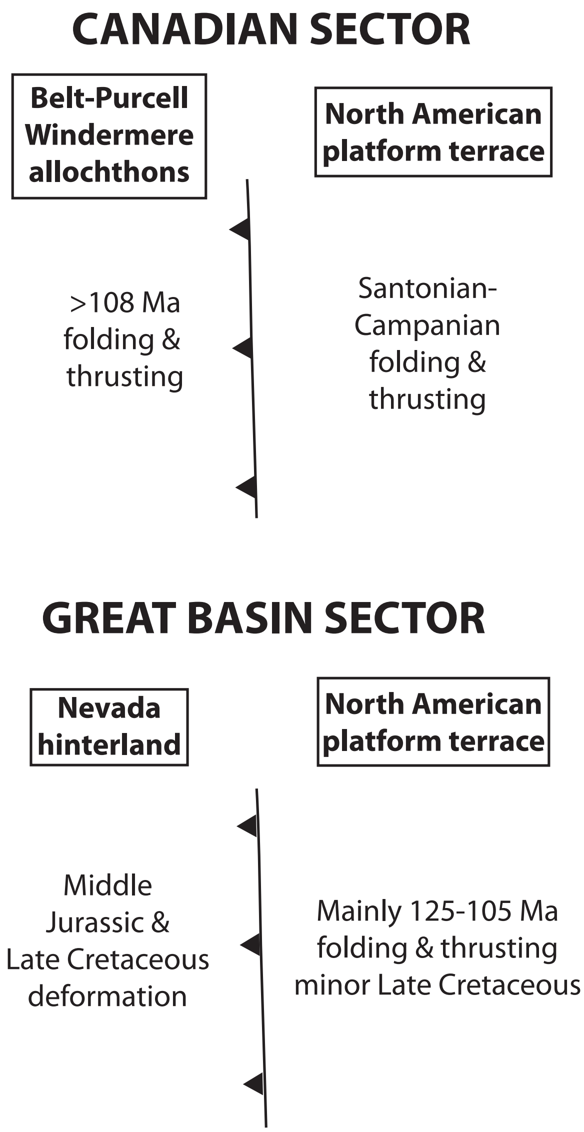


Figure 7. Diagram illustrating the mismatched age differences in deformation of two immediately adjacent areas along the North American margin. In the Canadian sector, rocks of the Windermere and Belt-Purcell supergroups were deformed and thrust prior to 108 Ma, whereas rocks of the North American platform terrace, located just a few kilometers to the east, were not deformed and thrust until the Santonian–Campanian. In the Great Basin sector, rocks of the North American platform terrace were deformed and thrust to the east mainly between 124 and 105 Ma, but rocks in the hinterland belt located just to the west exhibit only Jurassic and latest Cretaceous deformation.

the Atlanta lobe of the Idaho batholith might be restored, was hypothesized to lie buried beneath much younger lavas of the Snake River plain (Hildebrand, 2009). An additional segment recognized here is the Alaskan segment, which lies mainly west of a conspicuous break, now mostly covered by younger deposits, between the Canadian terranes, such as Yukon-Tanana and Selwyn basin, and the bulk of the northern and western terranes of Alaska (Fig. 5).

Understanding the geological and tectonic development within each sector is critical to unraveling the development of the orogen because, not only are there large differences in magmatic and structural evolution between segments (Armstrong, 1974), but a considerable body of evidence suggests that there was so much latitudinal migration of terranes along the orogen that a terrane might have been located in one sector for a period of time and then subsequently transferred to another, and also because there might be profound changes in deformational style or magmatism along strike in adjacent segments (for example, Oldow et al., 1989). The Coast plutonic complex provides a good example, for it was located within the Sonoran segment at 80 Ma, but by 58 Ma was entirely within the Canadian sector. Similarly, the eastern part of the Canadian segment is dominated by thin-skinned thrusting of the Rocky Mountain fold-thrust belt (Price, 1981), which was coeval with thick-skinned Laramide deformation in the Great Basin segment. Thus, reconstructions and models that don't consider these aspects are unlikely to succeed.

Here I start with a short review of the shared North American, or Laurentian, passive margin, located within the Canadian and Great Basin sectors, then describe the geology of the individual sectors, before presenting a plausible and testable model for their origin and final assembly.

LAURENTIAN PASSIVE MARGIN

The oldest passive margin rocks deposited on the thermally subsiding North American craton (Bond and Kominz, 1984) were Early Cambrian sequences of quartzose siliciclastic rocks overlain by shelf-to-slope carbonate rocks that pass abruptly westward into sparse and thin, shaly basinal-facies rocks (Rigo, 1968; Stewart, 1970). The Paleozoic carbonate shelf-to-basin transitions are observable today in the Wyoming salient (Fig. 8) of the eastern Sevier fold-thrust belt near the Utah-Wyoming border (Peterson, 1977; Rose, 1977; Doelling, 1980; Palmer and Hintze, 1992) and in the Main Ranges of the Canadian Rockies (Cook, 1970; Aitken, 1971). The basal Cambrian sandstone also fines westwardly into shale at the shelf edge (Oriel and Armstrong, 1971; Middleton, 2001). Overall, the platform-to-rise transition persisted in more or less the same position, except for occasional eastward transgressions, and Pennsylvanian uplift and sags in east-central Utah, from the Cambrian at least through the Jurassic (Hansen, 1976; Koch, 1976; Rose, 1977; Blakey, 2008). The shelf-slope transition was termed by some the Wasatch hinge line (Hintze, 1988; Poole et al., 1992), which is not strictly correct, as the hinge line is the most landward point of lithospheric stretching, and the shelf edge probably marks the most landward point of upper-crustal extension (brittle faulting). Lower-crustal stretching likely continued farther west.

During the Cambrian, North America was rimmed by carbonate platforms, and there is nowhere evidence for any topographically high-standing terrain that might have shed significant

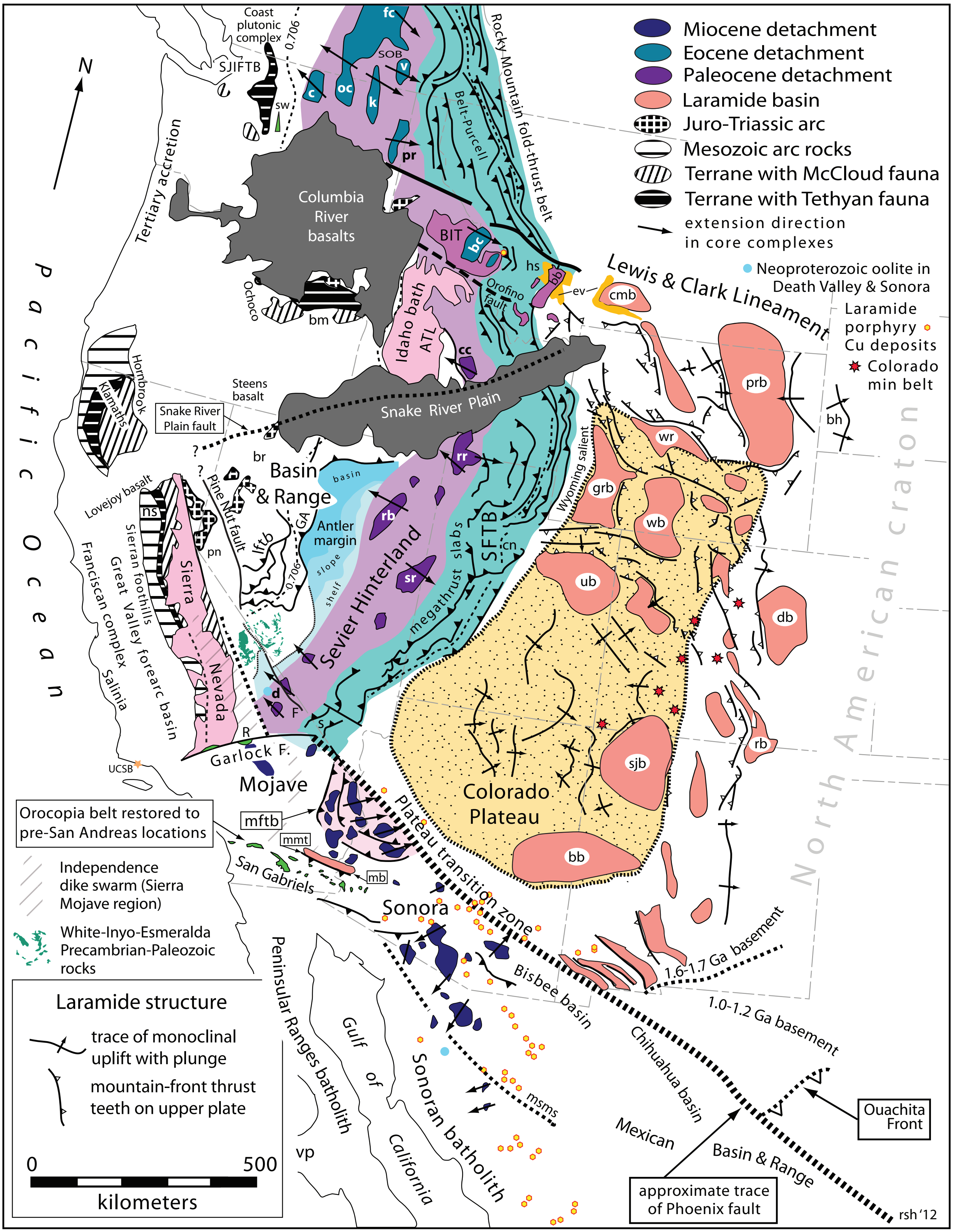

Miocene detachment
Eocene detachment
Paleocene detachment
Laramide basin
Juro-Triassic arc
Mesozoic arc rocks
Terrane with McCloud fauna
Terrane with Tethyan fauna
extension direction in core complexes
Neoproterozoic oolite in Death Valley & Sonora
Laramide porphyry Cu deposits
Colorado min belt
Coast plutonic complex
SJIFTB
Columbia River basalts
Rocky Mountain fold-thrust belt
Belt-Purcell
Lewis & Clark Lineament
Snake River Plain
Snake River Plain fault
Steens basalt
Basin & Range
Antler margin
Sevier Hinterland
megathrust slabs
SFTB
Wyoming salient
Colorado Plateau
Idaho bath
Orofino fault
Tertiary accretion
Pacific Ocean
Hornbrook
Klamaths
Lovejoy basalt
Pine Nut fault
Sierran foothills
Great Valley forearc basin
Franciscan complex
Salinia
Sierra Nevada
Garlock F.
Mojave
Orocopia belt restored to pre-San Andreas locations
Independence dike swarm (Sierra Mojave region)
White-Inyo-Esmeralda Precambrian-Paleozoic rocks
Laramide structure
trace of monoclinal uplift with plunge
mountain-front thrust teeth on upper plate
0
500
kilometers
San Gabriels
Sonora
Plateau transition zone
Peninsular Ranges batholith
Gulf of California
Sonoran batholith
Bisbee basin
Chihuahua basin
Mexican Basin & Range
approximate trace of Phoenix fault
1.6-1.7 Ga basement
1.0-1.2 Ga basement
Ouachita Front
North American craton
rsh '12

Figure 8. Geological sketch map of northwest Mexico, western United States, and southwesternmost Canada, illustrating the various tectonic elements discussed in the text. Approximate eastern limit of exotic allochthons marked by dashed line in thrust belt. Note that this is also the suture. Abbreviations: ATL—Atlanta lobe; bb—Baca basin; bb—Boulder batholith; bc—Bitterroot complex; bh—Black Hills; BIT—Bitterroot lobe; bm—Blue Mountains terranes; br—Black Rock Desert; c—Cascades core; cc—Clearwater complex; cmb—Crazy Mountains Basin; cn—Charleston-Nebo salient; d—Death Valley; db—Denver Basin; ev—Elkhorn volcanics; F—Furnace Creek fault; fc—Frenchman Cap; GA—Golconda allochthon; grb—Green River Basin; hs—Helena salient; k—Kettle complex or dome; lftb—Luning-Fencemaker thrust belt; mb—McCoy Mountains formation; mftb—Big Maria fold-thrust belt; mmt—Mule Mountains thrust system; msms—Mojave-Sonora megashear; ns—northern Sierra Nevada; oc—Okanagan complex or dome; pn—Pine Nut block; pr—Priest River complex; prb—Powder River basin; R—Rand schist; rb—Raton basin; rr—Albion–Raft River–Grouse Creek Ranges; s—Spring Mountains; SFTB—Sevier fold-thrust belt; sjb—San Juan Basin; SJIFTB—San Juan Islands fold-thrust belt; SOB—Shuswap–southern Omineca belt; sw—Swakane gneiss; ub—Uinta Basin; UCSB—University of California at Santa Barbara; v—Valhalla complex or dome; vp—Vizcaino Peninsula; wb—Washakie Basin; wr—Wind River Basin. Orocopia belt is restored to pre–San Andreas fault configuration after Nourse (2002), but the modern shoreline is left as is for reference. Note also that the extension within the younger Basin and Range Province is not restored. In this and subsequent figures, I use the geomorphological Colorado Plateau. Extension directions in Cordilleran core complexes after Wust (1986).

amounts of clastic detritus outboard of the carbonate platform. Thus, the Cambrian basinal facies of North America were probably quite starved. The overall scheme for western North America would have been an abrupt carbonate rim with a narrow slope-facies debris fan and a thin, sediment-poor, contourite-dominated rise facies.

Located to the west of the off-platform rocks is another west-facing platform, which I (Hildebrand, 2009) termed the Antler margin (Figs. 1 and 8). Rocks of this platform were overthrust by Paleozoic rocks of the Roberts Mountain and Golconda allochthons. This margin is older than the North American margin as it formed during the Neoproterozoic and contains an Early Cambrian carbonate bank characterized by abundant Archeocyathids and distinctive oolitic and intraclast beds (Fritz, 1975; Pope and Sears, 1997). The west-facing Paleozoic Antler platform edge is well documented in Nevada and California (Fig. 1), and there is no doubt that it is a different platform than the Sevier platform (Kepper, 1981; McCollum and McCollum, 1984; Heck and Speed, 1987; Montañez and Osleger, 1996; Morrow and Sandberg, 2008; Sheehan, 1986; Harris and Sheehan, 1998; Stevens et al., 1998; Stevens and Stone, 2007). This does not mean that the margin was not originally located in the western Americas, but merely that if so, it was separated from the margin and is exotic in its current location.

In the Canadian Cordillera, the story is similar to that in the United States. A westwardly thickening wedge of mature and shallow marine clastic rocks sits unconformably on cratonic basement (Fig. 9). The lower clastic rocks are overlain by two margin-parallel facies: an inner carbonate bank and an outer slope to basinal facies, which are separated by an algal reef complex known as the Kicking Horse Rim (Aitken, 1971). The shelf-slope facies transition (Fig. 10) occurs today in the Main Ranges close to the Alberta–British Columbia border, where the Middle Cambrian carbonate platform terminates into the shaly slope-to-basinal facies of which the best known accumulation is the Burgess Shale (Cook, 1970; Price and Mountjoy, 19707; McIlreath, 1977). The facies change is the locus of faulting and huge gravity collapse scarps, and marks changes in penetrative strain and styles of folding (Dahlstrom, 1977; Stewart et al., 1993).

The equivalent of the Antler shelf in Canada is the Cassiar platform, which lies to the west of the off-shelf basinal sedimentary rocks (Johnston, 2008). Like their counterparts in the Great Basin region, rocks of the Cassiar platform constitute a shallow-water, mixed carbonate-siliciclastic platform ranging in age from Neoproterozoic to Mesozoic. A number of lines of evidence, including faunal provinciality, basement ages, and contrasting Mesozoic structural evolution, led Johnston (2008) to suggest that rocks of the Cassiar platform were part of a much larger ribbon continent that was exotic with respect to North America prior to the Cretaceous.

GREAT BASIN SECTOR

Sevier Fold-Thrust Belt

Forming an important eastern element within the Cordilleran orogen is the Sevier fold-thrust belt, which is a thin-skinned deformational belt that extends from the Spring Mountains of Nevada in the south to the Orofino fault in the north (Figs. 5 and 8). Within the belt, rocks of the North American passive margin were detached from their basement and transported eastward on a basal décollement, typically located within the shaly Cambrian part of the marginal section (Armstrong and Oriel, 1965; Armstrong, 1968; Allmendinger, 1992; Burchfiel et al., 1974a, 1974b, 1992, 1998; DeCelles and Coogan, 2006). While most workers consider the thrust belt to be a distal, back-arc manifestation of eastwardly-directed subduction of oceanic lithosphere beneath North America, Hildebrand (2009) proposed that the thrust belt represents a typical collisional thrust belt formed as the western margin of North America was pulled beneath the previously amalgamated Rubian superterrane (Fig. 11). Structurally above thrust slices of the North American platform terrace are huge thrust sheets, up to 28 km thick by hundreds wide and long (Sears, 1988; DeCelles and Coogan, 2006; Fermor and Moffat, 1992). These include the megathrust sheets, containing sections

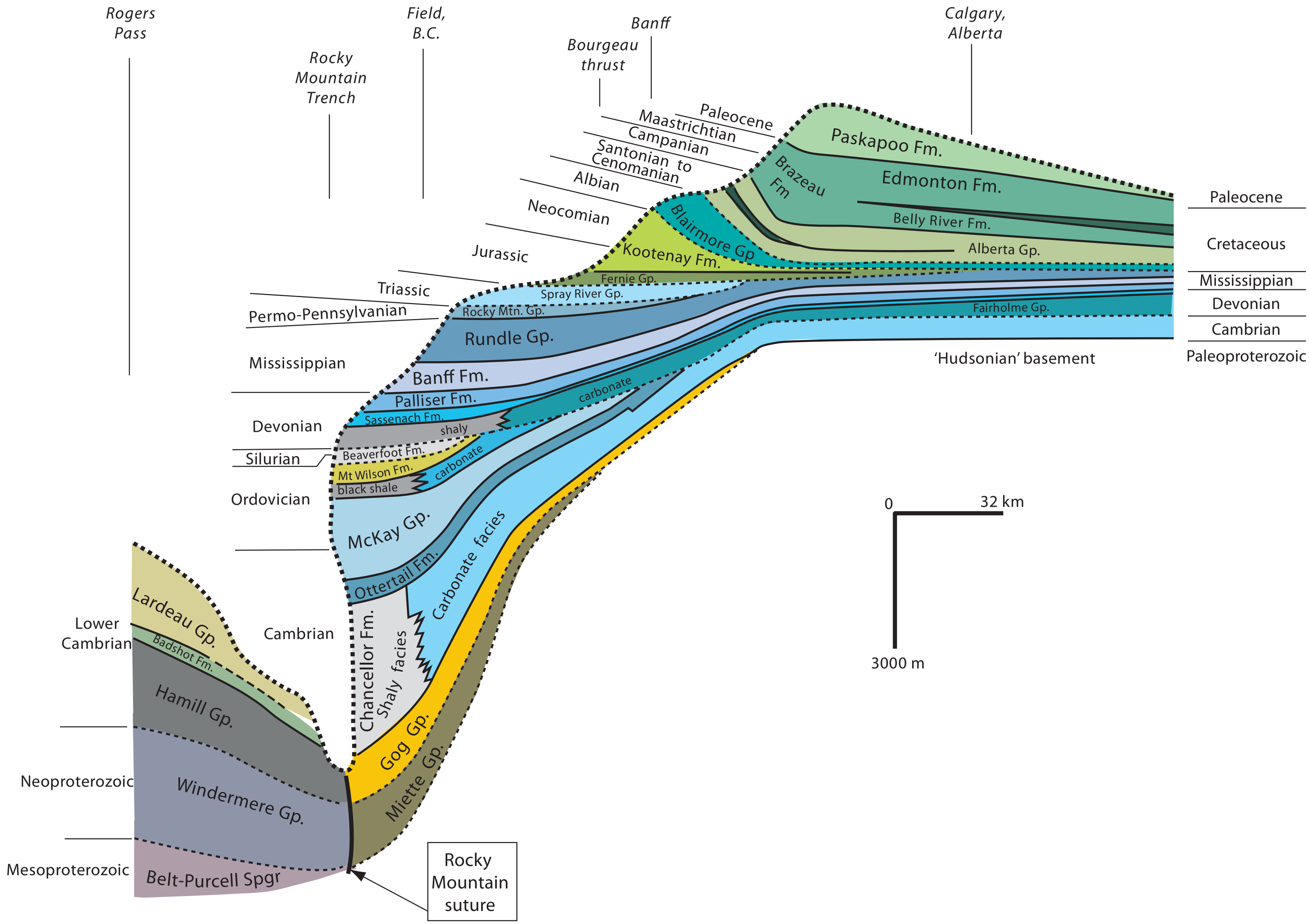


Figure 9. Schematic relations of stratigraphic units on the North American margin of Canada, and those located just to the west and commonly correlated with it; however, in this contribution the western rocks are considered to be exotic. Figure modified from Price and Monger (2003).

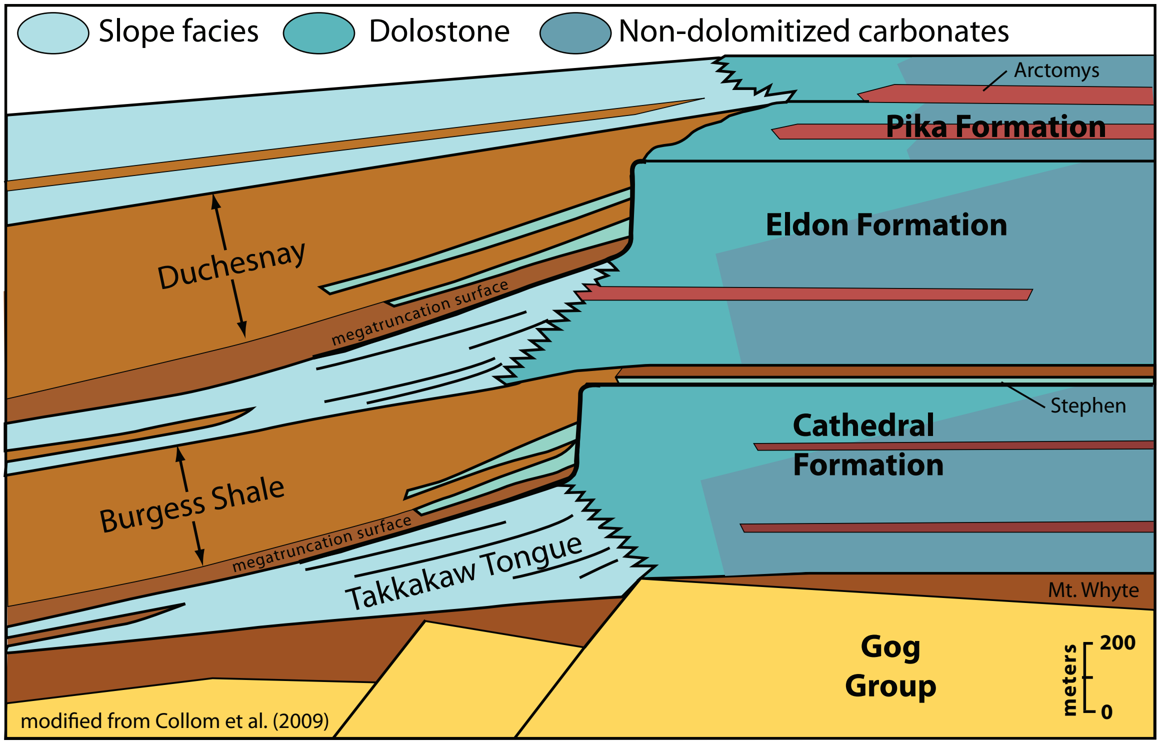


Figure 10. Illustrative cross section, modified from Collom et al. (2009), showing the relationships between platformal, slope, and basinal facies rocks at the Kicking Horse Rim of the Main Ranges, southern Canadian Rockies. Both the Cathedral and Eldon Formation had ~200 m of relief on their western escarpments.

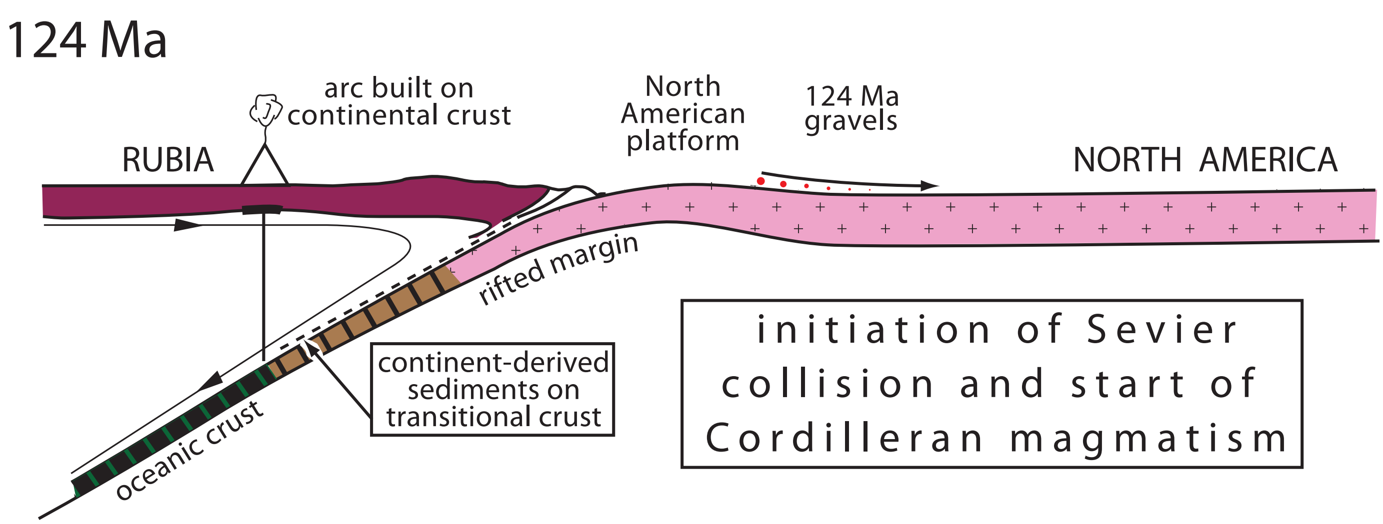


Figure 11. Plate model sketch illustrating the model of Hildebrand (2009) for the Sevier orogeny in the Great Basin sector of the orogen. A similar scenario, with westward subduction of North America beneath the Rubian ribbon continent, is also envisioned here for the Laramide event in both the Canadian and Sonoran sectors of the orogeny.

more than 7–10 km thick of Neoproterozoic–Cambrian sedimentary rocks within the Great Basin area (Christie-Blick, 1982, 1997; DeCelles and Coogan, 2006). To the north and located in both the Canadian and Great Basin sectors are huge allochthons containing rocks of the Belt-Purcell-Windermere supergroups carried on the Lewis-Eldorado-Hoadley thrust complex (Cook and van der Velden, 1995; Mudge and Earhart, 1980; Sears, 2001). Both the megathrust sheets and the Belt-Purcell-Windermere allochthons have no stratigraphic equivalents on the North American craton; so Hildebrand (2009) considered them to be exotic.

The sedimentary record of collision started at different times depending on the location, but nearly everywhere was predated by the deposition of gravels and conglomerates—comprising a wide variety of sedimentary clasts such as chert, quartzite, limestone, and siltstone—eroded from older rocks of the North American platform and then dispersed eastward to form a thin veneer over a regional unconformity and a calcrete-silcrete paleosol complex developed on the Morrison Formation and its lateral equivalents (Schultheis and Mountjoy, 1978; Leckie and Smith, 1992; Heller and Paola, 1989; Yingling and Heller, 1992; Currie, 2002; Ross et al., 2005; Zaleha and Wiesemann, 2005; Zaleha, 2006; Roca and Nadon, 2007; Greenhalgh and Britt, 2007). These gravels are known by various local names, such as the Cadomin, Kootenai, Lakota, Cloverly, Ephraim, Buckhorn, Pryor, etc., and they are extensive (Heller et al., 2003), occurring up and down the margin (Fig. 12). In the Great Basin sector the gravels are overlain by marine mudstones and siltstones of Albian and Cenomanian age, which mark the first sedimentary rocks of the Sevier foredeep, known regionally as the Western Interior basin (Kauffman, 1977; Hunt et al., 2011). Newly collected and analyzed detrital zircons from the Cadomin Formation in the Canadian segment indicate that it is considerably younger, perhaps by 30 Myr or more, than its generally considered equivalents in the Great Basin sector

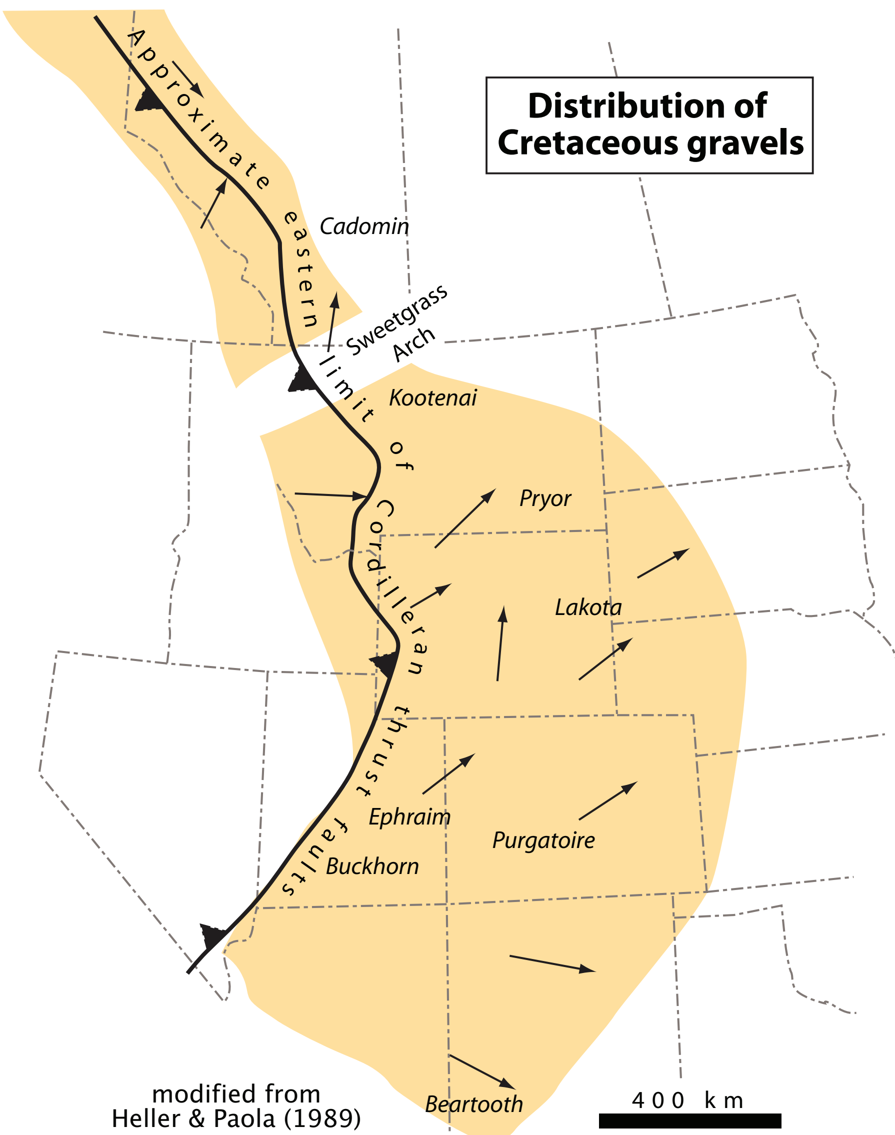


Figure 12. Sketch map showing the distribution of Cretaceous gravels at the base of the Cordilleran foredeep. The gravels contain clasts of sedimentary rocks that are generally interpreted to have been derived from sedimentary units of the North American platform. The Sweetgrass arch appears to have initially separated two different parts of the basin, developed at different times. Hildebrand (2009) suggested that the gravels were derived from the erosion of the North American platform as it rode up and over the outer swell to the westward-dipping trench.

(Leier and Gehrels, 2011). The two regions are separated by the Sweetgrass Arch, a positive element of the North American craton during the Cretaceous (Lorentz, 1982; Podruski, 1988).

Hildebrand (2009) suggested that the unconformity and the overlying conglomerates marked the passage of the North American craton over the outer forebulge of the trench (Currie, 1998). As the continent passed over the bulge (McAdoo et al., 1978; Forsyth, 1980; Jacobi, 1981; Stockmal et al., 1986; Yu and Chou, 2001) it was flexed upward, its passive margin exposed to erosion, and so shed Paleozoic and Mesozoic sedimentary clasts eroded from the uplift into adjacent basins. This conglomerate is similar in clast lithologies and stratigraphic setting to conglomerates of the Tethyan Himalaya, which are also interpreted to mark flexural uplift and passage of the platform terrace over the outer swell of the trench (Zhang et al., 2012). Thus, in the model of Hildebrand (2009), the collision between Rubia and North America was initiated by the attempted westward subduction of the leading edge of North America. On the other hand, Heller et al. (2003) argued that the conglomerate was dispersed during continental tilting, which resulted from dynamic topography. However, the two are not mutually exclusive.

The beginning of thrusting in the Great Basin segment is constrained to be ~124–120 Ma based on U-Pb zircon dating of ash beds (Fig. 13) just above the gravels (Greenhalgh and Britt, 2007), detrital zircons (Britt et al., 2007), and from a 119.4 ± 2.6 Ma U-Pb age of uraniferous carbonate (Ludvigson et al., 2010). The end of thin-skinned deformation is marked by the transition from the dominantly marine foredeep to localized and sedimentologically isolated, nonmarine basins (Figs. 5, 8, and 14) typical of the thick-skinned Laramide deformation, and is constrained to be between 80 and 70 Ma (Dickinson et al., 1988; Raynolds and Johnson, 2003; Cather, 2004). In Utah, the basal Buckhorn conglomerate contains a detrital zircon profile indistinguishable from Paleozoic–Mesozoic sandstones of the Colorado Plateau and detrital zircons collected up section through the foredeep define inverted chronofacies that document a complete unroofing sequence of allochthonous strata farther west in the Sevier fold-thrust belt (Lawton et al., 2010; Hunt et al., 2011).

In southeastern Idaho, DeCelles et al. (1993) argued that coarse conglomerate of likely Aptian age contained within the finer-grained Bechler Formation represents material shed from the advancing Meade-Paris thrust system, the main thrust faults that carry the thick Neoproterozoic–lower Paleozoic megathrust sheets over the North American platform margin (Fig. 15). In central Utah, the Pavant thrust is the structurally lowest thrust carrying rocks of the megathrust sheets over the North American platform (DeCelles and Coogan, 2006), and it is interpreted to have shed coarse debris that constitute the Aptian–Albian San Pitch Formation (DeCelles et al., 1995).

Some workers (DeCelles and Currie, 1996; DeCelles, 2004; Fuentes et al., 2010) argued that Jurassic rocks of the Fernie Basin in Canada and the Morrison Formation in the western United States are backbulge deposits related to a “phantom foredeep” as suggested originally by Royse (1993b). Hildebrand (2009) recognized this is a possibility, but found it untestable, and thus unsatisfactory, as every trace of the foredeep was eradicated. Furthermore, there is an unaccounted for 25 Myr gap between sedimentation of the Morrison and deposition of the foredeep. During this interval, the Morrison basin was inverted and dissected, with up to 50 m of local relief and local bone-bed accumulations (Eberth et al., 2006). The few Triassic–Jurassic–age detrital zircons in the formation (Fuentes et al., 2009) could easily have been derived from reworking of air-fall tuff units, of which there are many throughout the basin (Kowallis et al., 1998, 2007).

Within the Great Basin sector, a major belt of thrust faults that root in a décollement located in Cambrian shales, transported rocks of the North American platform eastward over the craton and also carried thick Neoproterozoic successions within the so-called megathrust sheets over the platformal rocks (DeCelles, 2004; DeCelles and Coogan, 2006). On the southeastern side of the Snake River Plain the Sevier belt forms a broad salient that continues southward through Idaho, westernmost Wyoming, and northeastern Utah (Fig. 8). The sector contains at least eight major thrust systems and is the only area along the U.S. portion of the fold-thrust belt where the North American platform edge wasn’t overridden by thrust sheets carrying rocks of the Antler platform and its thick section of Neoproterozoic terrigenous clastic rocks (Peterson, 1977; Rose, 1977; Palmer and Hintze, 1992). Western thrusts, such as the Paris and Meade thrusts (Fig. 15), carry typically thick sections of Proterozoic sedimentary rocks, whereas other, more easterly thrusts root in a detachment within Cambrian shale (Armstrong and Cressman, 1963; Armstrong and Oriel, 1965; Royse et al., 1975; Lamerson, 1982; Royse, 1993a).

One of the thrust systems, the Ogden, has an antiformal duplex of Paleoproterozoic crystalline basement that now constitutes the Farmington complex (Bryant, 1984; Yonkee, 1992; Yonkee et al., 1989, 2003; Andreasen et al., 2011). Crystalline basement beneath this part of the thrust belt comprises Archean rocks of the Wyoming craton (Foster et al., 2006). The Farmington complex occurs west of the shelf edge in northeastern Utah (Rose, 1977) and structurally beneath the Paris thrust. The band of Paleoproterozoic rocks likely continues northward into Idaho, where Paleoproterozoic crystalline basement occurs within the Cabin–Medicine Lake system just east of the Idaho batholith (Skipp and Hait, 1977; Skipp, 1987) and in the Tendoy Range of southwestern Montana (DuBois, 1982). Palinspastic restoration of the shortening within the thrust belt restores rocks of the Farmington complex well to the west of North American crust, as indicated by Sr isotopes (Armstrong et al., 1977; Fleck and Criss, 1985), basement windows in the Ruby and nearby ranges, and xenoliths (Evans et al., 2002). That, coupled with isotopic and geological evidence that Archean basement likely continues to the edge of the craton (Hanan et al., 2008), indicate that the Farmington Canyon complex is exotic with respect to its present location.

Just south of Provo, Utah, is another eastward reentrant, named the Charleston-Nebo salient (Fig. 8), where thrusts of the

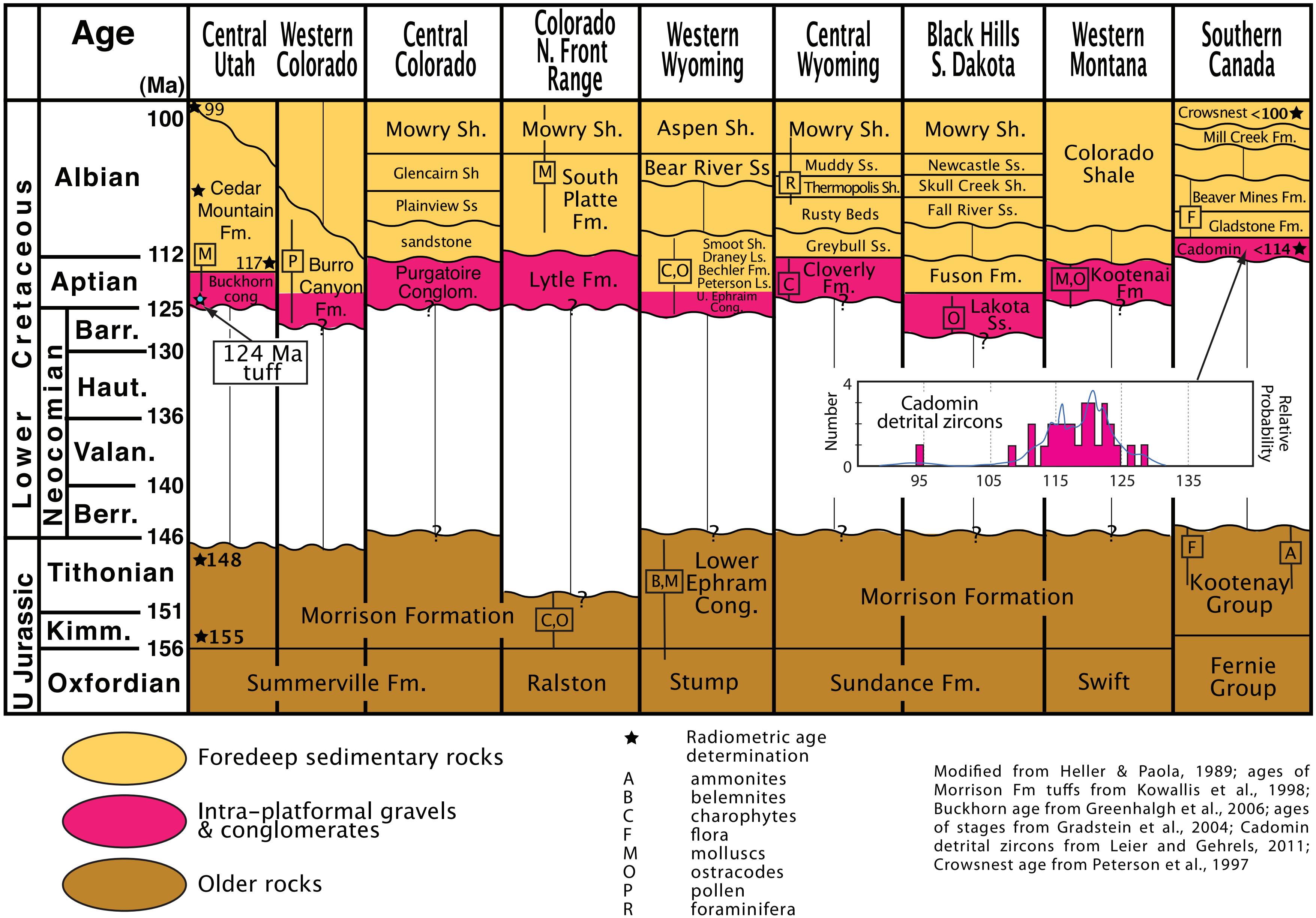
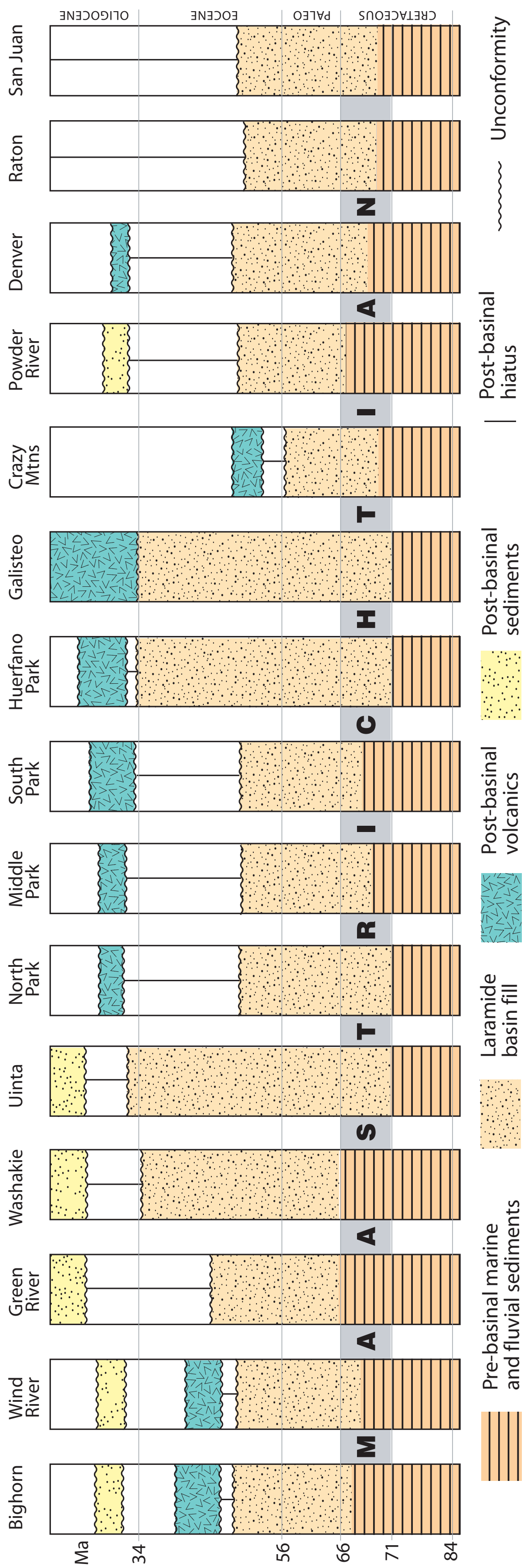


Figure 13. Stratigraphic correlation diagrams for the lower part of the Cordilleran foredeep and platform-derived gravels, modified from Heller and Paola (1989), illustrating the 125 Ma age for the gravels in the western United States and younger age in Canada north of the Sweetgrass arch.



Charleston-Nebo system carry a large, overturned, almost recumbent, anticline composed of thick sequences of Pennsylvanian–Permian rocks not present on the North American platform, and lesser amounts of Paleoproterozoic crystalline basement known as the Santequin complex (Tucker, 1983). Thrusts farther west, such as the Sheeprock, carry thick sections of Precambrian clastic rocks (Christie-Blick, 1982, 1983, 1997; Rodgers, 1989). Large areas of the eastern part of the thrust belt are buried by synorogenic sedimentary rock, and the entire area west of the Wasatch Mountain front was severely disrupted by Cenozoic normal faults.

There are four major thrust systems in south-central Utah, and the westernmost, the Canyon Range–Wah Wah–Pavant system, carries 4–10 km of dominantly siliciclastic rocks of Neoproterozoic age and as much as 12 km of Paleozoic strata, whereas the North American platform to the east is only ~1.5 km thick (Hintze, 1988). The Canyon Range thrust is the type "megathrust" of DeCelles (2004) and DeCelles and Coogan (2006). Like the Santaquin embayment to the north, 50 km of the frontal fold-thrust belt in south-central Utah is also dominantly buried by orogenic deposits (DeCelles, 2004) and broken by younger normal faults. The eastern faults in the area also root in a décollement in Cambrian shale as do those farther north (Lawton et al., 1997; DeCelles et al., 1995).

In the Las Vegas area, there are several major thrust systems. The structurally lowest Wilson Cliffs thrust places Cambrian carbonates and clastics atop eolian Aztec Sandstone of the North American platform (Burchfiel et al., 1974a, 1998). Just to the west, thick sequences of Neoproterozoic sedimentary rocks, collectively known as the Pahrump Group, along with their Paleozoic cover, sit unconformably on crystalline basement in a series of thrusts (Burchfiel et al., 1974a, 1974b; Brady et al., 2000; Snow, 1992; Wernicke et al., 1988).

Within the Great Basin sector, the emplacement of the megathrust sheets and their thick successions of Neoproterozoic sedimentary rocks onto the North American platform took place mainly during the Aptian–Cenomanian (124–94 Ma) as deduced from coarse sedimentary packages that either overlap the thrusts, such as the Canyon Range conglomerate, which overlies the Canyon Range thrust (DeCelles and Coogan, 2006; Lawton et al., 2007), or synthrusting deposits such as the Belcher conglomerate, which was overrun by the earliest thrusting of the Meade thrust (DeCelles et al., 1993). Within Utah, rocks that are considered proximal foredeep deposits are included in the

Figure 14. Chronostratigraphic diagrams for Laramide basins, showing age of inception, continuation, and termination of individual basins during the Laramide phase of the Cordilleran orogeny. The older, and more continuous foredeep basin formed during the thin-skinned Sevier phase of deformation was disrupted, and the younger isolated basins of the Laramide phase formed during thick-skinned deformation, generated during the Laramide collision between North America and the Rubian ribbon continent. Note that all basins formed during the Maastrichtian. Adapted from Dickinson et al. (1988).

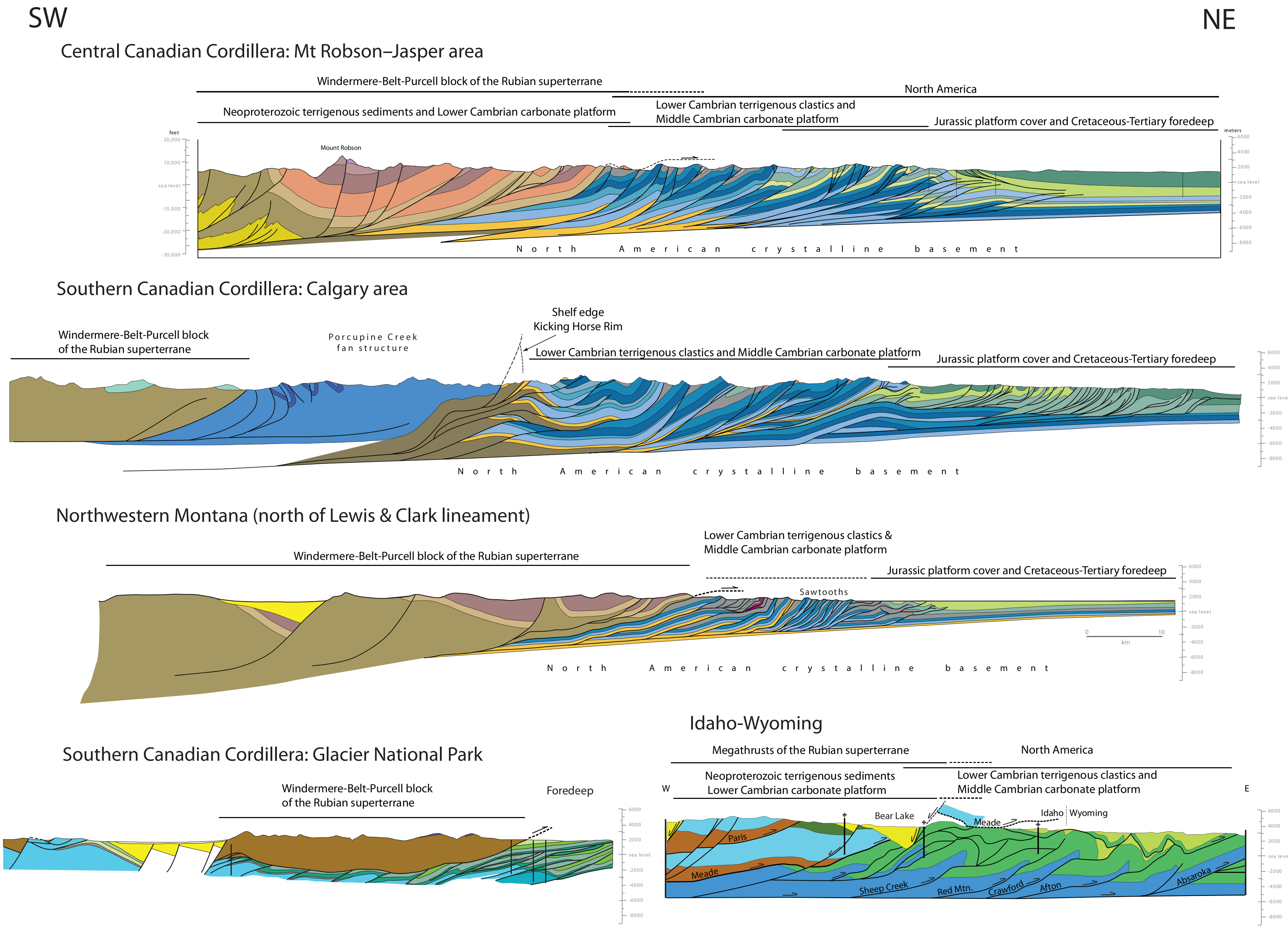

SW
NE
Central Canadian Cordillera: Mt Robson–Jasper area
Windermere-Belt-Purcell block of the Rubian superterrane
North America
Neoproterozoic terrigenous sediments and Lower Cambrian carbonate platform
Lower Cambrian terrigenous clastics and Middle Cambrian carbonate platform
Jurassic platform cover and Cretaceous-Tertiary foredeep
Mount Robson
feet
20,000
10,000
sea level
-10,000
-20,000
-30,000
meters
6000
4000
2000
sea level
-2000
-4000
-6000
-8000
North American crystalline basement
Southern Canadian Cordillera: Calgary area
Windermere-Belt-Purcell block of the Rubian superterrane
Porcupine Creek fan structure
Shelf edge
Kicking Horse Rim
Lower Cambrian terrigenous clastics and Middle Cambrian carbonate platform
Jurassic platform cover and Cretaceous-Tertiary foredeep
North American crystalline basement
Northwestern Montana (north of Lewis & Clark lineament)
Windermere-Belt-Purcell block of the Rubian superterrane
Lower Cambrian terrigenous clastics & Middle Cambrian carbonate platform
Jurassic platform cover and Cretaceous-Tertiary foredeep
Sawtooths
0
10
km
North American crystalline basement
Idaho-Wyoming
Megathrusts of the Rubian superterrane
North America
Neoproterozoic terrigenous sediments Lower Cambrian carbonate platform
Lower Cambrian terrigenous clastics and Middle Cambrian carbonate platform
W
E
Bear Lake
Meade
Idaho
Wyoming
Paris
Meade
Sheep Creek
Red Mtn.
Crawford
Afton
Absaroka
Southern Canadian Cordillera: Glacier National Park
Windermere-Belt-Purcell block of the Rubian superterrane
Foredeep

Figure 15. Comparative cross sections through the Rocky Mountain and Sevier fold-thrust belts illustrating the similarity of the westernmost thrust sheets in that they all carry much thicker sections of Precambrian–Paleozoic rocks, typical of the megathrust sheets up and down the Cordillera, which if derived from the outermost parts of North America, would require huge crustal ramps. In this paper, it is argued that rocks of the megathrust sheets are exotic with respect to North America. Mount Robson–Jasper area after Mountjoy (1979); Calgary area from Price and Fermor (1985); Northwestern Montana simplified from Fuentes et al. (2012); Glacier National Park (Canada) modified from Price (1962); Idaho-Wyoming section modified from DeCelles et al. (1993).

Indianola Group and Cedar Mountain Formation (DeCelles and Coogan, 2006; Hunt et al., 2011, and references therein). The uppermost member of the Cedar Mountain Formation, known as the Mussentuchit, is well dated radiometrically as earliest Cenomanian and lies directly beneath the much finer-grained and basin-wide Dakota Formation (Cifelli et al., 1997; Garrison et al., 2007; Biek et al., 2009). The lower part of the Canyon Range conglomerate is poorly dated, but Lawton et al. (2007) correlated a distinctive conglomerate rich in carbonate clasts with the lower-most member of the San Pitch Formation, located just to the east and containing palynomorphs ranging in age from mid- to upper-Albian (Sprinkel et al., 1999). Thus, the dominant period of emplacement of the megathrust sheets in the Great Basin sector took place during the Aptian–Cenomanian, 125–94 Ma, but is possibly no younger than mid-Albian (~105 Ma), if the Canyon Range–San Pitch correlation is correct. Sedimentary rocks of the upper Dakota contain significant detrital zircon peaks of 121, 116, and 110 Ma (Ludvigson et al., 2010) and reflect an entirely different source than the older and more proximal deposits, which as mentioned earlier, contain an inverted detrital zircon provenance derived from the Neoproterozoic and lower Paleozoic rocks of the megathrust sheets (Lawton et al., 2010; Hunt et al., 2011).

The Hinterland Belt

Lying directly west of the megathrust sheets and their Neoproterozoic sedimentary rocks within the Great Basin segment is the hinterland belt (Fig. 8). Ever since Armstrong (1968) recognized the belt, its origin has proved elusive, as it contains polydeformed and low-grade sedimentary rocks, high-grade metamorphic rocks, several ages of crystalline basement, and both metaluminous and peraluminous intrusions—all cut by thrust and normal faults of significant displacement. Its eastern and western boundaries are somewhat obscure owing to at least two major periods of extensional faulting and variable exhumation; but in a general sense the hinterland is a northerly trending strip characterized by Paleocene–Eocene core complexes, both Jurassic and Cretaceous thrust faults and metamorphism, dominantly westerly-vergent Jurassic folds, and generally sparse Jurassic–Cretaceous plutons.

Rocks of the hinterland are exposed in Paleocene–Eocene core complexes found in the Albion, Raft River, and Grouse Creek ranges in northeastern Nevada, northwestern Utah, and southern Idaho in what can best be termed the type area (Armstrong, 1968; Snoke, 1980; Howard, 1980; Todd, 1980; Snoke and Miller, 1988; Wells, 1992). Present-day structural relief within the hinterland is visible because of Paleocene–Eocene extensional collapse of the thickened and hot hinterland zone as well as even younger Basin and Range extension.

The rocks in the Albion–Raft River–Grouse Creek ranges (Fig. 8) are divided into an autochthon comprising Archean crystalline basement unconformably overlain by a thin veneer of quartzite and pelitic schist (Compton, 1972), structurally overlain by greenschist-amphibolite grade metasedimentary rocks of Paleozoic age (Wells et al., 1997). Within the autochthon, the deformation—as evidenced by small-scale structures—decreases downward from the basal thrust, whereas metamorphism postdates, or was possibly synchronous with, thrusting (Compton, 1980; Miller, 1980; Snoke and Miller, 1988). Early Cretaceous and Jurassic granites, present in the overlying allochthons, have not been described within the autochthon. The Archean crystalline basement of the autochthon is interpreted to represent cratonic North America of the Wyoming province (Miller, 1980; Snoke and Miller, 1988).

Rocks of the area generally contain evidence for two pulses of deformation, the first during the Late Jurassic, and the second during the Late Cretaceous (Camilleri et al., 1997; McGrew et al., 2000). Plutons of Jurassic age occur within the allochthons. The Jurassic magmatism overlapped in time with minor folding, thrusting, and development of local metamorphic aureoles adjacent to the intrusions (Camilleri et al., 1997).

The second deformational and metamorphic event is more intense and pervasive. Thrust faulting during this contractional pulse caused at least 70 km of shortening and as much as 30 km of crustal thickening (Camilleri et al., 1997). Metamorphic assemblages within the area indicate deep burial and metamorphism, perhaps as early as Late Jurassic, followed by higher temperature Late Cretaceous peak metamorphism at ~85 Ma with temperatures of 800 °C and pressures of >9 kb followed by a steep uplift path (McGrew et al., 2000). Exhumation was largely complete by the Eocene, when magmatic rocks intruded and overstepped extensional detachments (Miller et al., 1987; Camilleri, 1992).

The Ruby Range–East Humboldt Mountains (Fig. 8) just to the southwest in Nevada also contain rocks that record a complicated history of two Mesozoic orogenic events: (1) 153 Ma plutonic emplacement, polyphase folding, and upper amphibolite facies metamorphism; and (2) Late Cretaceous migmatization, metamorphism, and deformation (Snoke and Miller, 1988;

Hudec, 1992; McGrew et al., 2000). Cretaceous migmatitic upper amphibolite facies rocks are tectonically stacked and include a local recumbent isoclinal fold cored by Archean basement and a structurally overlying section of Neoproterozoic to Mississippian sedimentary rocks possibly sitting on Proterozoic gneiss (Howard et al., 1979; Lush et al., 1988; McGrew et al., 2000). Migmatization was synkinematic with nappe emplacement at 84.8 ± 2.8 Ma, and resulted from peak metamorphic conditions of 9–10 kb pressure and 750–800 °C (Hodges et al., 1992; McGrew et al., 2000). The initiation of exhumation and uplift is not precisely dated, but exhumation spans the range 63 to 50 Ma (Snoke and Miller, 1988).

South of the Ruby Range lies the massive Snake Range (Fig. 8), which appears to be structurally simpler but still contains evidence for two periods of plutonism and metamorphism, one at ~160 Ma and the other with peak metamorphism at 79 Ma, and initial exhumation between 57 and 50 Ma (Snoke and Miller, 1988; McGrew et al., 2000). Allochthonous Neoproterozoic and Paleozoic sequences in both the Snake Range and the East Humboldt–Ruby ranges appear to be similar to those in the megathrust sheets of the thrust belt.

The hinterland belt continues southward into the Death Valley area (Fig. 8), where rocks of the hinterland were metamorphosed during the Cretaceous with peak metamorphic conditions of ~620–680 °C and 7–9 kbar at 91.5 ± 1.4 Ma, followed by Late Cretaceous or early Tertiary extension, typical of other areas within the hinterland (Hodges and Walker, 1990, 1992; Applegate and Hodges, 1995; Mattinson et al., 2007).

Central Nevada

West of the megathrust slabs and the hinterland belt, and within the tectonic collage of generally recognized exotic allochthons, the most easterly assemblage of rocks are platformal Early Cambrian to Devonian limestones and siliciclastic rocks of the Antler shelf (Poole et al., 1977). Rocks of the platform were overthrust by additional terranes generally considered to be exotic with respect to North America (Silberling et al., 1992; Oldow et al., 1989). Rocks of the Roberts Mountain allochthon (Fig. 5), which comprises a structurally complex allochthonous stack of Cambrian–Devonian siltstone, chert, argillite, barite, and mafic volcanic rocks (Fig. 16), were emplaced upon the Rubian margin during the Antler orogeny, which occurred during the Late Devonian–Early Mississippian (Merriam and Anderson, 1942; Smith and Ketner, 1968; Poole and Sandberg, 1977; Nilsen and Stewart, 1980; Johnson and Pendergast, 1981; Johnson and Visconti, 1992; E.L. Miller et al., 1992b). During and after emplacement of the allochthons, coarse debris was shed eastward to form a clastic wedge over the pre-collisional Antler shelf (Poole, 1974, 1977; Harbaugh and Dickinson, 1981; Speed and Sleep, 1982).

Rocks of the Roberts Mountain allochthon also occur in the Pioneer Mountains north of the Snake River Plain in Idaho (Wilson et al., 1994; Link et al., 1996) and may continue northward into Canada. Several workers (Turner et al., 1989; Smith and Gehrels, 1992a, 1992b; Smith et al., 1993; Root, 2001) suggested that Paleozoic rocks and Middle to Late Devonian deformation and concomitant development of an orogenic foredeep extending from northern Washington to the Mackenzie delta in northern Canada were related to the Antler orogeny.

Sitting atop the Roberts Mountain allochthon are rocks of the Havallah sequence—another complexly deformed assemblage of allochthons collectively termed the Golconda allochthon (Fig. 5), and containing Upper Devonian to earliest Triassic chert-argillite sequences with intercalated lenses of pillow basalt, which were emplaced during the Early Triassic Sonoman orogeny (Silberling and Roberts, 1962; Speed, 1977; Silberling, 1975; Dickinson et al., 1983). Rocks that sit unconformably above those of the Roberts Mountain allochthon and beneath those of the Golconda allochthon were termed the Antler overlap sequence by Dickinson et al. (1983) and were deformed prior to the emplacement of the Golconda allochthon (Cashman et al., 2011; Trexler et al., 2003, 2004).

A sequence of 250–248 Ma intermediate to siliceous volcanic rocks, known as the Koipato volcanics, sits unconformably upon rocks of the Golconda allochthon and has Sr and Nd isotopic values suggestive of Paleoproterozoic crust, which led Vetz (2011) to argue that the Golconda allochthon was already emplaced upon the Roberts Mountain rocks to the east when the volcanic rocks were erupted. The short burst of magmatism might be a manifestation of slab break-off following Golconda–Roberts Mountain collision. A west-facing carbonate platform of Triassic age overlies the volcanics, but in the east clastic rocks lie between the two (Oldow, 1984).

White-Inyo Range

The White-Inyo Range lies just east of the Sierra Nevada and forms a mountain range nearly as imposing as the Sierra (Figs. 5 and 7). The range comprises a 7 km-thick section of Paleozoic–Mesozoic rocks (Stewart, 1970) that due to metamorphism and deformation do not obviously continue to the west into the Sierra Nevada but are known to occur in an arcuate band in Esmeralda County, Nevada (Fig. 8) to the east (Albers and Stewart, 1972). Rocks partly of the same age occur in the Death Valley region, and the two packages have been correlated with those in the White-Inyos, but the sections are very different and require facies changes at every stratigraphic interval: facies transitional between the two are absent despite good exposure. Rocks originally to the west of Death Valley appear to presently reside in Sonora, Mexico (Stewart et al., 1984, 2002; Stewart, 2005). The White-Inyo block appears to have been attached and adjacent to the Sierra Nevada block by at least the Jurassic as volcanic units of that age appear to cross Owens Valley along strike (Dunne and Walker, 1993). Similarly, it was likely attached to rocks in the Death Valley region to the southeast by the late Paleozoic as Permian thrust faults (Snow, 1992) appear to form a continuous band across their contact as do Jurassic plutons (Dunne et al., 1978).

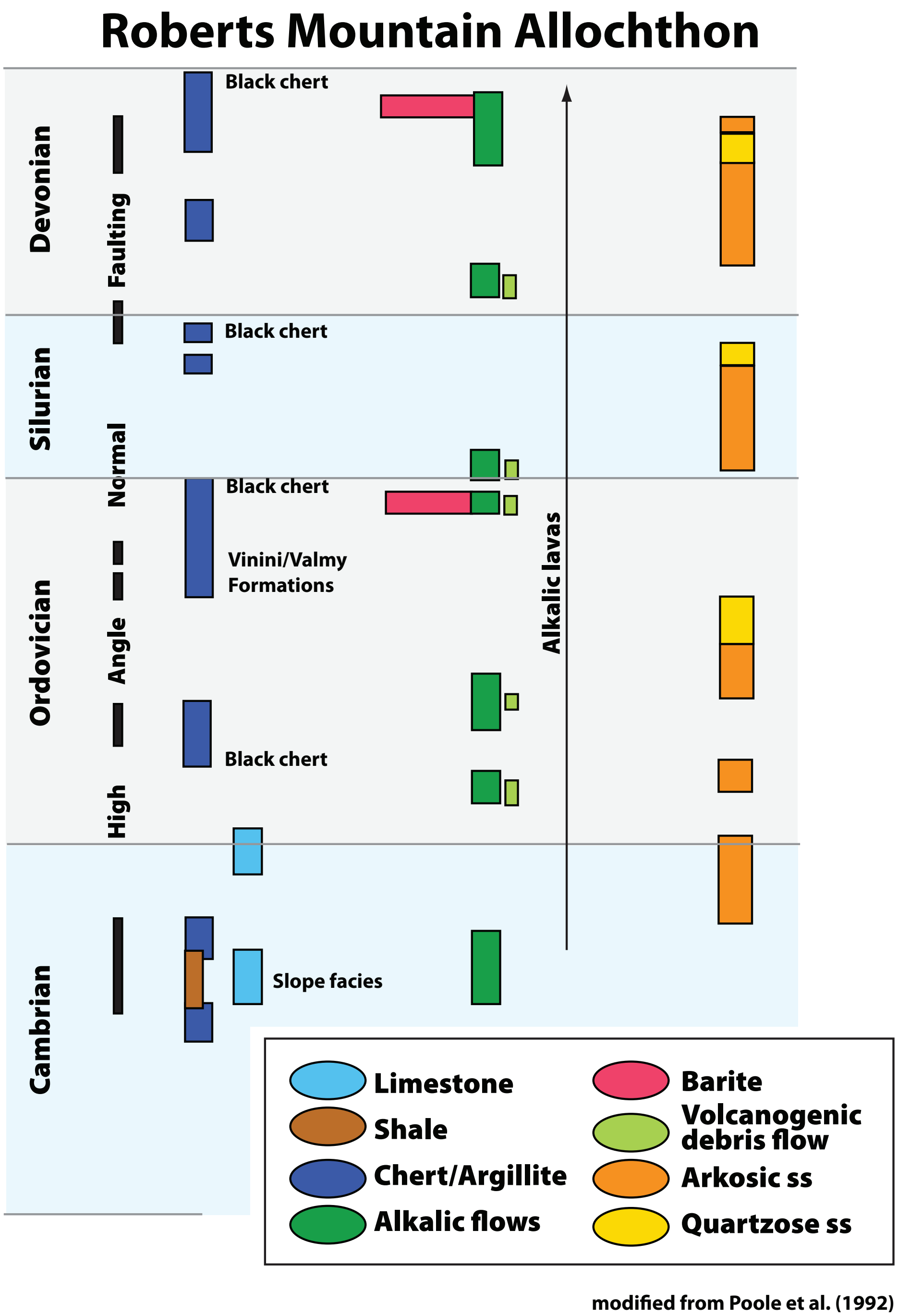


Figure 16. Composite and schematic stratigraphic section illustrating the different lithologies of the Roberts Mountain allochthon. Note the presence of alkalic lavas throughout the entire section, a feature that is incompatible with a cooling and magmatically dead passive margin. As discussed in the text, the lithologies are more typical of those in marginal basins.

Death Valley

In the Death Valley–southern Nevada sector (Fig. 8), rocks of the Pahrump Group and their crystalline basement are allochthonous (Burchfiel et al., 1974a, 1974b; Brady et al., 2000; Snow, 1992). Overlying Ediacaran–Cambrian sedimentary rocks of the Wood Canyon Formation contain detrital zircon peaks at ~1.1 Ga, and the slightly younger Zabriskie Quartzite contains abundant 3.0–3.4 Ga grains (Stewart et al., 2001)—source ages markedly absent in western Laurentia. Higher in the stratigraphic succession is the distinctive Middle to Late Cambrian Bonanza King Formation, which farther north is part of the Antler shelf of central Nevada and westernmost Utah (Kepper, 1981; McCollum and McCollum, 1984; Montañez and Osleger, 1996; Morrow and Sandberg, 2008). This is well west of the North American shelf edge, and rocks of the Bonanza King Formation match poorly with those of the time-correlative Muav Formation of the Colorado Plateau. These rocks and those of the White-Inyos were folded and transported on thrust faults dated to be Permian at ~294–284 Ma (Snow, 1992; Stevens et al., 1998; Stevens and Stone, 2007), an event unknown on the North American platform terrace. In the Spring Mountains near Las Vegas, carbonate rocks of the Bonanza King Formation sit structurally atop rocks of the Aztec Sandstone, a Jurassic eolianite of the North American platform (Burchfiel et al., 1998).

Stratigraphically higher in the sequence and crudely approximating the platform edge of the Bonanza King Formation are the Devonian and Late Ordovician–Silurian carbonate platform edges (Sheehan, 1986; Harris and Sheehan, 1998; Morrow and Sandberg, 2008). This westerly facing shelf-edge facies transition also crops out in east-central Nevada and northwesternmost Utah, where it occurs west of the hinterland belt as part of the Antler shelf (Fig. 1). Similarly, the Pennsylvanian–Early Permian Bird Spring carbonate shelf edge lies in eastern California and faces west (Stevens and Stone, 2007). Overall, the locations of the shelf edges, the non–North American detrital zircons, and the Permian thrust faults indicate that crystalline basement in the Death Valley area, the Pahrump Group, and overlying Paleozoic strata are most likely exotic with respect to North America.

We now jump over the great Mesozoic batholiths and Jurassic arc terranes to describe rocks located in the western Sierra Nevada. The reason for doing so is to provide a framework for discussion of the Cretaceous Sierran batholith by first describing their enveloping wall rocks.

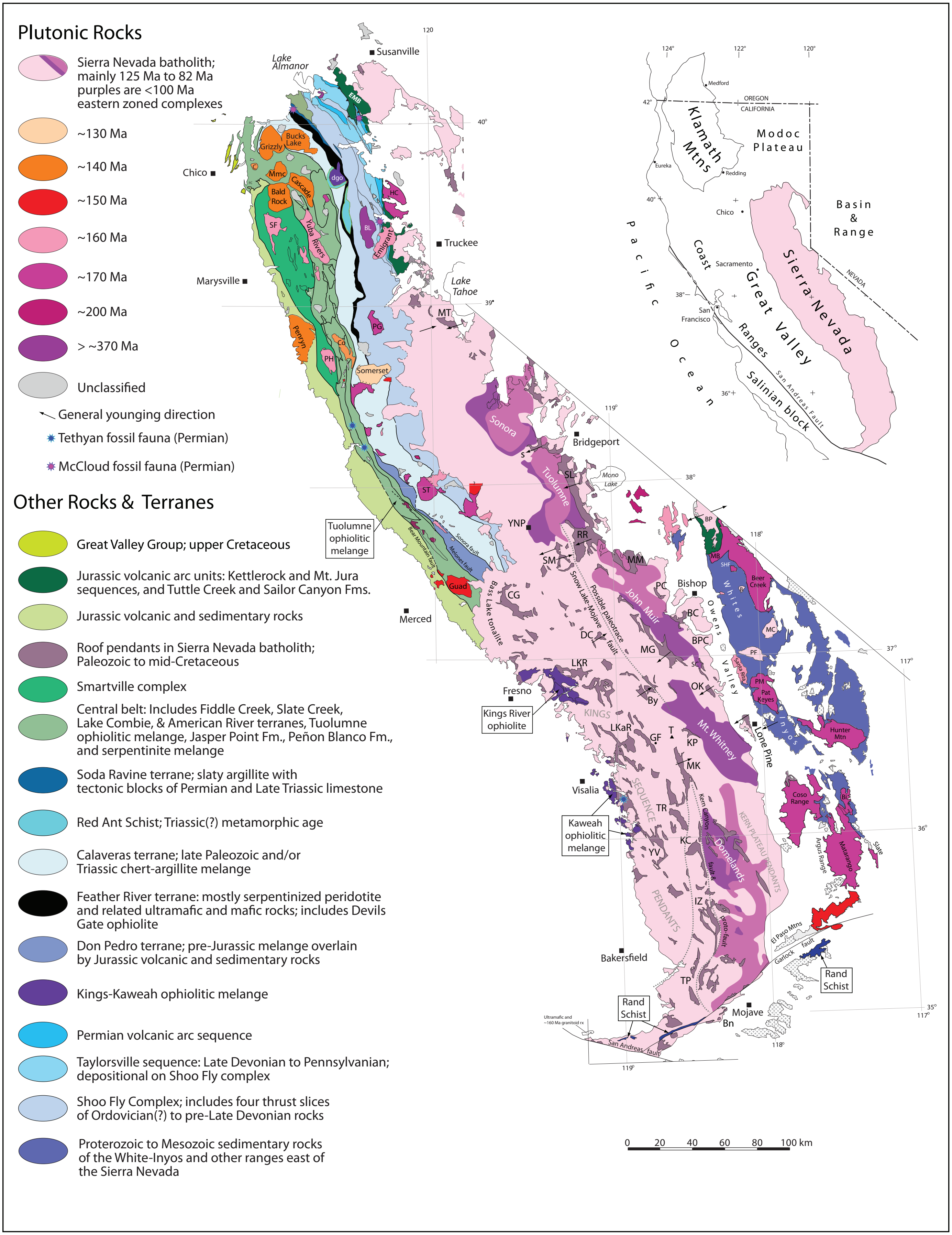
Western Sierra Nevada Metamorphic Belt

The western Sierra Nevada metamorphic belt (Fig. 17) is a collage of amalgamated Paleozoic–Mesozoic arc and subduction complexes that in a general sense young westward (Saleeby et al., 1989; Edelman, 1990), locally contain far-traveled Permian McCloud and Tethyan fauna (Miller, 1987), and are divided by some into four major tectonic belts or terranes (Day et al., 1985). The oldest, and easternmost, terrane is the northern Sierra terrane (Coney et al., 1980), comprising in its lower parts, the Paleozoic Shoo Fly complex (Fig. 17), which contains westerly-vergent thrust sheets of Ordovician to Silurian sedimentary rocks and Devono-Ordovician ophiolitic mélange, cut by the 385–364 Ma Bowman Lake batholith (Hanson et al., 1988; Harwood, 1992). Higher in the section are northeasterly-vergent thrust sheets containing Paleozoic metasedimentary rocks as well as volcanic and volcaniclastic rocks of three possible volcanic arcs: (1) the Devonian to Pennsylvanian Taylorsville sequence; (2) a Permian volcanic sequence; and (3) the upper Triassic to mid-Jurassic Kettle Rock–Mount Jura and Tuttle Creek–Sailor Canyon sequences of the eastern Mesozoic belt, which on their western side, are structurally overlain along a westerly-dipping thrust fault by rocks of the older complexes (Day et al., 1985; Christe and Hannah, 1990; Harwood, 1992; Christe, 2011). The Mesozoic arc sequences are part of the extensive belt of calc-alkaline volcanic rocks that extend along the Sierra and are generally interpreted to have been generated by easterly subduction beneath North America (Burchfiel and Davis, 1972, 1975). Based on the lack of an arc of the appropriate age, paleogeographic considerations, and detrital zircon analyses, Wright and Wyld (2006) argued that the Paleozoic Shoo Fly complex was a peri-Gondwanan terrane that migrated into the Pacific, but when it docked with more easterly terranes remains unresolved.

To the west of the Northern Sierra terrane is the fault-bounded Feather River peridotite (Fig. 17), a Paleozoic–Mesozoic suture zone (Edelman et al., 1989b) of metamorphosed and tectonized peridotite, dunite, serpentinite, and lesser amounts of metagabbro, amphibolite, and metasedimentary rocks, some of which are fault slices of lawsonitic blueschist known as Red Ant schist (Mayfield and Day, 2000; Schweickert et al., 1980; Hietanen, 1981). The fault along the western margin of the peridotite belt dips steeply eastward and juxtaposes the peridotite above the Calaveras mélange belt, but westward overturning of folds suggested to Day et al. (1985) that the bounding fault is overturned and originally dipped shallowly westward such that the mélange was thrust eastward over the peridotite belt. Hornblende ages from schists along the fault range from 345 to 235 Ma with some high-P metamorphism, so constitute a metamorphic sole, possibly formed at the initiation of intra-oceanic subduction (Smart and Wakabayashi, 2009).

West of the Feather River peridotite lies the Central belt (Fig. 17), which is a disrupted, isoclinally-folded, and variously metamorphosed amalgam of ultramafic, plutonic, volcanic, and sedimentary rocks (Day et al., 1985; Dilek, 1989). In addition to sedimentary mélange and broken formation of the Calaveras

Figure 17. Geological sketch map of the Sierra Nevada and Western metamorphic belt showing locations and geological features discussed in the text. Modified from Irwin and Wooden (2001) with additional data from Bateman (1992), Saleeby and Busby-Spera (1993), Dunne et al. (1978), and Saleeby et al. (1978). Abbreviations: B—Birch Creek pluton; BC—Bishop Creek pendant; Bi—Bendire pluton; BL—Bowman Lake batholith; Bn—Bean Canyon pendant; BP—Boundary Peak pluton; BPC—Big Pine Creek pendant; By—Boyden Cave pendant; C—Concow pluton; CG—Coarse Gold pendant; Co—Colima pluton; DC—Dinkey Creek pendant; dgo—Devil's Gate ophiolite; EMB—Eastern Mesozoic Belt; GF—Giant Forest; Guad—Guadalupe pluton; HC—Haypress Creek pluton; IZ—Isabella Lake pendants; KC—Kern Canyon pendant; KP—Kaweah Peaks pendant; LKaR—Lower Kaweah River pendants; LKR—Lower Kings River pendants; MB—Mount Barcroft pluton; MC—Marble Canyon pluton; MG—Mount Goddard pendant; MK—Mineral King pendant; MM—Mount Morrison pendant; Mmc—Merrimac pluton; MT—Mount Tallac pendant; OK—Oak Creek pendant; PC—Pine Creek pendant; PF—Papoose Flat pluton; PG—Pino Grande pluton; PH—Pine Hill complex; PM—Paiute Monument pluton; RR—Ritter Range pendant; S—Snow Lake pendant; SF—Swede Flat pluton; SHF—Sage Hen Flat pluton; SL—Saddlebag Lake pendant; SM—Strawberry Mine pendant; ST—Standard pluton; T—Triple Divide peak; TP—Tehachapi pendants; TR—Tule River pendants; YNP—Yosemite National Park; YV—Yokohl Valley pendant.


Plutonic Rocks
Sierra Nevada batholith; mainly 125 Ma to 82 Ma purples are <100 Ma eastern zoned complexes
~130 Ma
~140 Ma
~150 Ma
~160 Ma
~170 Ma
~200 Ma
> ~370 Ma
Unclassified
General younging direction
Tethyan fossil fauna (Permian)
McCloud fossil fauna (Permian)
Other Rocks & Terranes
Great Valley Group; upper Cretaceous
Jurassic volcanic arc units: Kettlerock and Mt. Jura sequences, and Tuttle Creek and Sailor Canyon Fms.
Jurassic volcanic and sedimentary rocks
Roof pendants in Sierra Nevada batholith; Paleozoic to mid-Cretaceous
Smartville complex
Central belt: Includes Fiddle Creek, Slate Creek, Lake Combie, & American River terranes, Tuolumne ophiolitic melange, Jasper Point Fm., Peñon Blanco Fm., and serpentinite melange
Soda Ravine terrane; slaty argillite with tectonic blocks of Permian and Late Triassic limestone
Red Ant Schist; Triassic(?) metamorphic age
Calaveras terrane; late Paleozoic and/or Triassic chert-argillite melange
Feather River terrane: mostly serpentinized peridotite and related ultramafic and mafic rocks; includes Devils Gate ophiolite
Don Pedro terrane; pre-Jurassic melange overlain by Jurassic volcanic and sedimentary rocks
Kings-Kaweah ophiolitic melange
Permian volcanic arc sequence
Taylorsville sequence: Late Devonian to Pennsylvanian; depositional on Shoo Fly complex
Shoo Fly Complex; includes four thrust slices of Ordovician(?) to pre-Late Devonian rocks
Proterozoic to Mesozoic sedimentary rocks of the White-Inyos and other ranges east of the Sierra Nevada
Susanville
Lake Almanor
Chico
Marysville
Truckee
Lake Tahoe
Bridgeport
Mono Lake
Merced
Fresno
Bishop
Visalia
Lone Pine
Bakersfield
Mojave
Tuolumne ophiolitic melange
Kings River ophiolite
Kaweah ophiolitic melange
Rand Schist
Sonora
Tuolumne
John Muir
Mt. Whitney
Domelands
Owens Valley
Whites
Inyos
Coso Range
Argus Range
Slate
Hunter Mtn
Bass Lake tonalite
Snow Lake-Mojave fault
Possible paleotrace
Kern Canyon fault
Proto-fault
Garlock fault
San Andreas fault
El Paso Mtns
Melones fault
Sonora fault
Bear Mountains fault
KINGS SEQUENCE
KERN PLATEAU PENDANTS
PENDANTS
Klamath Mtns
Modoc Plateau
Basin & Range
Pacific Ocean
Coast Ranges
Great Valley
Sierra Nevada
Salinian block
San Andreas Fault
Medford
OREGON
CALIFORNIA
Eureka
Redding
Sacramento
San Francisco
NEVADA
0 20 40 60 80 100 km

assemblage, there are two fault-bounded complexes containing both ultramafic rocks and volcano-plutonic successions interpreted as arcs: the Lake Combie and Slate Creek complexes (Fig. 18).

The Calaveras assemblage is a structural sequence of basaltic to andesitic pillowed lava and hyaloclastite, chert, volcaniclastic rocks, phyllite, and Permo-Carboniferous marble blocks that sit beneath the Feather River ophiolite and the Foothills suture (Hietanen, 1981; Schweickert and Bogen, 1983; Hacker, 1993). Limestone within the Calaveras contains McCloud fauna (Standlee and Nestell, 1985). Recent detrital zircon analysis suggests that the mélange could be younger than 159–150 Ma, the age of the youngest detrital age peaks from units in the western part of the assemblage (Van Guilder et al., 2010).

The Jurassic Slate Creek complex (Figs. 17 and 18), exposed as a fault-bounded unit west of the Calaveras complex and mainly atop the Fiddle Creek complex, comprises a tripartite pseudostratigraphy of (1) a basal zone of serpentinite matrix mélange holding blocks of plutonic, volcanic, and metasedimentary rocks; (2) a central plutonic interval consisting of amphibolitic gabbro, metadiabase, and tonalite; and (3) an upper volcanic unit of aphyric to augite porphyritic greenstone, tuff and locally derived volcaniclastic rocks (Day et al., 1985; Edelman et al., 1989a, 1989b; Fagan et al., 2001). It sits structurally upon rocks of the Fiddle Creek complex (Figs. 17 and 18), and the thrust fault is cut by the 167 Ma Scales pluton (Day and Bickford, 2004). Metaplutonic and metavolcanic rocks range in age from ~209 Ma to 172 Ma, whereas a suite of younger plutons (Fig. 18) ranges in age from 160 to 150 Ma (Edelman et al., 1989a, 1989b; Saleeby et al., 1989; Fagan et al., 2001; Day and Bickford, 2004).

The Fiddle Creek complex consists of two distinct associations: (1) ophiolitic mélange with abundant blocks of ophiolite and diorite cut by dioritic dikes; and (2) chert-argillite units, volcaniclastic sandstones and olistostromes holding blocks of amphibolite, marble, and scarce pillow basalt (Dilek, 1989; Edelman et al., 1989b). Rocks of the area are poorly dated but within the second group apparently range from Middle Triassic to ~174 Ma (Hietanen, 1981; Hacker, 1993).

The Lake Combie complex (Fig. 18) is another fault-bounded belt of Jurassic rocks crudely similar to, and generally correlated with, the Slate Creek complex, in that it contains a foliated and lineated basal ultramafic unit, tectonically overlain by gabbro to quartz dioritic intrusions, and an upper unit, more than 5 km thick, consisting of mafic flows at the base grading upwards to dominantly tuff, flow-breccia, and volcaniclastic rocks (Day et al., 1985).

The westernmost terrane, the Smartville complex (Figs. 17 and 18), is a partial ophiolite, with serpentinized ultramafic rocks, gabbro, pillowed basalts, and sheeted dikes—all overlain by an arc suite, 1.5–2 km thick, of pyroxene andesite tuff breccia, basaltic to andesitic flows and pillow lavas with minor dacitic extrusives, sandstones, and conglomerate (Xenophontos and Bond, 1978; Menzies et al., 1980). The volcanic rocks are generally divided into a lower tholeiitic mass and an upper Middle Jurassic calc-alkaline suite intruded by an extensive sheeted dike swarm between ~163 and 159 Ma (Beard and Day; 1987; Saleeby et al., 1989; Day and Bickford, 2004). The sheeted dikes are interpreted to indicate an extensional period within the arc (Beard and Day, 1987; Dilek, 1989). The complex forms a large hanging-wall anticline on an eastwardly vergent thrust that places the complex upon mélange and broken formation of the Central belt (Day et al., 1985; Moores, 2011, personal commun.). Based on fossils, overlap succession, and a dated tuff, volcanic rocks in the upper part of the complex are Oxfordian–Kimmeridgian, slightly older than the 157 Ma Yuba Rivers pluton (Xenophontos, 1984; Saleeby et al., 1989; Day and Bickford, 2004), which sits along the thrust between the Smartville and Slate Creek–Combie belts and apparently metamorphosed rocks of the Smartville complex (Bobbitt, 1982). Therefore, the collision of the Smartville block with rocks located to the east occurred at ~162 Ma (Day and Bickford, 2004) and magmatism within the arc continued up to the time of collision.

Farther south, but along strike with the Smartville Complex and probably part of the same arc (Fig. 17), are additional thick sequences of Jurassic arc rocks—the Peñon Blanco, Logtown Ridge, and Jasper Point formations—that sit on ophiolitic basement, are augite porphyritic, and also occupy a hanging-wall anticline (Bogen, 1985). They appear related to a sparse suite of similar age 170–160 Ma peridotite-diorite intrusive complexes that occur along the western Sierran foothills and are interpreted as products of arc magmatism (Snoke et al., 1982). These rocks are cut by the 153–151 Ma Guadalupe Igneous complex (Ernst et al., 2009b; Haeussler and Paterson, 1993; Saleeby et al., 1989). This presents a problem because fine-grained metasedimentary rocks of the highly deformed Mariposa Formation are thought to interfinger with the volcanics (Bogen, 1985; Snow and Ernst, 2008), yet many detrital zircons collected and analyzed from the metasedimentary rocks are in the 155 to 152 Ma range making the intrusive complex syn-Mariposa (Ernst et al., 2009b); and, in fact, detrital zircons from the inferred lowermost sandstone units, which are upsection from the intercalated volcanic units, yielded youngest zircons of 160 Ma. This suggests that there are two superposed basins or possibly a significant intra-basinal hiatus.

Klamath Mountains

The Klamath Mountains (Figs. 5, 8, and 19) are reasonably well known and constitute an isolated block within northwesternmost California and southwestern Oregon. They comprise an imbricate stack of terranes separated from one another by thrust faults, generally considered to be easterly dipping (Irwin, 1981). The terrane concept was hatched here by Irwin (1972), who more recently divided the Klamaths into three terranes, Eastern, Central and Western, each of which are themselves subdivided into several subterranes and generally correlated with rocks of the western Sierra Nevada (Irwin, 1994, 2003).

The Eastern Klamath terrane comprises three subterranes (Fig. 19): Yreka, Trinity, and Redding and contains the oldest rocks

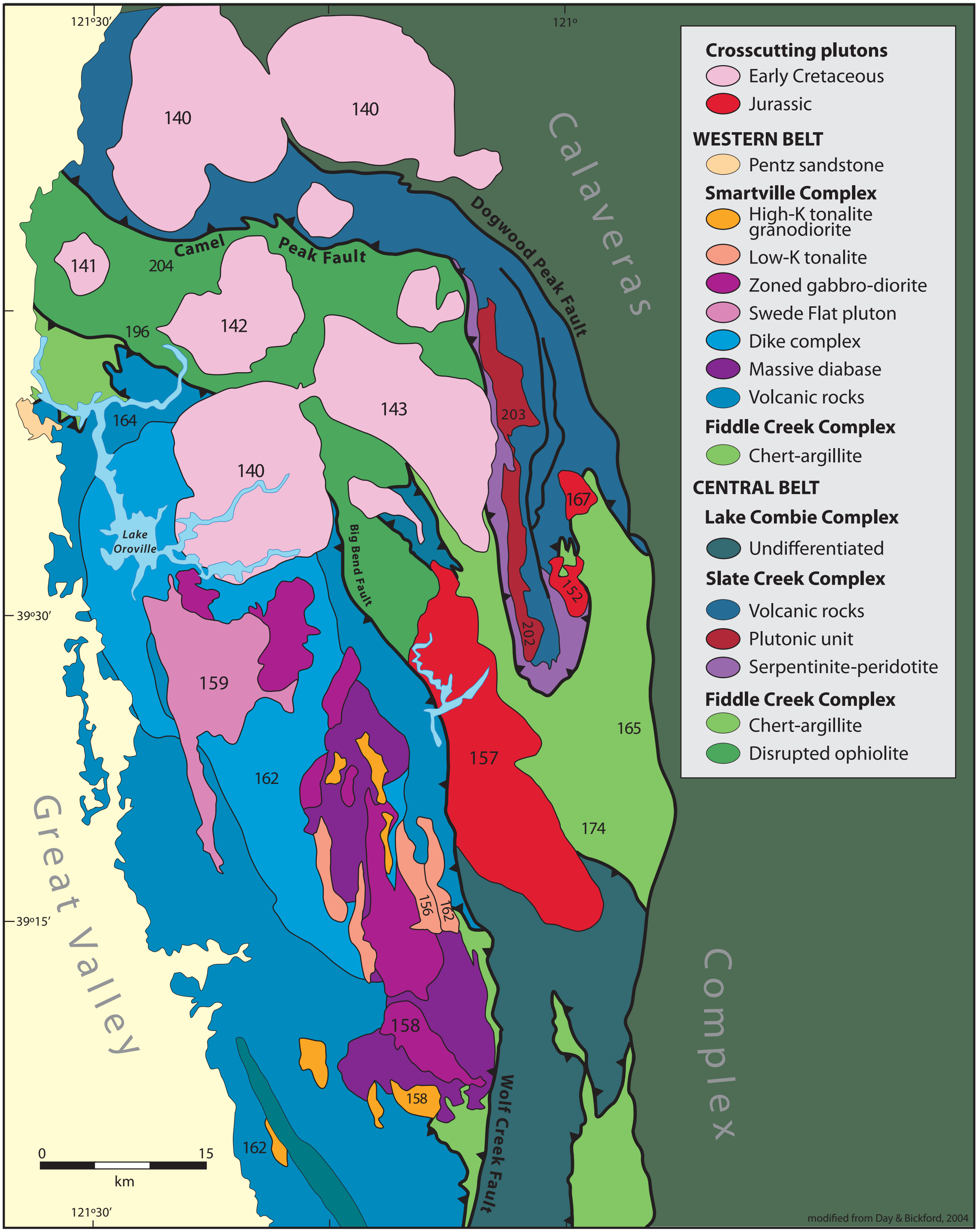


Figure 18. Geological sketch map showing the geology of the northern Western and Central belts, Western metamorphic belt. Modified after Day and Bickford (2004).

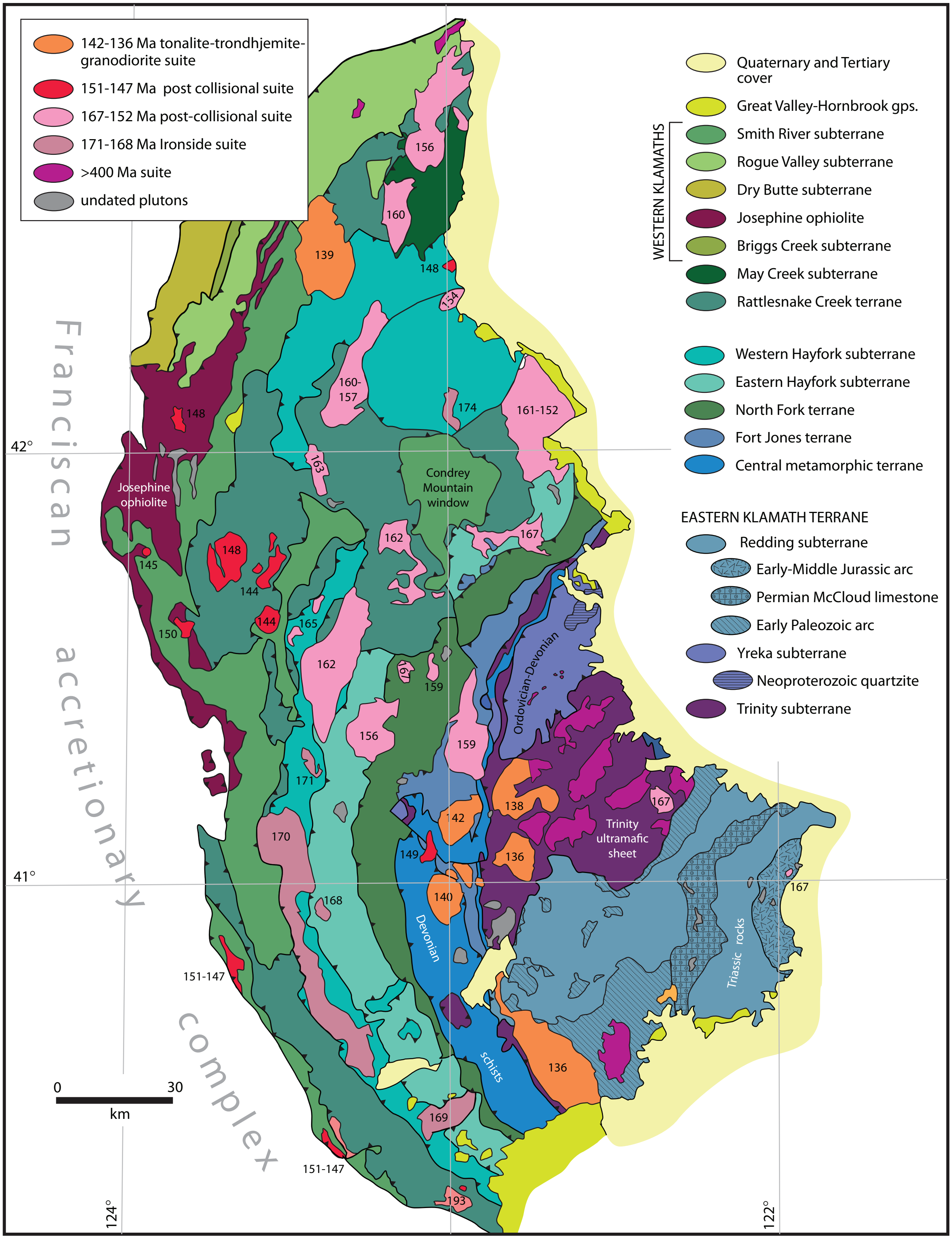


Figure 19. Geological sketch map showing terranes, subterranes, and various plutonic suites of the Klamath Mountains area. Modified after Irwin (2003); Snoke and Barnes (2006); and Allen and Barnes (2006).

of the Klamath block. Trinity subterrane comprises serpentinized peridotite cut by Neoproterozoic and Paleozoic plagiogranite and gabbro, overlain by mafic volcanic rocks; the Yreka subterrane consists of an early Paleozoic imbricate stack of metasedimentary nappes, amphibolite, ultramafic rocks, mélange, and the Neoproterozoic Antelope Mountain quartzite; and the Redding subterrane contains an early Paleozoic arc terrane, Mississippian metasedimentary rocks, a Permian arc terrane, which contains the type McCloud fauna, and a Triassic–Jurassic arc sequence (Potter et al., 1977; Hotz, 1977; Boudier et al., 1989; Peacock and Norris, 1989; Irwin, 2003). Rocks of the terrane roughly correlate with the Shoo Fly, Taylorsville, and Feather River belts of the Northern Sierra terrane (Irwin and Wooden, 2001; Irwin, 2003).

Rocks of the Central Metamorphic terrane (Fig. 19), locally separated from the Eastern terrane by rocks of the Fort Jones subterrane, an accretionary prism containing 220 Ma blueschists, are dominantly amphibolite, structurally overlain by schist and marble (Hacker and Peacock, 1990). The boundary between the Central and Trinity is a deep-seated east-dipping fault, known as the Trinity thrust, which is interpreted to extend eastward beneath Trinity subterrane for 100 km or more (Zucca et al., 1986). The Paleozoic to Early Jurassic North Fork subterrane (Fig. 19) consists of metamorphosed ophiolite, mafic volcanic rocks, radiolarian chert, and limestone, which Irwin (2003) correlated with the Sailor Canyon, Mount Jura, and Kettle Rock sequences in the Northern Sierra.

Lying to the west of the North Fork terrane is the composite Hayfork terrane (Fig. 19), which comprises (1) an eastern sector of Permo-Triassic broken formation and mélange of volcanic and sedimentary rocks, including chert and blocks of amphibolite, limestone, schist, and serpentinized ultramafic rocks, and (2) a western Middle Jurassic basaltic-andesitic volcanic arc that was accreted to the Eastern Hayfork terrane (Wright and Fahan, 1988; Irwin, 2010). Volcanic and sedimentary rocks of the western arc terrane, dated to be 177–168 Ma, are intruded by a number of high K_2O intrusive complexes of olivine-clinopyroxene ultramafic rocks, two-pyroxene gabbros, diorites and monzodiorites, and intermediate to siliceous hornblende-bearing rocks that cluster in age near 170 Ma, suggesting that the arc was accreted to the more easterly terranes between 169 and 164 Ma (Wright and Fahan, 1988; Barnes et al., 2006a). Prior to collision the Hayfork arc (1) sat atop the Rattlesnake Creek terrane, a Triassic subduction complex; (2) graded upwards from basal oceanic basalt to arc basalts; and (3) contained small gabbroic to quartz dioritic plutons ranging in age from 207 to 193 Ma (Wright and Wyld, 1994).

To the west, another arc complex, known as the Rogue-Chetco arc (Garcia, 1979, 1982), is separated from the Hayfork-Rattlesnake complex by the Josephine ophiolitic slab and its basinal cover sequence, the Galice flysch. The arc rocks were transported over Galice flysch of the Smith River subterrane on the Orleans thrust after 153 Ma, but before 150 Ma as the Galice rocks contain detrital zircons of that age and a 150 Ma pluton cuts the thrust fault (Miller et al., 2003). The Galice flysch sits depositionally atop the 164–162 Ma Josephine ophiolite, which in turn lies structurally above the Rogue Valley subterrane, a coeval 157 Ma arc, parts of which are located to the west (Harper et al., 1994).

Several important suites of Jurassic–Cretaceous plutons (Fig. 19) occur within the Klamaths: (1) a compositionally diverse suite of 167–152 Ma ultramafic, gabbroic, quartz dioritic, tonalitic, trondhjemitic, and granodioritic bodies known as the Wooley Creek suite; (2) a western suite of 151–147 Ma ultramafic to granodioritic plutons emplaced into the area after and during the waning stages of 155–150 Ma deformation; and (3) a suite of ~140 Ma tonalitic, trondhjemitic, granodioritic bodies that commonly show evidence for magma mixing and older crustal input (Allen and Barnes, 2006; Barnes et al., 2006b).

Jurassic Magmatic Rocks

A diverse belt of dominantly Jurassic volcanic, sedimentary, and plutonic rocks—generally interpreted to represent a low-standing, westward-facing arc located near the hypothesized and unexposed paleowestern edge of the North American craton (Hamilton, 1969a; Burchfiel and Davis, 1972, 1975; Busby-Spera, 1988; Busby-Spera et al., 1990; Fisher, 1990)—forms a linear belt from northern California southward through the Sierra to the Mojave Desert of California (Fig. 5). A similar band trends ESE across the Sonoran Desert through southern Arizona–northern Mexico and is generally considered correlative (Tosdal et al., 1989; Haxel et al., 2005). Yet another band trends NE through northwestern Nevada to the Snake River Plain (Crafford, 2007, 2008). Abrupt changes in strike and geology occur between both the Nevadan and Mojave-Sonoran sectors and the Sierran sector, suggesting that major faults lie between them (Figs. 5 and 7). These complex and varied rocks are described by area.

Sierran Region

In the northern Sierra (Fig. 17) the Jurassic arc rocks are exposed in thrust sheets around Mount Jura and are represented by the uppermost Triassic to Jurassic Tuttle Creek–Mount Jura sequence and Sailor Canyon Formation, which are composed of thick, steeply-dipping to overturned sections of terrigenous sedimentary rocks, largely derived from volcanic sources, and a spectrum of andesitic-dacitic-rhyolitic lava flows, proximal breccias, and tuff that reach an aggregate thickness of 8–11 km (Christe and Hannah, 1990; Harwood, 1992, 1993; Stewart et al., 1997; Lewis and Girty, 2001; Templeton and Hanson, 2003). The Kettle Rock succession contains high-K volcanic rocks dated at 180 Ma, cut by intrusions with related porphyry Cu-Au deposits (178 Ma) and unconformably overlain by rocks of the Mount Jura sequence dated at 161–148 Ma (Christe, 1993, 2011; Dilles and Stephens, 2011). The Emigrant Gap composite pluton and the Haypress Creek pluton (Fig. 17) cut rocks of the previously deformed Sailor Canyon and Tuttle Creek formations, as well as the Shoo Fly complex, and various phases of those complexes range in age from 168 to 163 Ma (Girty et al., 1993a, 1995;

John et al., 1994; Hanson et al., 1996, 2000). However, thin tuff units high in the section—and locally overturned beneath the east-vergent Taylorsville thrust—were recently dated to be 128 ± 3 Ma (Christe, 2010, 2011).

Metamorphosed basaltic-andesitic breccias, conglomerates, and hypabyssal intrusions, probable correlatives of the Tuttle Creek Formation, also occur to the southeast in the Verdi Range, just north of Lake Tahoe (Pauly and Brooks, 2002). Southwest of Lake Tahoe, similar age rocks are exposed in the Mount Tallac roof pendant (Saucedo, 2005) and comprise volcaniclastic sandstone, conglomerate, and monolithologic volcanic breccia, intruded by Pyramid Peak granite at 164 ± 7 Ma and smaller intermediate plutons (Sabine, 1992; Fisher, 1990). Southeast of Lake Tahoe, in the Walker River drainage, Schweickert (1976) described several Jurassic epizonal plutons and related volcanic and epiclastic rocks, all of which were folded.

In the Pine Nut Range, just over the border in Nevada and southwest of the Pine Nut fault (Fig. 8), lies ~2 km of mainly volcaniclastic rocks with minor andesite flows and intercalated carbonates of Triassic age overlain by a succession, ranging in age from 170 to 162 Ma, and comprising more than 3 km thick of siltstone, volcanic conglomerate, andesitic lavas, and ash-flow tuffs, cut by undated, but probable Jurassic and Cretaceous, plutons (Wyld and Wright, 1993).

Near Yerington, Nevada, the 168.5 Ma Yerington and the 166 Ma Shamrock batholiths intrude a thick section of Triassic andesitic and rhyolitic lavas, cut by 233 Ma intrusions, and overlain by 1800 m of Late Triassic–Early Jurassic dominantly nonvolcanic sandstones and limestones, which are in turn, overlain by a short-lived 169–165 Ma burst of magmatism that led to eruption of andesites, dacites, and basaltic lava flows associated with minor sedimentary and pyroclastic rocks (Dilles and Wright, 1988; John et al., 1994; Proffett and Dilles, 1984, 2008). The Yerington batholith hosts rich porphyry copper mineralization (Dilles, 1987).

Within the Sierra Nevada (Fig. 17), the rocks of the many pendants were divided into several sequences: (1) dominantly westward-dipping, metamorphosed, and polydeformed Paleozoic rocks of the Mount Morrison block located in the extreme east around Mono Lake and southward; (2) Triassic–Jurassic metavolcanic and metasedimentary rocks of the Koip sequence, which unconformably overlie Paleozoic rocks and in a general way young westward; (3) Cretaceous metavolcanic and sedimentary rocks that unconformably overlie or are in fault contact with rocks of the Koip and Kings sequences; (4) dominantly eastward-facing volcanic and sedimentary rocks of the Kings sequence, which occur in the western part of the batholith in the Kings-Kaweah and Yokohl Valley pendants; (5) the Kern Plateau pendants, which contain metasedimentary rocks of unknown provenance but are considered by some to be correlative with rocks of the Roberts Mountain allochthon (Dunne and Suczek, 1991; Chapman et al., 2012); and (6) the strongly-deformed, dominantly metasedimentary Isabella pendants (Bateman, 1992; Saleeby et al., 1978; Saleeby et al., 1990; Saleeby and Busby-Spera, 1986, 1993; Stevens and Greene, 2000). The general strike of rocks within the pendants is slightly more easterly than the Sierra Nevada.

One pendant (Fig. 17) that appears to be different from others is the Snow Lake pendant (Wahrhaftig, 2000), which comprises 148 Ma gabbroic dikes correlated with the 150–148 Ma Independence swarm and late Precambrian–Cambrian metasedimentary rocks with similar characteristics, such as detrital zircon profile, and presence of Scolithus, to those of rocks well to the south in the Death Valley–Mojave area (Smith, 1962; Stewart, 1970), from where they are hypothesized to have been transported (Lahren et al., 1990; Lahren and Schweickert, 1989, 1994; Schweickert and Lahren, 1990; Grasse et al., 2001a; Memeti et al., 2010a). These workers suggested that rocks to the south in the Kings sequence might be part of the same sequence, and so separated from rocks to the east by the intrabatholithic Snow Lake–Mojave fault; however, the defining characteristics don't appear to be present in those pendants (Saleeby et al., 1978), so the Snow Lake pendant might represent a fault wedge transported northward from the Mojave, and there might be two intrabatholithic faults or splays of one fault system (Kistler, 1993).

Just to the east in the Saddlebag Lake pendant (Fig. 17), and nearby in several smaller pendants, is a several-km-thick accumulation of westerly-dipping conglomerate, rhyolitic ash-flow tuff, and andesitic lava, breccia, and associated epiclastic rocks that sits unconformably upon Paleozoic and Lower Triassic rocks (Schweickert and Lahren, 1993a). At least one caldera complex—with a rhyolitic outflow-facies sheet dated by U-Pb on zircons at 222 ± 5 Ma and a possible associated pluton with a U-Pb zircon age of ~232–219 Ma—occurs within the Saddlebag Lake pendant and was folded and transported along eastward-vergent thrust faults prior to the emplacement of the 168 Ma granodiorite of Mono Dome (Schweickert and Lahren, 1993a, 1993b, 1999; Barth et al., 2011).

Along strike with the Saddlebag Lake pendant to the southeast lies the Ritter Range pendant (Fig. 17), which comprises a westerly-dipping sequence containing polydeformed Paleozoic sedimentary rocks and a probable thrust stack of Late Triassic to Early Jurassic (214–186 Ma) metavolcanic and metasedimentary rocks unconformably overlain by a 164 Ma sequence of meta-ignimbrite, lavas and breccias, small fault-bounded slivers of 140–130 Ma volcanic rocks, and—separated from the other rocks by a fault—a westerly-dipping Middle Cretaceous (~100 Ma) caldera-fill sequence, intruded at 98 Ma by what is interpreted as a resurgent pluton (Rinehart and Ross, 1964; Huber and Rinehart, 1965; Russell and Nokleberg, 1977; Fiske and Tobisch, 1978, 1994; Greene et al., 1997; Tobisch et al., 1986, 2000). A suite of 226–218 Ma plutonic rocks, known as the Scheelite intrusive suite, intrudes the older volcanic rocks (Bateman, 1965a, 1992; Barth et al., 2011).

To the southeast, the narrow Mount Goddard pendant (Fig. 17) contains metamorphosed and deformed, southwest-dipping and facing volcaniclastic rocks intercalated and interleaved with sparse tuff, tuff-breccia, ash-flow tuffs (143 ± 3 Ma and 131 ± 6 Ma), lava flows (156 ± 2 Ma and 140 ± 1 Ma), intrusions (159–156 Ma), some of which are older than their wall rocks, and minor carbonate beds (Bateman, 1965b; Bateman and Moore, 1965; Lockwood and Lydon, 1975; Moore, 1978; Tobisch et al., 1986). Sequences (and older plutons) there were regionally metamorphosed to amphibolite grade during penetrative deformation, display both older over younger and younger over older structural relations, with deformation bracketed to be between 131 ± 6 Ma, the age of the youngest dated tuff, and 90 ± 3.5 Ma, the age of the foliated Mount Givens granodiorite, which intrudes the Goddard pendant on the southwest (Tobisch et al., 1986; Bateman, 1992).

Connected along strike to the southeast with the Mount Goddard pendant by a series of sill-like Jurassic plutons, are southwesterly-dipping rocks of the Oak Creek pendant (Fig. 17), which comprises a 165 Ma sequence of metamorphosed and deformed lava flows, rhyolitic and dacitic ash-flow tuffs, and temporally associated plutons, all overlain by intermediate-mafic composition lavas, tuffs, and volcaniclastics of ~109 ± 2 Ma (Moore, 1963; Saleeby et al., 1990). Adjacent to the Oak Creek pendant a mylonitic orthogneiss was dated at 164 ± 4 Ma and deformed Jurassic plutons were dated as 165–164 Ma (Mahan et al., 2003). These older deformed plutons are cut by the Independence dike swarm, and this is the type area where they were first described (Moore and Hopson, 1961; Moore, 1963). Westward into the Giant Forest area (Fig. 17) of Sequoia National Park, several pendants contain Jurassic supracrustal rocks and plutons, some as old as 162 Ma, and many of those pendants contain dikes that might be part of the Independence dike swarm (Sisson and Moore, 1994; Moore and Sisson, 1987). In the area around Triple Divide Peak itself (Moore, 1981; Moore and Sisson, 1987), the dikes might be folded along with their wall rocks as the folds appear slightly more open than typical, and where the limbs strike NW, the dikes strike NW, and where the limbs strike SW, the dikes strike SW. These relations suggest that, at least locally, the dikes might originally have been sills.

About 30–40 km to the west of the Oak Creek pendant is the Boyden Cave pendant (Fig. 17), where an ~5-km-thick section of steeply-dipping Paleozoic metasedimentary rocks, including quartzite, marble, pelite, and sandstone, is separated from a Middle Cretaceous sequence, consisting of steeply dipping 105–100 Ma rhyolitic ash-flow tuff and mixed siliceous–intermediate volcanic and volcaniclastic rocks, by younger intrusions (Saleeby et al., 1990). Southwest of the Boyden Cave pendant in the Giant Forest area there are several small pendants that contain Jurassic–Cretaceous rocks. The granodiorite of Yucca Mountain, dated at 162 Ma, intrudes an undated metasedimentary succession containing metabasalt and andesitic lavas, whereas a few scattered bits of Cretaceous ash-flow tuff sit unconformably upon the older rocks (Sisson and Moore, 1994).

To the south, most of the rocks within the pendants are strongly metamorphosed and tectonized (Wood and Saleeby, 1998), which makes protolith recognition difficult, but small areas of known Cretaceous rocks outcrop within the Isabella Lake pendants (Fig. 17), where they unconformably overlie previously deformed metasedimentary rocks of the King's sequence and are known as the Erskin Lake sequence (Saleeby and Busby-Spera, 1986). Included within the sequence are steep eastwardly-dipping, ignimbritic units, volcanic breccias, and a possible volcanic neck that together yielded a narrow spectrum of U-Pb zircon ages ranging from ~107 to 102 Ma (Saleeby et al., 2008).

Jurassic rocks crop out in the Alabama Hills just west of Lone Pine (Stone et al., 2000). There (Fig. 17), two folded, but dominantly southwest-dipping, sequences are exposed: a lower sequence comprising ~2 km of rhyolitic ash-flow tuff and volcaniclastic rocks, all intimately intruded by dikes and sills; and an unconformably overlying upper sequence consisting of deformed and altered sedimentary rocks along with a 170 ± 4 Ma rhyolitic ash-flow tuff more than 450 m thick (Dunne and Walker, 1993).

There are several areas within the White-Inyo Mountains (Figs. 5 and 17) where Jurassic magmatic rocks occur. In the north, the oldest are siliceous volcanic rocks interbedded with marble, exposed as pendants within the 165 Ma Barcroft pluton, and the lower parts of a 3-km-thick assemblage of metamorphosed basaltic andesite to rhyolitic lavas, tuffs and hypabyssal intrusions cut by the pluton and overlain by another sequence of metavolcanic rocks that contains a rhyolitic tuff dated to be ~154 Ma (Hanson et al., 1987). They also reported that a hypabyssal intrusion that cuts the sequence is 137 Ma. To the northeast, within Esmeralda County, Nevada, several plutons of likely Jurassic age, such as the Sylvania and Palmetto, crop out, but similar age volcanic rocks are unknown (Albers and Stewart, 1972).

In the southern Inyo Mountains there are three separate unconformity-bound sequences of dominantly southwesterly-dipping and facing metasedimentary and metavolcanic rocks that sit unconformably upon Lower Triassic rocks: (1) a 500 m sequence of conglomerate, breccia, sandstone and small quantities of basaltic lavas; (2) a 169–168 Ma sequence, up to 800 m thick, dominated by sheets of welded ash-flow tuff, andesitic and rhyolitic lavas with lesser amounts of volcanic cobbly conglomerate; and (3) 2.5–3 km of volcanogenic conglomerate, sandstone, and finer-grained sedimentary rocks intercalated with minor ash-flow tuff, lava flows, and debris flows, with a dacitic tuff dated at 148 Ma (Dunne and Walker, 1993; Dunne et al., 1978, 1998).

To the south, in the Argus, Slate, and nearby ranges (Fig. 17), are several areas of poorly dated Jurassic rocks, including andesitic to basaltic lavas intercalated with volcaniclastic rocks, siliceous ash-flow and air-fall tuffs, some cut by plutonic complexes such as the 186 Ma Bendire pluton, the 174 Ma Hunter Mountain batholith (Dunne, 1979), and the 161 Ma Maturango pluton (Dunne et al., 1978). Another sequence of volcanogenic conglomerate, sandstone, ash-flow tuff, and porphyritic andesite, known as the Warm Spring Formation, crops out in the Butte Valley area of the Panamint Range, and is cut by plutons with K-Ar ages in the range 145 to 137 Ma (Abbott, 1972).

In the El Paso Mountains (Fig. 17), which sit along the north side of the Garlock fault, a quartz dioritic–quartz monzodioritic pluton, with K-Ar ages on hornblende of 152 Ma and on biotite

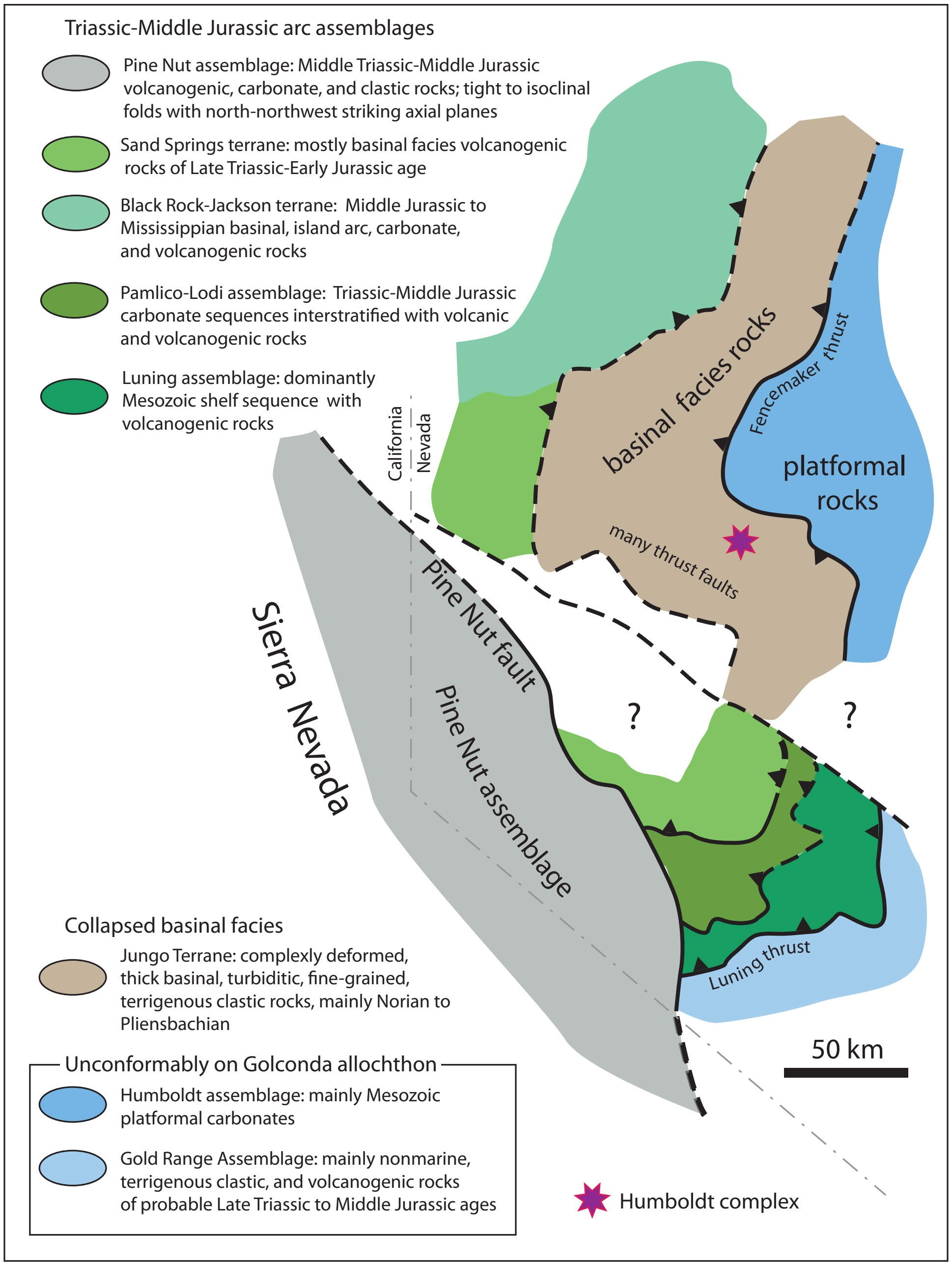


Figure 20. Sketch map illustrating the relations between arc, platform, and basinal terranes of western Nevada. Location of Humboldt complex is approximate as it occurs in several thrust slices over a broad area (Dilek and Moores, 1995). From Crafford (2007, 2008) and Oldow (1984).

of 146 Ma, intruded and metamorphosed previously metamorphosed and deformed rocks (Carr et al., 1997). They mapped and described two tectonically juxtaposed and distinct lower Paleozoic sequences: a western finer-grained meta-argillitic and argillic-metachert unit with greenstones, and an eastern platformal facies comprising meta-argillite, marble, orthoquartzite, and graptolitic slate, all unconformably overlain by Mississippian metaconglomerate and meta-argillite—interpreted as foredeep fill related to the Antler orogeny (Carr et al., 1980)—and intruded by Late Permian gneissic plutons and a nonfoliated Early Triassic plutonic suite ranging in composition from gabbro to granite.

Northwestern Nevada

In northwestern Nevada north of the Pine Nut fault (Figs. 8 and 20), there is a complex northeasterly-trending band of Triassic to Cretaceous sedimentary, metamorphic, and magmatic rocks (Crafford, 2007, 2008; Oldow, 1984; Oldow et al., 1989). There, movement along faults of the southeast-vergent Luning-Fencemaker fold-thrust belt—considered by some to be part of the same progressive eastwardly migrating back-arc shortening that created the Sevier fold-thrust belt farther east (Burchfiel et al., 1992; Saleeby and Busby-Spera, 1992; DeCelles and Coogan, 2006)—juxtaposed magmatic rocks of the Black Rock arc terrane and the Humboldt igneous complex above several km of highly deformed Triassic basinal facies rocks (Burke and Silberling, 1973; Speed. 1978), apparently without known basement, with the basinal facies rocks thrust over a little-deformed west-facing Triassic carbonate platform that was deposited atop the previously deformed Golconda and Roberts Mountains allochthons (Oldow, 1984; Elison and Speed, 1988, 1989; Quinn et al., 1997; Dilek et al., 1988; Dilek and Moores, 1995; Wyld, 2000, 2002). The basinal facies rocks are dominated by detrital zircons in the age range 1145 to 948 Ma (Manuszak et al., 2000), which suggests that the basinal facies rocks were not derived from North America as originally suggested by Elison and Speed (1988). The arc terrane was apparently in contact with the basinal facies rocks by the Early Jurassic, but the depositional relationship between the basinal facies rocks and the eastern shelf is unknown.

The Black Rock terrane comprises up to 10 km of upper Paleozoic sedimentary and volcanic strata overlain by Middle Triassic to Early Jurassic intermediate to mafic volcanogenic rocks and a variety of Jurassic intrusions, several in the 170 to 163 Ma range (Quinn et al., 1997; Wyld, 2000). The ~165 Ma Humboldt igneous complex comprises a cogenetic suite of ultramafic to granitic plutonic rocks unconformably overlain by pillow basalts, andesitic-dacitic lavas, and a variety of volcanic breccias and tuff (Dilek and Moores, 1995). Deformation within the Black Rock terrane may have begun in the earliest Jurassic, but the main pulse coincided with its emplacement above the collapsed basin just after 163 Ma, followed by 7–14 km of exhumation, which makes a connection between this deformation and sedimentation of Upper Jurassic age on cratonic North America highly unlikely as there is no deformation or orogenic sedimentation there of that age (Wyld et al., 2003; Wyld and Wright, 2009). Lower Cretaceous deposits of alluvial debris intercalated with minor ash-flow tuffs dated at 125–124 Ma sit unconformably upon the uplifted and eroded Jurassic rocks and are cut by a pluton dated as 123 ± 1 Ma (Martin et al., 2010). Upper Cretaceous plutons (93–88.5 Ma), largely similar in age and composition to those of the Sierran Crest magnatic event along the Sierran crest, intrude the older rocks (Smith et al., 1971; Van Buer and Miller, 2010).

Mojave Desert Region

Much of the geology of the Mojave Desert appears to be built upon a Paleozoic substrate sitting atop scattered outcrops of Precambrian crystalline basement, but as there are many Cenozoic strike-slip faults and over half of the area is covered by alluvium or Cenozoic volcanic rocks and unexposed, it is a challenging area to reconstruct Mesozoic and older paleogeography (Burchfiel and Davis, 1981).

Within the Mojave Desert, two different terranes: one characterized by deep water sedimentary rocks, Late Permian deformation, and 260 Ma plutonism/andesitic volcanism—all deformed by west-vergent folds and thrusts; and the other consisting of sedimentary rocks deposited in shallow water and with post-Mississippian deformation, were juxtaposed prior to the emplacement of 246–243 Ma postkinematic plutons (Martin and Walker, 1995; Miller et al., 1995; Carr et al., 1997; Walker et al., 2002). Sitting unconformably upon plutons dated at 243 and 241 Ma (Miller et al., 1995; Barth et al., 1997) in the area around Victorville (Fig. 21) is the Fairview Valley Formation, which is an isoclinally folded sequence of poorly dated, but possibly Jurassic, metasedimentary rocks (Walker, 1987; Schermer et al., 2002).

Along the north side of Antelope Valley in the Tehachapi Mountains (Fig. 21), but south of the Garlock fault, is an intermittent band, mostly surrounded by plutons presumed to be Cretaceous, of marble, schist, metavolcanic rocks, and ultramafic rocks, collectively known as the Bean Canyon Formation (Dibblee, 1967; Ross, 1989). The age of the unit was poorly constrained as no fossils were found, but based on thick marble and basalt units, Wood and Saleeby (1998) speculated that it might be correlative with the Late Triassic–Early Jurassic Kings sequence in the Isabella Lake pendant; however, a dacitic metatuff was recently dated by Chapman et al. (2012) and is clearly mid-Permian at 273 Ma.

An extensive area of Middle Jurassic volcanic rocks occurs in the central Mojave region, where they are known as the Sidewinder volcanic series (Bowen, 1954). The series, which might be temporally correlative with the Fairview Valley Formation, is divided into a lower section of 179–164 Ma rhyolitic-dacitic intracauldron-facies ash-flow tuffs unconformably overlain by an upper sequence of rhyolitic-basaltic lavas intruded by a rhyolitic dike dated to be 152 ± 6 Ma (Schermer and Busby, 1994; Schermer et al., 2002). They also dated, by U-Pb on zircon, a thick ignimbrite, separated from the lower sequence by a period of erosion, deposition, and possible faulting, at 151 ± 1 Ma, which indicates that this tuff is much younger than the lower sequence, and might therefore be more closely related to the upper sequence. Recently, Fohey-Breting et al. (2010) used an ion microprobe to date three of the major ash-flow units in the Sidewinder volcanics to confirm the earlier results. He attempted to relate the tuffs to specific intrusives in the area: the oldest at 180 ± 3 Ma had no recognized intrusive equivalent; a 161 Ma tuff is approximately the same age as the 167–161 Ma Bullion Mountains intrusive suite and related plutons, which are part of the Kitt Peak–Trigo Mountains super unit of Tosdal et al. (1989); and a 150 ± 2 Ma unit more or less contemporaneous with 155–151 Ma calc-alkaline plutons and a 149 Ma syenitic body.

Rocks of about the same age occur in the Rodman, Ord, and Fry mountains (Fig. 21), where thick sequences of metamorphosed and deformed ash-flow tuff, andesite flows, and a variety of epiclastic rocks are cut by Middle Jurassic plutons dated at 171–166 Ma and later by nondeformed latest Jurassic Independence dikes, some of which may have fed lava flows (Karish et al., 1987; James, 1989).

In several small ranges, such as the Cowhole Mountains, located just south of Baker (Fig. 21), 170 Ma dacitic-rhyodacitic ash-flow tuff, volcaniclastic rocks, and andesitic breccias are

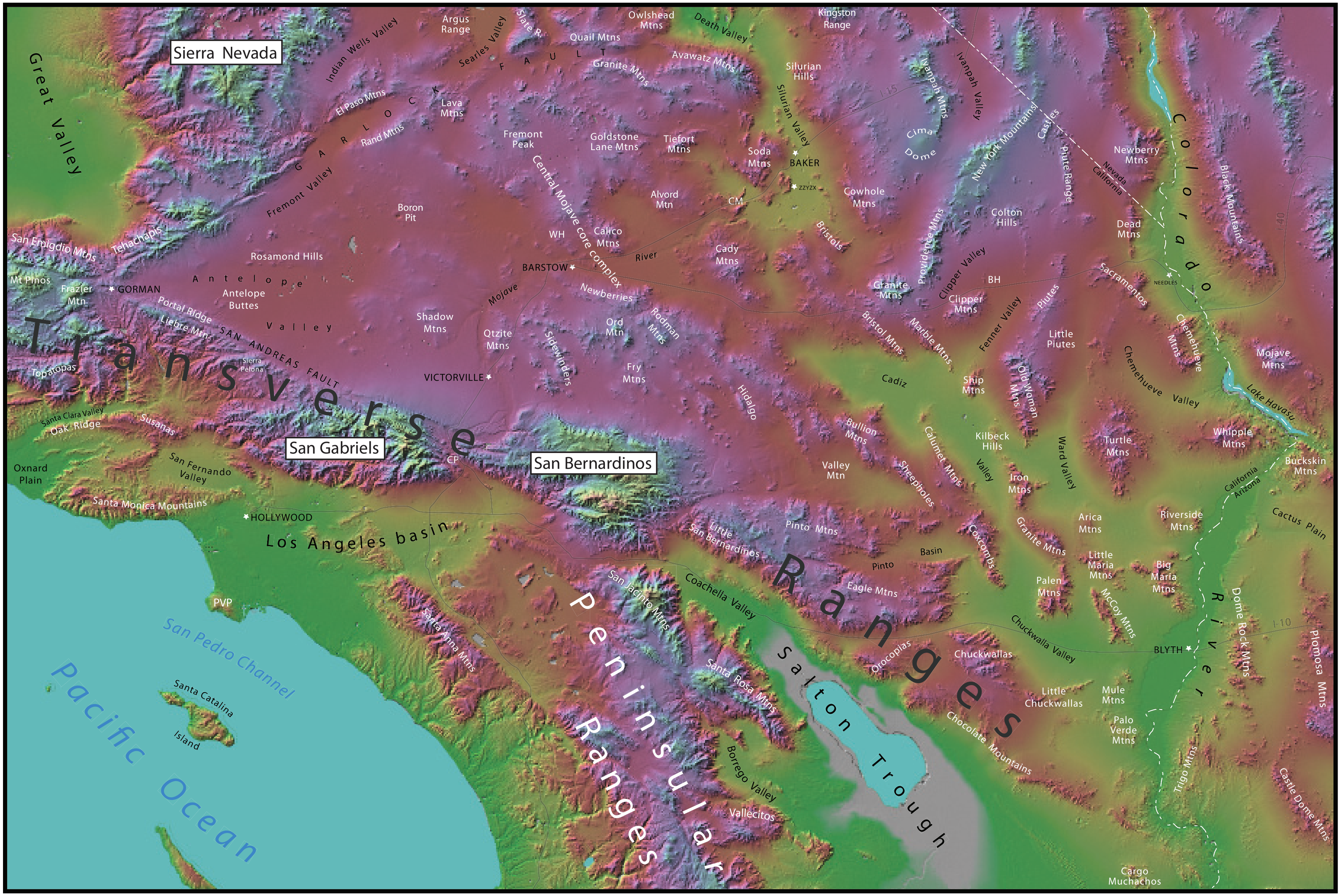


Figure 21. Color, shaded relief map showing locations of ranges in the Mojave Desert and southern California discussed in the text. The base image was made by Ray Sterner at the Johns Hopkins Applied Physics Laboratory using 1 arc second Shuttle Radar Topography Mission (SRTM) digital elevation data obtained by the SRTM, the 11-day STS-99 mission by the space shuttle *Endeavor* in February 2000.

intercalated with eolian sandstones in fault-bounded paleodepressions and overlain by the Cowhole volcanics, which comprise ~500 m of welded ash-flow tuff, various volcanogenic breccias, dacitic lava, and near the top, rhyolitic breccias intruded by a hypabyssal sill dated at 169 ± 2 Ma—all intruded by probable Independence dikes (Marzolf and Cole, 1987; Wadsworth et al., 1995; Busby et al., 2002).

A cluster of ranges, the Clipper, Ship, Piute, and Old Woman mountains, as well as the Kilbeck Hills (Fig. 21), within the east-central Mojave block lie close to the NE limit of Jurassic magmatism there and contain a wide variety of plutons and deformational features (Howard et al., 1997; Howard, 2002). Howard et al. (1995) argued that the 161 ± 10 Ma Goldhammer pluton, a dominantly monzodioritic pluton, was emplaced during thrusting of Proterozoic crystalline basement over Paleozoic strata. Postkinematic plutons include the 150–145 Ma Ship Mountains pluton, a mingled complex of granites, quartz monzonite, gabbro, diorite, and monzodiorite, and a suite of dikes, dated to be ~145 Ma (Gerber et al., 1995).

In the Palen Mountains (Fig. 21), 4 km of dacitic to rhyolitic tuff, lava flows, hypabyssal intrusions, a dome complex, and various epiclastic rocks of the 174 ± 8–162 ± 3 Ma Dome Rock sequence unconformably overlie conglomerates and sandstones of the Palen Formation and are themselves unconformably overlain by rocks of the dominantly Cretaceous McCoy Mountains Formation (Fackler-Adams et al., 1997; Busby et al., 2002). To the east, and exposed within the lower plate of the Whipple Mountains core complex is a suite of deformed granodioritic-quartz dioritic plutons, known as the Whipple Wash suite, dated to be 89 ± 3 Ma (Anderson and Cullers, 1990).

Just to the southeast over the border in Arizona, the Dome Rock Mountains (Fig. 21) contain sections of Paleozoic metasedimentary rocks and Mesozoic volcanic rocks intruded by Mesozoic plutons. A rhyolitic tuff, dated at 165 ± 3 Ma, and a granodiorite, dated at 164 Ma, predate recumbent southwest-vergent folding, whereas a 161 Ma leucogranite postdates the deformation (Boettcher et al., 2002).

A 100 × 20 km band of strongly metamorphosed rocks deformed during the Jurassic and Cretaceous is exposed in the footwall of the Miocene Central Mojave metamorphic core complex (Fig. 21) (Fletcher et al., 1995). Porphyritic metavolcanic rocks in the footwall, known as the Hodge volcanic sequence, yielded U-Pb zircon ages of ~170–164 Ma, whereas a postkinematic granite gave an age of 151 ± 11 Ma, and a muscovite-garnet granite that crosscuts Cretaceous deformational fabrics provided an age of 83 ± 1 Ma (Boettcher and Walker, 1993). In the Shadow Mountains (Fig. 21), Martin et al. (2002) reported on a Neoproterozoic–Mesozoic sequence of rocks, similar to sequences in the Death Valley region that were deformed and metamorphosed during recumbent folding prior to intrusion of gabbro and diorite at 148 ± 1.5 Ma and younger granite at 144–143 Ma.

Southeast of Barstow, an extensive 167 Ma intrusive complex, known as the Bullion Mountains intrusive suite, is dominated by granite, quartz monzonite, and quartz monzodiorite; outcrops in the Bullion, Pinto, and Eagle mountains (Fig. 21); and may be related to the coeval Dale Lake volcanics, which are mainly intermediate composition lavas, epiclastic rocks, and tuffs containing distinctive oval, lavender alkali feldspars similar to those in the quartz monzonitic phase of the intrusive complex (Mayo et al., 1998; Howard, 2002).

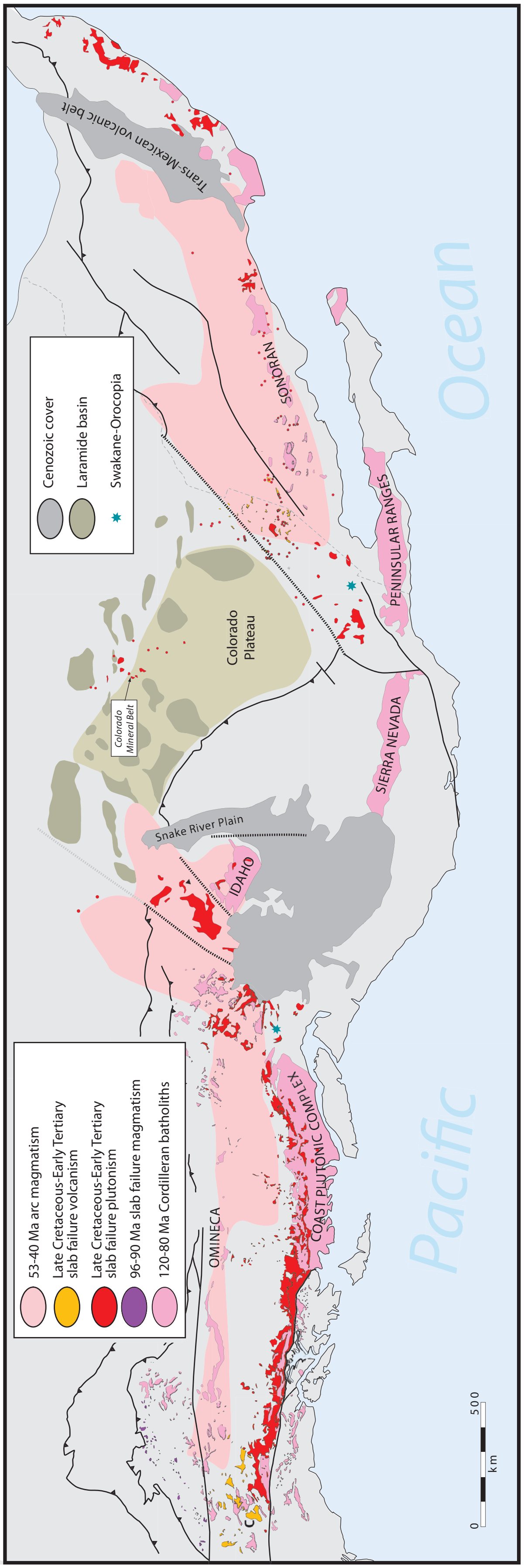
Northeast of Barstow, in the Tiefort Mountains (Fig. 21), Schermer et al. (2001) described metasedimentary rocks of unknown age, and metavolcanic rocks, inferred to be Mesozoic, intruded by foliated plutons dated by U-Pb at 164–160 Ma, northerly trending siliceous and mafic dikes (148 ± 14 Ma) that crosscut the foliation in the older plutons, and 82 Ma pegmatites, and orthogneiss from South Tiefort Mountain with a U-Pb age of 105 Ma, indicating both Jurassic and Cretaceous deformation.

To the east in the Cronese Hills (Fig. 21), locally highly sheared and thrust greenschist grade metavolcanic rocks derived from tuffs and lavas, metaplutonic rocks (166 ± 3 Ma), and metasedimentary rocks are overturned and cut by a postkinematic granite dated at 155 ± 1 Ma (Walker et al., 1990b). Just to the west at Alvord Mountain (Fig. 21), Miller and Walker (2002) described a foliated monzodioritic-quartz monzonitic pluton, dated at 179 Ma by Miller et al. (1995), that is cut by nonfoliated gabbroic and hornblende diabase dikes, dated at 149 ± 3 Ma, and intermediate composition porphyritic dikes dated at 83 Ma.

Cretaceous Batholithic Rocks

Several Cordilleran-type batholiths occur mostly as parts of long-lived, composite magmatic belts in western North America (Figs. 5 and 22). Hildebrand (2009) broke out distinct periods of magmatism (Fig. 23), separated by lulls, within several of the volcano-plutonic belts: (1) an Upper Jurassic–Early Cretaceous extensional arc phase, with well-preserved volcanic rocks intercalated with fluvial and shallow marine terrigenous clastics; (2) a 120 Ma to ~80 Ma stage (Fig. 24) when the main masses of the Cordilleran batholiths were formed; (3) a 75–60 Ma slab break-off phase of magmatism, in places only tens of km wide, yet thousands long, consisting of intermediate to siliceous volcanic and plutonic rocks; and (4) a magmatic arc stage containing typical arc rocks that were erupted starting at ~53 Ma, were generated by eastward subduction, and continue to form today in the Pacific Northwest. Short, but significant, magmatic gaps of 5–10 Myr generally occur between phases. While all of the batholiths share many features in common, the most impressive commonality is that the most voluminous phase, the Cordilleran batholithic phase, occurred during the same time period: ~120–80 Ma (Fig. 24).

The Sierra Nevada batholith is probably the best studied of the Mesozoic batholiths, although the Peninsular Ranges batholith and Coast plutonic complex are also reasonably well known. The main bulk of the batholith is Cretaceous, but older Mesozoic rocks have been widely considered to be genetically related and are commonly included within it (Bateman and Wahrhaftig, 1966;



Bateman et al., 1963; Bateman, 1992; Saleeby et al., 2008). In this volume, I separate the Cretaceous magmatism from older and younger suites, not only because the older rocks shed light on arc polarity and timing of subduction initiation that generated younger rocks, but also because there are significant magmatic gaps, at times coupled with periods of deformation, that I view to be important markers separating different tectonic regimes. Because the rocks of the Cretaceous batholith trend slightly more northerly than older rocks, plutons not strictly part of the batholith as defined here, and mainly older, occur northwest of the main batholithic mass in the Foothills metamorphic belt and to the southeast, mainly in the White-Inyo Mountains: only a few relict masses occuring within the main batholithic mass were discussed in a previous section.

I also subdivide the 120–80 Ma Cordilleran batholithic phase and examine its origins in more detail, because this period of magmatism in the Coast plutonic complex, the Peninsular Ranges batholith, and the Sierra Nevada batholith appears to comprise two phases, or parts (Gromet and Silver, 1987; Silver and Chappell, 1988; Kistler, 1990, 1993; Bateman et al., 1991; Todd et al., 2003; Lee et al., 2007; Lackey et al., 2008), that were deformed and joined by collision between about 105 and 100 Ma (Kimbrough et al., 2001; Tulloch and Kimbrough, 2003; Saleeby et al., 2008; Gehrels et al., 2009). In a general sense, the older pre-deformational magmatic rocks occur to the west, whereas to the east younger post-collisional magmatism dominates.

Sierra Nevada Batholith

The plutonic rocks within the Sierran batholith (Figs. 5, 8, and 17) range in composition from gabbro to leucogranite, but the most common rock types are tonalite, granodiorite, and granite (Bateman and Wahrhaftig, 1966; Bateman et al., 1963; Bateman, 1992; Ross, 1989). In general, the hundreds of plutons within the batholith have sharp contacts with each other or are separated by minor screens of older metamorphic rock (Bateman, 1992).

Crystalline basement beneath the batholith is unknown, but based on the composition and size of the plutonic bodies it must be continental (Hildebrand and Bowring, 1984). Ever since Moore (1959) recognized that the more mafic plutons lay west of more intermediate composition bodies, others have confirmed that the Sierra can be divided into older western and younger eastern parts based on geochemistry, magnetic susceptibility, age, radiometric and stable isotopes, wall rock provenance, and basement types (Chen and Tilton, 1991; Bateman et al., 1991; Kistler, 1990, 1993; Saleeby et al., 2008; Lackey et al., 2008, 2012a, 2012b; Chapman et al., 2012). All known wall rocks within the Sierran batholith are older than 98 Ma and they are all deformed

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Figure 22. Sketch map showing the distribution of Cordilleran-type batholiths, postcollisional slab-failure magmatism, and younger 53–40 Ma arc magmatism. A larger image of Figure 22 is on the loose insert accompanying this volume.

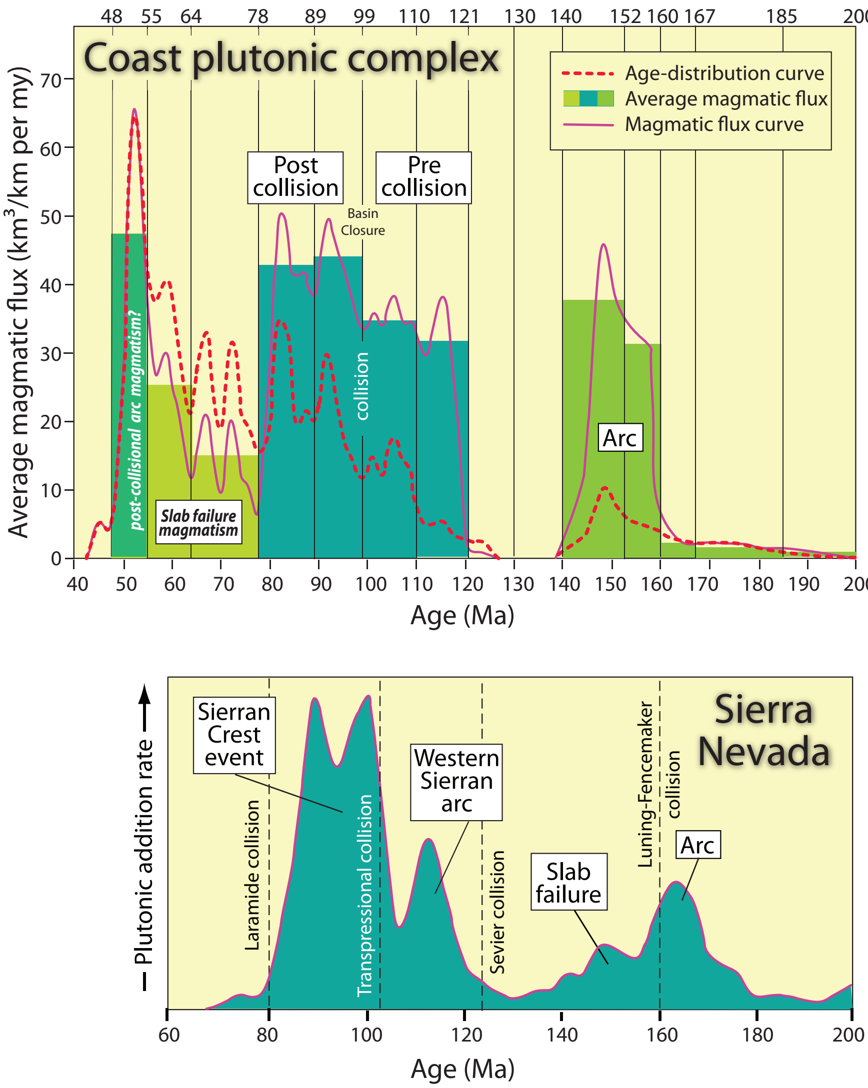


Figure 23. Average magmatic flux (based on plutonic rocks) plotted against age for the Coast plutonic complex of coastal British Columbia compared with average plutonic addition rate for the Sierra Nevada batholith of California. The plot for the Coast plutonic complex illustrates the four interpreted phases of magmatism from Hildebrand (2009). Note the overall similarities in flux between the two regions. The main Cordilleran batholithic phase (120–80 Ma) is bounded temporally by the Sevier and Laramide collisions. Most workers consider the Cordilleran-type magmatism to be arc magmatism from a single subduction zone, but in this contribution the magmatism is divided into two parts that likely have very different origins. An important difference between the Coast and Sierran regions is that the Sierran batholith is devoid of post-Laramide magmatism. Coast plutonic complex modified from Gehrels et al. (2009) and Sierra Nevada batholith modified from Paterson et al. (2012).

(Peck, 1980; Nokleberg and Kistler, 1980; Bateman et al., 1983a; Saleeby and Busby-Spera, 1986; Memeti et al., 2010a; Wood, 1997; Saleeby et al., 1990). The bulk of this deformation apparently occurred prior to the emplacement of 98–85 Ma plutons of the Sierra Crest magmatic event (Coleman and Glazner, 1998; Davis et al., 2012) and, based on metamorphic studies, prior to 95 Ma in the southernmost Sierran batholith (Saleeby et al., 2007, 2008).

Over most of the length of the batholith plutonic rocks were emplaced at mesozonal-epizonal levels as indicated by their narrow to moderate metamorphic aureoles of hornblende hornfels (Bateman, 1992). Detailed geobarometry indicates that the main mass of the visible batholith crystallized at 3–4 kb, except at its southern end where P>6 kb and most rocks are at upper amphibolite or granulite grade; and along the eastern side where pressures were dominantly 1–2.5 kb (Ague and Brimhall, 1988; Ross, 1989; Wood and Saleeby, 1998; Saleeby et al., 2007; Nadin and Saleeby, 2008).

While there are small areas of Triassic and Jurassic bodies within the Sierra, the main mass of plutons ranges in age from 125 to 82 Ma (Stern et al., 1981; Chen and Moore, 1982; Bateman, 1992; Irwin, 2003). In the east-central part of the batholith, Bateman (1992) broke out several intrusive suites of cogenetic, but not necessarily comagmatic, plutons that have distinctive petrographic, compositional, and textural characteristics as well as spatial proximity. The best known are the <100 Ma compositional zoned complexes such as the Tuolumne intrusive suite, which comprise apparently nested units that are progressively younger and more leucocratic inward (Calkins, 1930; Bateman and Chappell, 1979; Bateman et al., 1983b; Huber et al., 1989). The "nested" intrusive complexes (Fig. 17) were emplaced along the eastern Sierran crest between ~98 Ma and 86 Ma, and are characterized by an outer, older tonalite and granodiorite in sharp contact inward with younger hornblende porphyritic granodiorite, and cored by even younger K-feldspar megacrystic granite and granodiorite (Bateman, 1992; Coleman

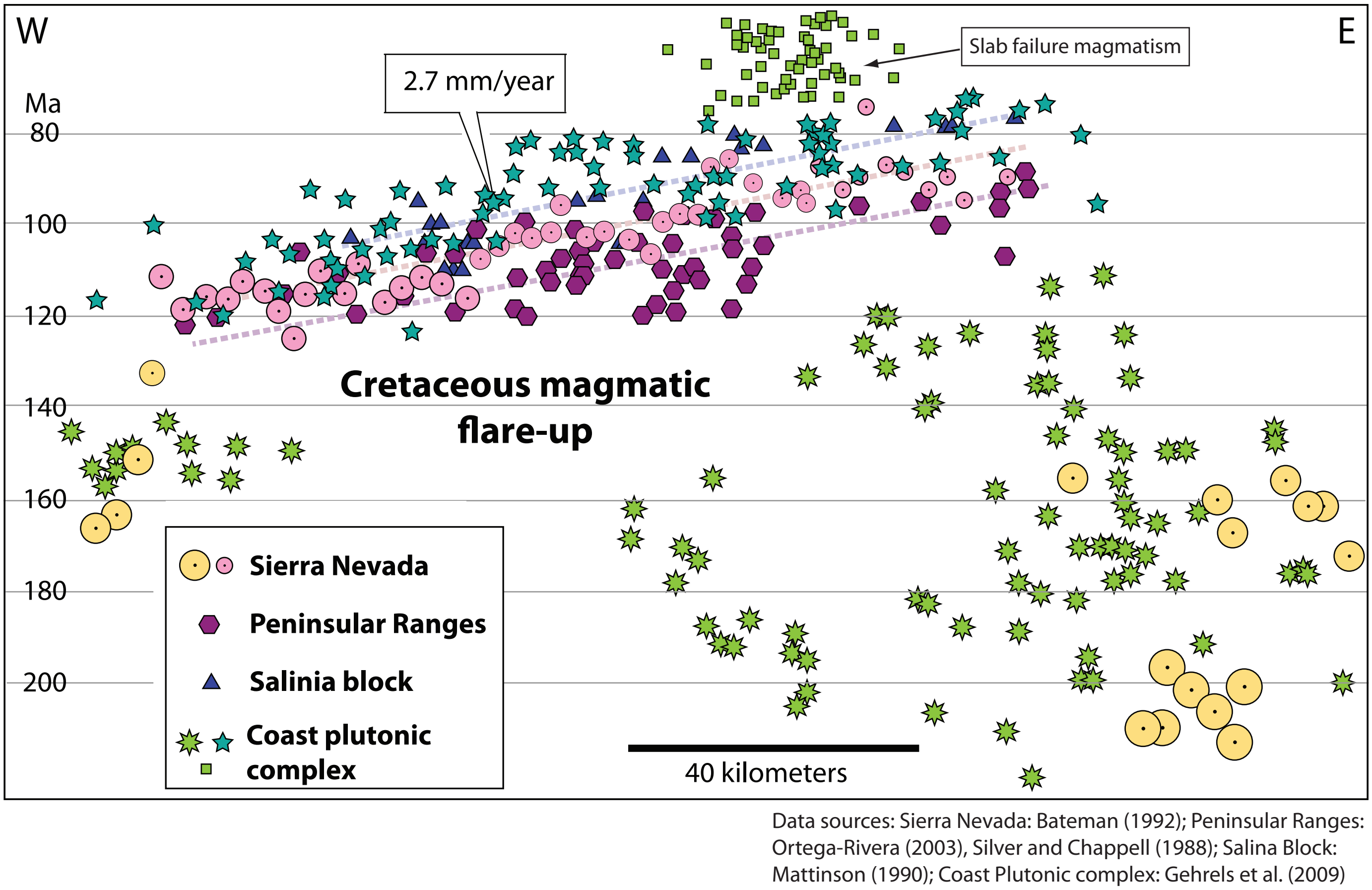


Figure 24. Age versus distance plots for four flare-up Cordilleran-type batholiths showing dominant 120–80 Ma ages. Note tightly focused <80 Ma magmatism of the Coast plutonic complex interpreted to be slab-failure magmatism (Hildebrand, 2009). Also well displayed are the two pre–105 Ma magmatic suites of the Coast plutonic complex joined during the ~100 Ma transpressional deformation characteristic of most of the Cordilleran-type batholiths.

and Glazner, 1998; Hirt, 2007). Another similar nested complex, the Sahwave, occurs in northwestern Nevada, where the regional trend is NNE (Van Buer and Miller, 2010). Based on the narrow grouping of U-Pb ages, Coleman and Glazner (1998) considered the Tuolomne intrusive suite, along with other similar complexes both north and south along the Sierran crest, such as the Whitney and Mono Pass intrusive suites (Gaschnig et al., 2006; Hirt, 2007), to have formed during one 10 Myr magmatic burst, which they named the Sierran Crest magmatic event.

To the south, Nadin and Saleeby (2008) divided the Sierra Nevada into three longitudinal zones: (1) a western zone rich in mafic and tonalitic rocks; (2) an axial zone of plutons with pendants of mid-Cretaceous silicic metavolcanic rocks and associated shallow intrusions; and (3) an eastern zone comprising large-volume, composite plutons generally ranging inward from tonalite, quartz diorite, granodiorite, and K-feldspar porphyritic granite that are similar overall to those of the Sierra Crest magmatic event. They defined the boundary between the two western zones, as did Kistler (1990) before them, to be delineated by the $Sr_i = 0.706$ isopleth, and the boundary between the axial and eastern zones as the western boundary of the large composite plutonic complexes.

Rocks in the southern Sierra and Tehachapis were exhumed from 9 kb at about 100 Ma to 4 kb by about 95 Ma (Saleeby et al., 2007). Also in the south, Saleeby et al. (2008) recognized that a major 2–5-km-wide shear zone (Busby-Spera and Saleeby, 1990), the proto–Kern Canyon fault–Eastern Tehachapi shear zone, separated various plutonic suites, with the 105–98 Ma Bear Valley suite, the 110–95 Needles suite, and the 105–102 Ma Kern River suite cropping out west of the shear, whereas the voluminous 95–84 Ma Domelands suite and the 100–94 Ma South Fork suite occur east of the fault. Nadin and Saleeby (2008) used geobarometry to suggest that the shear zone had ~10 ± 5 km of east side up displacement in its central part, but based on disruption of batholithic zonation, it might have as much as 25 km of shortening across its southern part. They also suggested that this deformation was under way by at least 95 Ma and was overprinted by dominantly dextral shear fabrics by 90 Ma as also deduced by Wong (2005). The area also contains a 77±5 Ma cooling event apparently caused by rapid collapse of the region on the southern Sierra detachment system (Chapman et al., 2012). This may be the same rapid 83–79 Ma cooling event noted by Maheo et al. (2004) to have occurred in the Mount Whitney intrusive suite farther north.

Some paleomagnetic results indicate that the Sierra Nevada was located 700 ± 500 km farther south (Housen and Dorsey, 2005); whereas others suggest it moved very little with respect to North America at least since ~83 Ma (Hillhouse and Grommé, 2011). New North American poles calculated by Kent and Irving (2010) add ~500 km to the possible discordance. However, reannealing of magmatic minerals and fabrics due to long residence times at high, but subsolidus, temperatures might have altered textures, ages, and magnetite grains (Pullaiah et al., 1975).

Salinian Block

Cretaceous plutonic rocks of the Salinian block (Figs. 5, 8, and 25), which is tectonically isolated by faults to both the NE and SW (Ross, 1978), are texturally, compositionally, and temporally similar to those of the Sierra Nevada, including an overall eastward younging (Ross, 1972; Mattinson, 1978b, 1990; Mattinson and James, 1985). Based on recently collected U-Pb data, Barth et al. (2003) suggested that many of the plutonic rocks of the Salinian block correlate closely with those of the Cathedral Peak intrusive series within the Sierra, and that the schist of Sierra de Salinas, generally correlated with the Rand-Pelona-Orocopia schists farther south (Ross, 1976; Jacobson et al., 2011) was, like their southern counterparts, thrust beneath the Cretaceous granitoids.

Post-Miocene right-lateral displacement of ~320 km on the San Andreas fault (Crowell, 1962, 1975, 1981), based on separation of the 23.5 Ma Pinnacles and Neenach volcanics (Matthews, 1976); submarine fan deposits of the Oligocene–Miocene La Honda basin in the Santa Cruz Mountains with the San Joaquin basin in the Temblor Range (Graham et al., 1989; Critelli and Nilsen, 2000); and the Logan gabbro of the Coast Ranges with similar rocks at Eagle Rest Peak in the San Emigdio Mountains (Ross, 1970; James et al., 1993) restores a substantial portion of the Salinian block south of the Sierra Nevada, but much still would lie west of the Great Valley, even after restoration of ~130 km dextral displacement on other faults such as the San Gregorio–Hosgri system (Ross, 1984). Thus, current restorations are problematic, and there must be separation on additional faults.

Great Valley Sedimentation and Deformation

According to Ingersoll (2008, p. 414), the Great Valley forearc basin (Figs. 8 and 25) "is the most thoroughly studied, best understood forearc basin on Earth … and is the type forearc basin against which all other forearc basins are compared." Although Ojakangas (1968) was the first to study sedimentation in the basin, it was before the advent of plate tectonics and so it was left to Dickinson (1970, 1971, 1976) to develop the generally accepted fore-arc model. The fill of the basin is represented by the latest Jurassic–earliest Cretaceous to Maastrichtian Great Valley Group (DeGraaff-Surpless et al., 2002; Surpless et al., 2006), which includes a stratigraphic succession greater than 15 km thick that nonconformably overlies slabs and breccias of the Coast Range ophiolite, the Tehama-Colusa serpentinite belt, and chert to the west and Sierran basement to the east (Ingersoll, 1982; C.A. Hopson et al., 2008).

On the basis of detrital mineralogy, Ingersoll (1983) divided rocks of the northern Great Valley Group into six petrofacies. The oldest three petrofacies contain significant quantities of sedimentary and metamorphic debris, possibly shed from the Klamath and northern Sierra terranes; whereas the younger middle Late Cretaceous petrofacies have much higher percentages of plutonic debris (Ingersoll, 1983). DeGraaff-Surpless et al. (2002) presented U-Pb analyses of detrital zircons from several sections within the Great Valley sequence. In general, the lowermost petrofacies, the Stony Creek, preserves rocks from the Tithonian biostratigraphic zone and contains youngest zircons with ages from 144 Ma to 135 Ma, whereas the remaining units have youngest zircons in the range 97 to 72 Ma, documenting that they were deposited in the Upper Cretaceous (Surpless et al., 2006). The Upper Cretaceous part of the Great Valley Group consists mainly of turbidites deposited in basin plain, fan, slope, and shelf depositional environments within a northerly-trending, asymmetric basin (Ingersoll, 1979) presently located between the Sierra Nevada foothills on the east and the Coast Range ophiolite–Franciscan complex to the west.

Constenius et al. (2000) documented a major discontinuity (Fig. 26)—across which are changes in structure, composition, and overall depositional pattern—between the lower two petrofacies, which coincides with the Barremian–Aptian boundary at 125 ± 1 Ma (Gradstein et al., 2004). In the northern Great Valley, rocks below the discontinuity are faulted, warped, and locally eroded. The origin of this deformation is poorly understood, and there is no consensus as to its origin (Constenius et al., 2000; Wright and Wyld, 2007; Dumitru et al., 2010).

During the Campanian–Maastrichtian, the basin was dramatically altered as its western side was abruptly uplifted, its depocenter migrated eastward (Fig. 27), the sedimentary regime went from deep water to shallow marine and alluvial, and paleocurrents switched from westerly to southerly (Moxon and Graham, 1987; Moxon, 1988; Mitchell et al., 2010). Based on the composition of sediments, there appears to have been little variation in source terrane, and the changes appear related to rapid uplift along the western side of the basin (Almgren, 1984; McGuire, 1988). High-grade metamorphic debris from the Franciscan complex, located to the west, appears in sediments of the Great Valley Group during the Maastrichtian (Berkland, 1973). Some workers (Wentworth et al., 1984; Unruh et al., 1991; Wakabayashi and Unruh, 1995) suggest a period of easterly directed tectonic wedging, which although it may have been

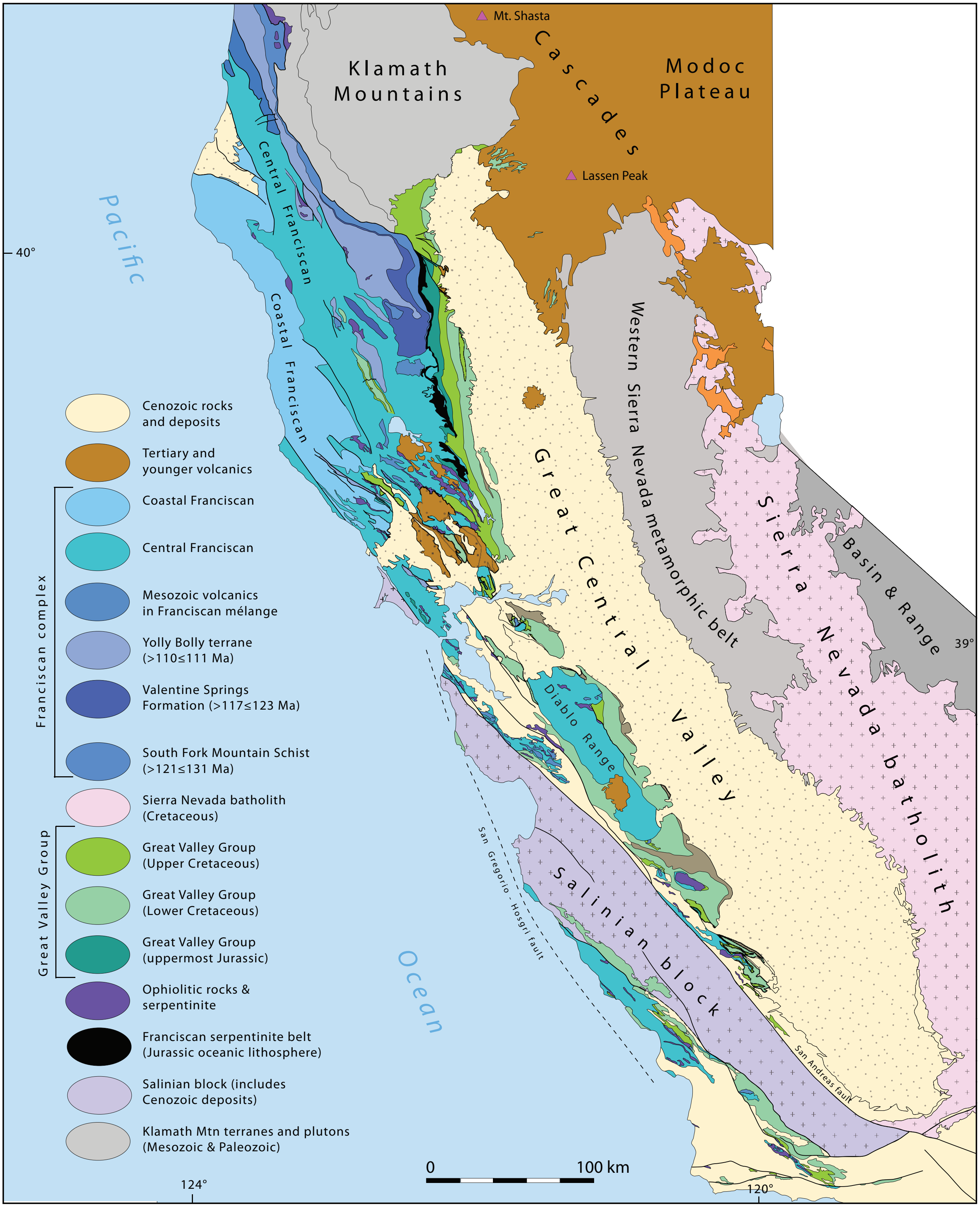


Figure 25. Geological sketch map showing the main belts and terranes of the Franciscan complex, the serpentinite belt, and the Great Valley sequence. Modified from Jennings (1977) and Dumitru et al. (2010).

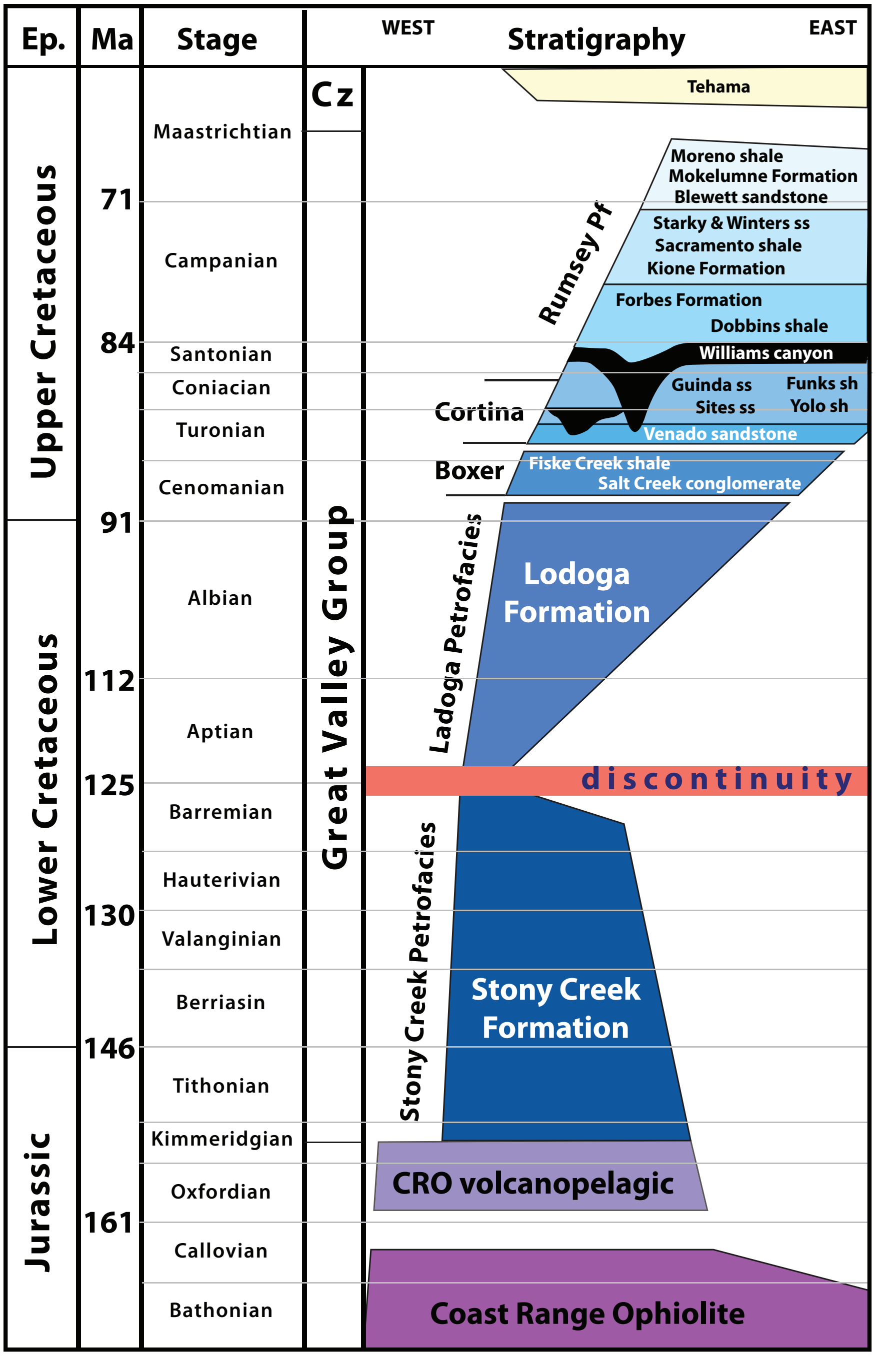


Figure 26. Petrofacies of the Great Valley group after Constenius et al. (2000) illustrating the 125 Ma discontinuity and the Upper Cretaceous disruption and channels as discussed in the text. CRO—Coast Range ophiolite; ss—sandstone; sh—shale.

most active during the early Tertiary, appears to be continuing today (Unruh and Moores, 1992).

While the petrological profiles through the Great Valley Group are widely assumed to represent deroofing of the Sierra Nevada, their provenance is not restricted to the Sierran batholith and would fit erosional deroofing of any of the Cordilleran batholiths of western North America. In fact, Wright and Wyld (2007) examined detrital zircon suites from the lowermost Cretaceous–uppermost Jurassic parts of the sequence and found many with ages of ~980 Ma and 1.4–1.6 Ga, which suggested to them that rocks of the western Great Valley Group were displaced from an original locus of deposition near Oaxaquia (Fig. 5) presently located in Mexico.

Thus, the sedimentary rocks of the Great Valley group reflect significant changes in the basin at ~125 Ma, when the Late Jurassic–earliest Cretaceous sediments were deformed and locally eroded, and at ~80–75 Ma, when the western side of the basin was uplifted and sedimentation went from deep to shallow marine. The uplift and exhumation were rapid enough that blueschists of the Franciscan complex were exposed at the

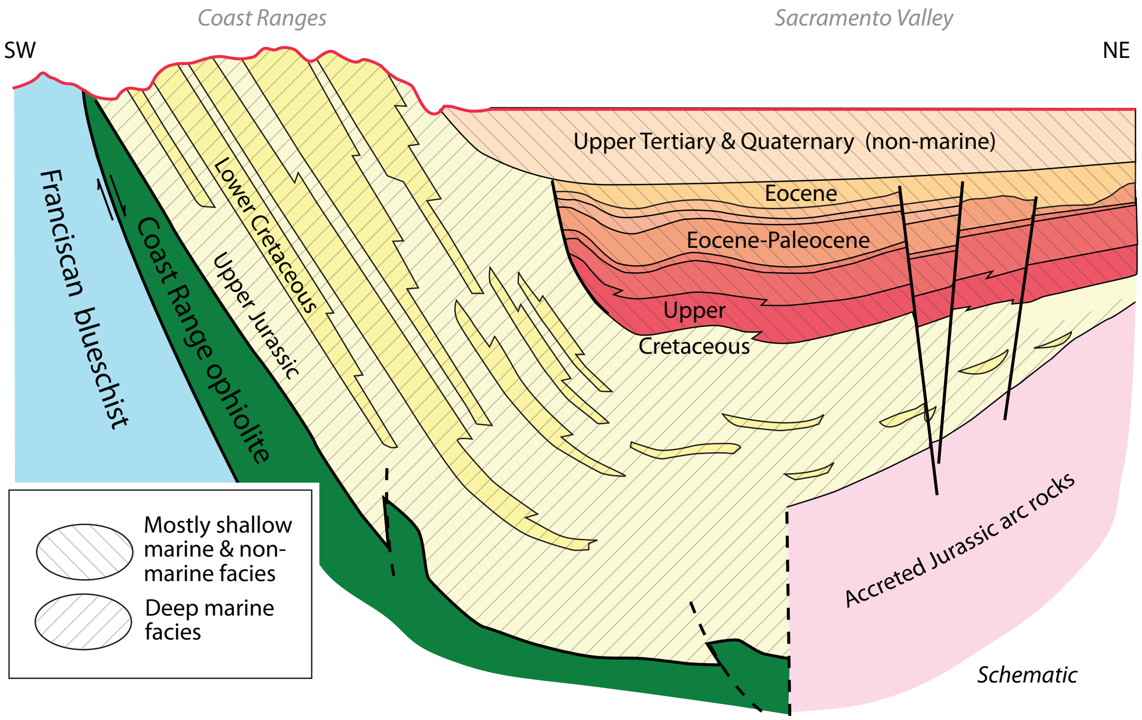


Figure 27. Schematic cross section of the Great Valley group from Dickinson and Seeley (1979), illustrating the eastward migration of the basin axis and the abrupt change during the Upper Cretaceous from deep marine to shallow marine facies. The uplift is interpreted here to have been created by post–slab-failure exhumation.

surface, eroded, and shed into the basin during the Maastrichtian at ~67 Ma (Berkland, 1973).

A basinal remnant called the Hornbrook basin occurs along the northeast side of the Klamath Mountains (Fig. 8) and is generally considered to represent a fragment of the same fore-arc basin as that of the Great Valley Group (Nilsen, 1986; Kleinhans et al., 1984; D.M. Miller et al., 1992). The Hornbrook is an east-northeasterly-dipping remnant of a 125–85 Ma sequence of terrigenous clastics that sits unconformably on rocks of the eastern Klamath Mountain terranes and occupies an arcuate region along the California-Oregon border, where they constitute basal alluvial fan deposits fining upwards through sandstones to siltstones (Nilsen, 1993; Beverly, 2008). Although the rocks were generally considered to have been derived dominantly from the Klamaths, a recent detrital zircon study of rocks in the basin showed that they contain large numbers of zircons much younger than any known magmatic rock outcropping nearby and so, except for the lowermost parts, were probably mostly derived from one of the Cordilleran batholiths such as the Sierra Nevada or Coast plutonic complex (Beverly, 2008).

Coast Range Ophiolite

The Coast Range ophiolite (Fig. 25) is a dismembered sequence of 168–161 Ma ultramafic, gabbroic, and basaltic rocks—interpreted to represent ancient sea floor formed at a spreading ridge—that are disconformably overlain by distal Oxfordian tuffaceous radiolarian chert and mudstone grading up into more proximal Tithonian volcaniclastic facies (Hull et al., 1993; C.A. Hopson et al., 2008). Paleomagnetic and biostratigraphic evidence indicate that all of the ophiolitic remnants formed close to the paleoequator and migrated northward through a zone of nondeposition, followed by volcanogenic deposition, locally a Late Jurassic (152–144 Ma) disruption that created ophiolitic breccias, and finally to be overwhelmed and buried by a thick apron of uppermost Jurassic siliciclastic turbidites of the Great Valley Group (Pessagno et al., 2000; C.A. Hopson et al., 2008). While these workers suggested that the ophiolite developed in the open ocean, other workers favor a suprasubduction origin either in a fore-arc (Shervais, 2001; Shervais et al., 2004, 2005) or back-arc (Godfrey and Dilek, 2000) setting. The geochemical arguments for and against each model are summarized in the preceding references and are beyond the scope of this paper.

A peculiar and unexplained feature of the section that may be germane to our discussion is the ophiolitic breccia unit, 0–500 m thick, comprising fragments of ophiolitic debris, found locally between the ophiolite and the Great Valley Group (Hopson et al., 1981; Robertson, 1990; C.A. Hopson et al., 2008). As summarized by C.A. Hopson et al. (2008), the ophiolite beneath the breccia unit was broken by a complex system of faults that lived long enough such that blocks of earlier-formed and partially-cemented breccias were shed from higher-standing blocks. Mafic magmatism and hydrothermal alteration accompanied the faulting and in the Elder Creek remnant a dike, dated at 154 ± 5 Ma, is overlain by breccia (Blake et al., 1987). Robertson (1990) favored simple normal faulting and collapse of high-standing blocks for the origin of the breccias, but C.A. Hopson et al. (2008) found that model unable to explain the volumes of fragmental debris and so suggested that the breccias formed due to passage of the sea floor through a transform zone. Another possibility, that might create similar breccias, alteration, and magmatism, is the system of normal faults, many kms long with 100–500 m vertical separations, found associated with 4–9 Ma fields of alkali basaltic lavas, peperites, and reworked hyaloclastites, entering the Japan trench today, and apparently generated as the oceanic plate was flexed prior to passing over the 800-m-high outer trench rise (Hirano et al., 2001, 2006). Both the flexural and the transform models would fit the observation that volcanopelagic sedimentary rocks locally sit atop the breccia (Robertson, 1990; C.A. Hopson et al., 2008). The Franciscan accretionary complex also contains a suite of alkaline gabbros emplaced into the sediments shortly before their incorporation into the accretionary prism (Mattinson and Echeverria, 1980; Mertz et al., 2001), which probably also reflect passage over the outer trench rise.

A serpentinite matrix mélange, known as the Tehama-Colusa serpentinite mélange (Hopson and Pessagno, 2005), occurs between rocks of the Franciscan complex and those of the Great Valley Group along the eastern side of the Coast Ranges west of the Sacramento Valley (Fig. 25). It is typically included on maps as part of the Coast Range ophiolite (Jennings, 1977), but geochemical study suggests that the mélange represents dismembered Franciscan oceanic crust and mantle plus its abyssal sedimentary veneer (Shervais and Kimbrough, 1985). The mélange, apparently affected only by low-temperature hydrous alteration, generally occurs in fault-bounded slivers, sits structurally beneath the Coast Range ophiolite–Great Valley package, and is interpreted to represent Jurassic oceanic material that originated in paleoequatorial regions similar to the Coast Range ophiolite (Hopson and Pessagno, 2005).

Franciscan Complex

The Franciscan complex outcrops in the California Coast Ranges (Fig. 25) and due to its lithologies, chaotic nature, high P–low T metamorphism, and a systematic eastward progression in metamorphic grade, is generally considered as the archetypical example of a subduction complex (Hamilton, 1969a; Ernst, 1970; Blake et al., 1988). Rocks included within the complex are generally divided into three belts: Eastern, Central, and Coastal following the scheme of Berkland et al. (1972).

The Eastern belt contains two distinct terranes: the Pickett Peak and Yolla Bolly, rocks of which were both metamorphosed to the blueschist facies (Blake and Jones, 1981) and are separated by east-dipping thrust faults. The Pickett Peak—which contains two subunits, also separated by east-dipping thrusts: the South Fork Mountain schist (SFMS) and Valentine Spring formation (VSF)—is the structurally highest and lies east of the Yolla Bolly (Worrall, 1981). Locally, the SFMS contains some thrust sheets of mid-ocean ridge basalt (MORB)–like metabasalt (Wakabayashi et al., 2010) topped with chert, which contained detrital zircons dated at 137 Ma, but overall it is dominantly very fine grained and its protolith was likely mudstone (Dumitru et al., 2010). The best estimate of its metamorphic age is a $^{40}Ar/^{39}Ar$ plateau age of 121 Ma from white mica (Wakabayashi and Dumitru, 2007; Dumitru et al., 2010). The Valentine Spring Formation—which sits structurally beneath the SFMS, contains more abundant metagraywacke, and is apparently less metamorphosed than those rocks—has detrital zircons as young as 123 Ma and gave $^{40}Ar/^{39}Ar$ ages of ~117 Ma (Dumitru et al., 2010). The Yolla Bolly terrane contains more chert than the Pickett Peak and was intruded by sills of alkali basalt (Blake and Jones, 1981; Isozaki and Blake, 1994) at ~119 Ma (Mertz et al., 2001). Although generally considered to be trench magmatism, such intrusions seem compositionally quite similar to magmatism erupted just outboard of the outer swell off Japan at 5.9 Ma, but now broken by normal faults and located within the Japan Trench (Hirano et al., 2001, 2006).

The Central belt is dominantly a tectonic mélange containing slabs and blocks of blueschist derived from the Eastern belt, pillow lava capped by either chert or limestone (~88 Ma), and a variety of exotic blocks of high-grade blueschist, eclogite, and amphibolite, all engulfed in a sheared argillitic matrix containing interbeds of graywacke (Blake and Jones, 1981). Several workers have analyzed a few detrital zircons here and there within the belt and the youngest range from 110 to 78 Ma (Snow et al., 2010; Morisani et al., 2005; Tripathy et al., 2005; Joesten et al., 2004).

The Coastal belt contains the westernmost units of the Franciscan complex and is exposed in the Coast Ranges of northern California (Fig. 25) where it comprises disrupted terrigenous clastics holding blocks of pillow basalt, pelagic limestone, and rare blocks of blueschist (Blake et al., 1988). Sandstones within the eastern parts of the belt are more arkosic, contain sparse volcanic and cherty debris, and contain laumontite; whereas overall the belt is not known to contain newly formed blueschist minerals and is of lower grade than more eastern belts (Blake and Jones, 1981). Based upon detrital zircons and microfossils, it appears to be early to mid-Tertiary, probably mostly Eocene to Miocene, in age (Evitt and Pierce, 1975; Blake et al., 1988; Tagami and Dumitru, 1996; Snow et al., 2010).

In a landmark paper, Dumitru et al. (2010) demonstrated with U-Pb analyses of detrital zircons that the main pulse of sedimentation within the Franciscan mélange started at ~123 Ma, and that high-grade exotic blocks and small slabs within the mélange are distinctly older, with metamorphic ages in the range 169(?) to 132 Ma (Mattinson, 1986; Anczkiewicz et al., 2004). Many of the exotic blocks are blueschist, amphibolite, and eclogite with rinds of actinolite-chlorite suggesting that they were formerly engulfed in serpentinite (Coleman and Lanphere, 1971; Cloos, 1986) and thus not necessarily related to the same subduction zone as the main bulk of the mélange. Northerly-trending belts of rocks exposed just to the east in the Tehama-Colusa serpentinite or the Coast Range ophiolite (C.A. Hopson et al., 2008) or in the western Sierra, such as the Kings-Kaweah serpentinite mélange (Saleeby, 1977; Saleeby and Sharp, 1980), the Smartville-Foothills arc block (Menzies et al., 1980; Dilek, 1989; Day and Bickford, 2004), and the Slate Creek–Lake Combie arc belt (Edelman et al., 1989a, 1989b; Fagan et al., 2001) and their intervening sutures may have been the source for the exotic blocks, but there might also have been considerable along-strike movement along buried faults beneath the Great Central valley (Wright and Wyld, 2007) so their source might be obscure.

Based on the ages of many high-grade blocks within the Franciscan, it is commonly assumed that eastward subduction started ~169–165 Ma (Wakabayashi and Unruh, 1995; Anczkiewicz et al., 2004; Wakabayashi and Dumitru, 2007; Dumitru et al., 2010), but as discussed earlier, the Smartville complex, an arc located on the upper plate, wasn't accreted until 159 Ma, so it seems unlikely that subduction would step westward to the other side of the arc until the collision was complete, although it is possible. All of the oldest published ages for blocks of the Franciscan, except one Lu-Hf age on garnet, are within analytical error of 159 Ma (Wakabayashi and Dumitru, 2007), and the single older age of 168 Ma reported by Anczkiewicz et al. (2004) suffers from uncertainties in the ^{176}Lu decay constant. Thus, although pieces may have been derived from older terranes, the Franciscan itself shouldn't be older than 159 Ma.

It is, however, possible that subduction had commenced prior to 140 Ma as a crudely N-S linear group of plutons (Figs. 17, 18, and 19) intruded the Klamaths and the Sierran Foothills belt at ~140 Ma (Saleeby et al., 1989; Irwin and Wooden, 2001; Day and Bickford, 2004). They may be the first magmatic products of the subduction ultimately responsible for the Franciscan complex, but as they represent an isolated and short-lived 2 or 3 Myr pulse of magmatism, another cause, such as slab failure, might be more likely.

Apparently, the youngest blueschists in the Franciscan complex are of Coniacian to Santonian age and are located in the Burnt Hills terrane and at Pacheco Pass in the Diablo Range (Fig. 25) (Blake et al., 1985, 1988; Wakabayashi and Unruh, 1995; Wakabayashi and Dumitru, 2007; Ernst et al., 2009a; A. Jayko, 2010, personal commun.), which suggests a significant change in the subduction regime at about that time.

Late Cretaceous Deformation and Metamorphism

The best known deformational features of the Cordillera are the Late Cretaceous to Eocene basement-involved Laramide uplifts and associated basins of the Rocky Mountain region (Fig. 8) that occur within the Great Basin segment (Dana, 1896). The uplifts generally have the form of asymmetrical anticlines with cores of Precambrian basement bounded by thrust faults, or steep to overturned monoclines that faced, and in many cases overrode, deep basins that subsided to receive sediment during rise of adjacent areas. Many of these features have lengths of tens to hundreds of kilometers, have structural relief between basin-uplift pairs of 5–12 km, and involve the entire crust (Grose, 1974; Smithson et al., 1979; Brewer et al., 1982; Rodgers, 1987; Hamilton, 1988b).

The Laramide features reflect fundamental changes in structural style and sedimentation within the Great Basin segment of the Cordillera from the Sevier deformation in that deformation changed from thin skinned to thick skinned, and the sedimentation from dominantly marine foreland basin sedimentation to deposition in localized, isolated nonmarine basins (Dickinson et al., 1988; Beck et al., 1988). Although there is some spatial and temporal overlap between the two styles of deformation (Kulik and Schmidt, 1988), the overall pattern of laterally continuous foreland basin sedimentation was generally followed by the development of localized depocenters and associated thick-skinned deformation such that there are two deformational episodes with only minor temporal overlap between the two (Armstrong, 1968).

Thrusting within the Sevier fold and thrust belt continued, but it was much diminished compared to the earlier Cretaceous phase. In the Wyoming salient, the dominant thrusting occurred on the Crawford and Absaroka thrusts, and older Aptian–Cenomanian thrusts were folded into large anticlines (Yonkee and Weil, 2011). Farther south, the older Canyon Range thrust, also Aptian–Cenomanian, was folded into a large anticlinorium, and movement on smaller thrust and duplexes took place (DeCelles and Coogan, 2006).

The hinterland belt in the Great Basin sector has two phases of deformation, one Jurassic and the other Late Cretaceous (see, for example, Snoke and Miller, 1988). The two deformations and at least two periods of intense normal faulting make it difficult to resolve many finer details, but in general the Late Cretaceous deformation included thrusting, back folding, and nappe formation (Camilleri et al., 1997; Snoke et al., 1997; McGrew and Peters, 1997).

Late Cretaceous deformation occurred in the Tehachapi Mountains. There 100 Ma plutons were recumbently folded and thrust westward prior to 95 Ma exhumation (Wood, 1997; Saleeby et al., 2007, 2008). Later, rocks of the probable Late Cretaceous Witnet Formation, which sit unconformably upon 92 Ma granitoids (Chapman et al., 2012), were folded and overthrust from the south by 92 Ma granitoids (Wood, 1997).

Within the Sierra Nevada, rocks of the Goddard pendant were deformed after 131 Ma and before 90 Ma (Tobisch et al., 1995; Bateman, 1992). Elsewhere within the Sierra Nevada, most of the Cretaceous rocks within the pendants were folded at least twice.

West of the Sierra Nevada, rocks of the Great Valley group were deformed in the latest Cretaceous–early Tertiary by folding and thrusting (Unruh et al., 1991). Large submarine canyons were cut into older rocks (Fig. 26) at that time (Williams et al., 1998).

Rand-Pelona-Orocopia-Swakane Subduction Complex

Generally considered to represent rocks of a subduction complex and outcropping in a NW-SE trending band extending from just south of Monterrey Bay to SW Arizona (Jacobson et al., 2011; Haxel, 2002) are discontinuous exposures of peculiar schists, variously named Sierra de Salinas, San Emigdio, Rand, Pelona, and Orocopia schists (Fig. 28). The schists—originally interpreted to lie beneath a major thrust fault, but now generally considered to sit beneath low-angle normal faults where they occupy cores of exhumed areas—are dominantly quartzofeldspathic, with transposed lithologic layering, metamorphic mineral assemblages that belong mainly to the albite-epidote amphibolite facies, and interpreted to have been graywacke, sandstone, chert, mudstone, and basalt prior to deformation and metamorphism (Haxel and Dillon, 1978; Ehlig, 1981; Frost et al., 1982; Malin et al., 1995; Oyarzabal et al., 1997; Wood and Saleeby, 1998; Haxel et al., 2002; Jacobson et al., 1988, 1996, 2002, 2007, 2011). In general the protolith age is within ~5 Myr of their exhumation age and in their present distribution, both the protolith and emplacement ages of the schists are progressively older from southeast to northwest, ranging from less than 60 Ma in the southeast to 90 Ma in the northwest (Grove et al., 2003b; Jacobson et al., 2011). When displacements on the faults of southern California are restored (Powell, 1993; Nourse, 2002), the eastern occurrences form an E-W band extending across much of southern California and western Arizona (Fig. 8).

Although recent workers have lumped all the schists together to form a continuum of NE-SE decreasing age (Jacobson et al., 2011), it is possible that there are two, or even three, different periods of formation. In most pre–San Andreas reconstructions, one group of schists, the Rand, San Emigdio, Sierra de Salinas, and Portal Ridge exposures, lies near the southern end of the Sierra Nevada and is separated from, and mostly older than, the

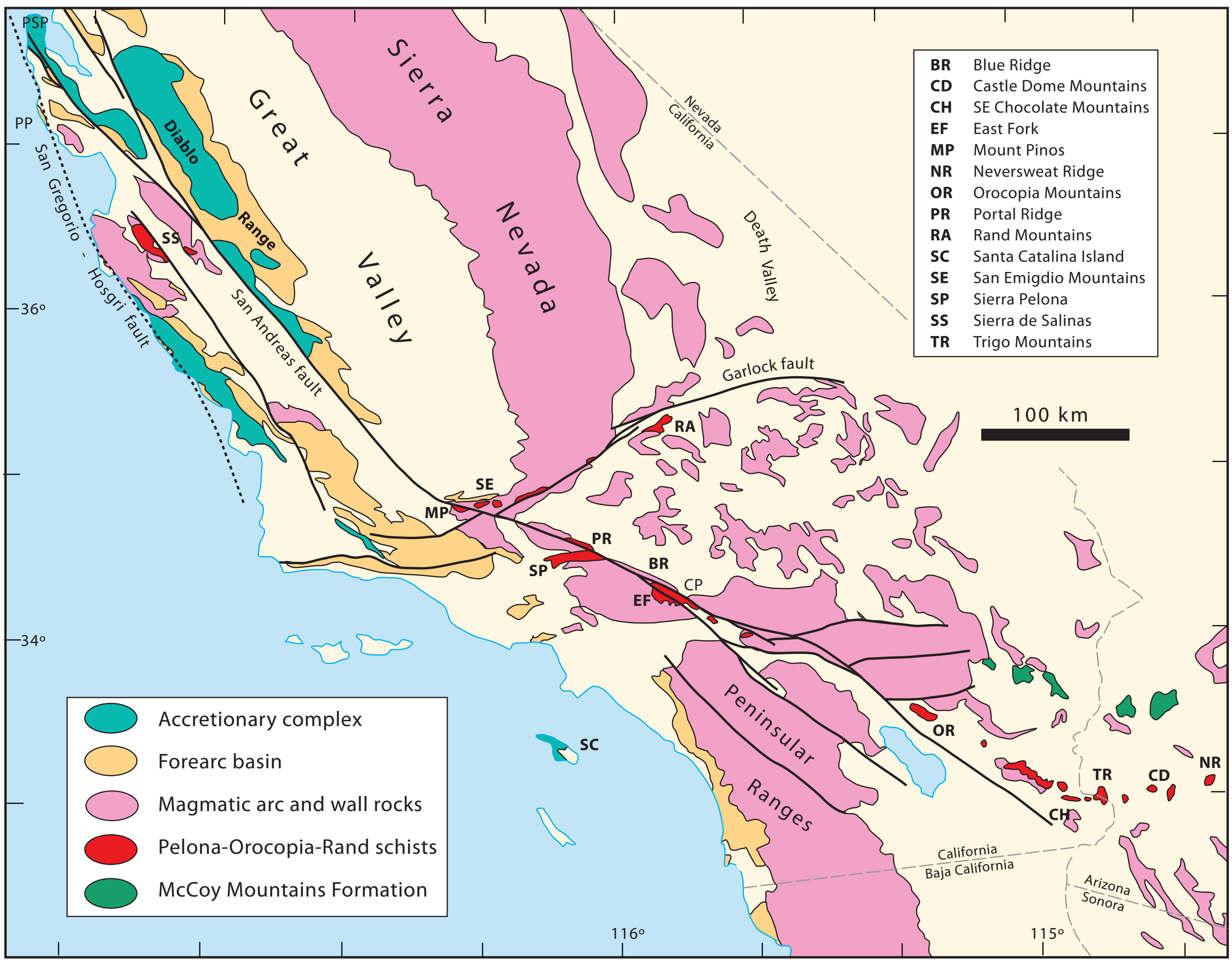


Figure 28. Generalized geologic map showing distribution of subduction schists. Modified from Jennings (1977) and Jacobson et al. (2011).

Orocopia-Pelona exposures, which are dominated by Early to Middle Cretaceous detrital zircons; whereas many, but not all, of the Pelona-Orocopia exposures to the southeast are dominated by Proterozoic and Late Cretaceous detrital zircons (Grove et al., 2003b; Jacobson et al., 2011). Additionally, the exposures within the San Emigdio Mountains and southern Sierra Nevada are some 10–15 Myr older, and have very different detrital zircon profiles, than the Rand schists south of the Garlock fault, but are quite similar, both in terms of age and provenance, to the Catalina schist (Jacobson et al., 2011; A. Chapman, 2011, personal commun.). Thus, the best division might be by age: the pre-Laramide schists, Santa Catalina, San Emigdio, and Portal Ridge; and the post-Laramide schists (Fig. 29).

Remarkably similar rocks, generally not considered with the Rand-Pelona-Orocopia outcrops, belong to the Swakane gneiss, located within the North Cascades of Washington (Fig. 8) at the south end of the Coast plutonic complex (Matzel et al., 2004). The gneiss is a quartzo-feldspathic amphibolite-grade rock (9–12 kbar and 640–740 °C) of sedimentary protolith containing detrital zircons ranging in age from 161 to 73 Ma, a leucogranite inferred to be a partial melt of the schist dated at 68 Ma, and a hornblende $^{40}Ar/^{39}Ar$ date of 57.9 ± 0.5 Ma (Matzel et al., 2004). Recently, Gatewood and Stowell (2012) argued that deposition must have taken place prior to about 75 Ma and that younger zircons are metamorphic, but zircon morphologies suggest that they are magmatic, not metamorphic (S. Bowring, 2012, personal communication). Overall, the gneiss is most similar temporally to the Pelona-Orocopia grouping of schists (Fig. 29).

SONORAN SECTOR

Transverse Ranges

The Transverse Ranges are a group of mountain ranges that extend in a more or less easterly direction from the California coast to southeasternmost California (Fig. 21). At an earlier time the ranges trended more northerly, but in the past 20 Myr, the western ranges were captured by the Pacific plate and rotated clockwise ~80°–110° (Kamerling and Luyendyk, 1985; Nicholson et al., 1994).

In the northern San Gabriel Mountains (Fig. 21), the Vincent thrust carries 1.7 Ga granulitic Mendenhall gneiss and a 1.2 Ga anorthosite-syenite-gabbro complex over Pelona schist (Barth et al., 1995). Several Cretaceous terranes were delineated in the southern part of the range by May and Walker (1989) including Lower Cretaceous granulite facies gneisses of the Cucamonga

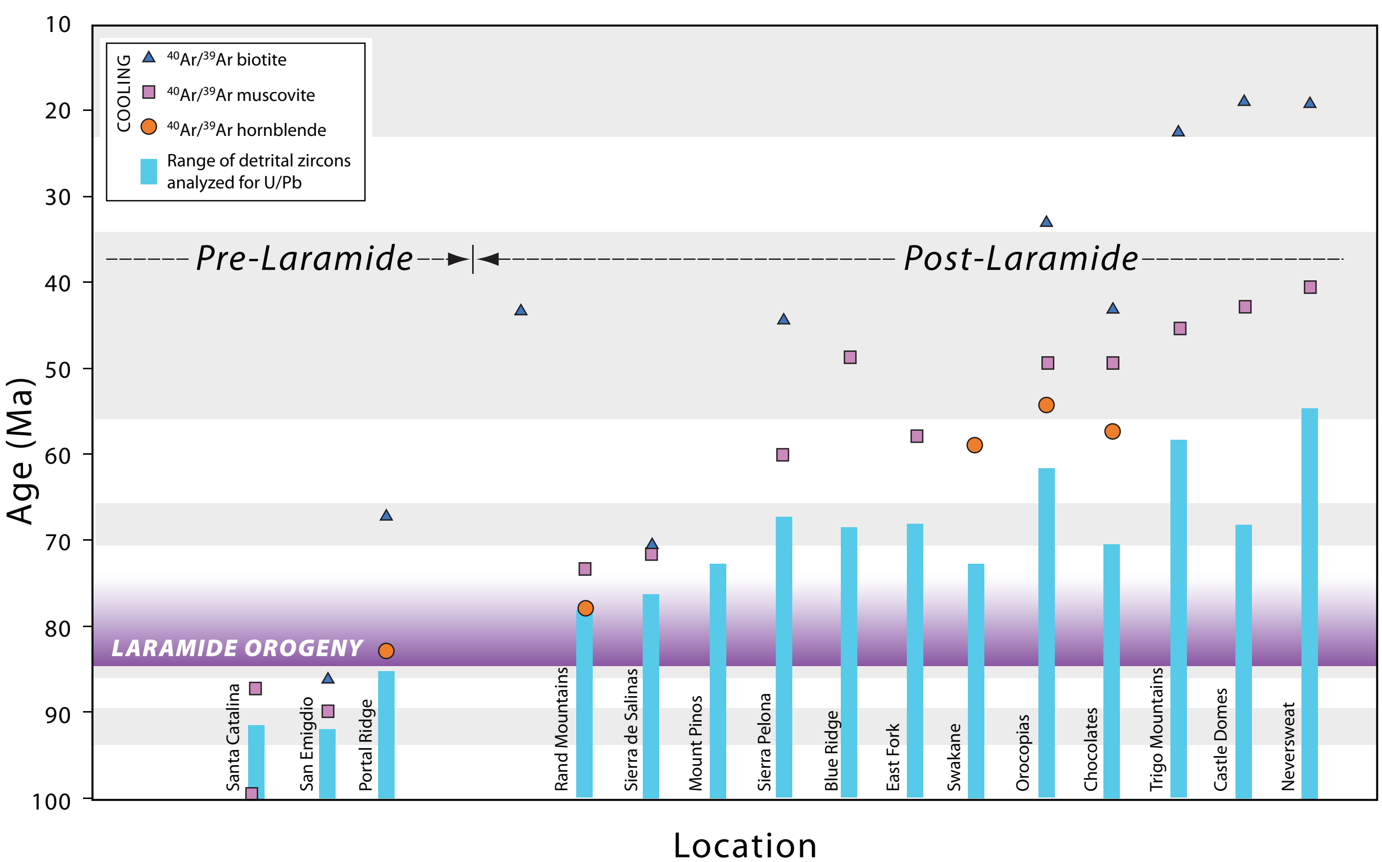


Figure 29. Argon isotopic ages and U-Pb detrital zircon ages for Rand-Swakane-Orocopia-Pelona schists and similar rocks. Note that they are readily divisible into pre- and post-Laramide groups. Modified from Jacobson et al. (2011); Swakane data from Matzel et al. (2004).

block and the San Antonio block comprising Late Cretaceous plutons, mylonites, and gneisses with pendants of amphibolite-grade metasedimentary rocks.

The bulk of the rocks of the San Bernardino Mountains occur east of the San Andreas fault (Fig. 21) and comprise Paleoproterozoic granitoids and gneisses unconformably overlain by Neoproterozoic and Paleozoic metasedimentary rocks (Cameron, 1982).

Powell (1993) divided the various Mesozoic magmatic rocks in the Transverse Ranges (Fig. 21) into three broad northwesterly trending belts: eastern, central, and western, with the eastern belt characterized by quartz-poor, alkali feldspar porphyritic bodies ranging in age from ~165 to 150 Ma, the Central belt mainly characterized by the presence of Proterozoic–Paleozoic basement, and the western belt by dominantly foliated 120–85 Ma plutons. More recent U-Pb dating of zircons by Barth et al. (2008a) and Needy et al. (2009) revealed many Late Jurassic alkaline and high-K calc-alkaline dioritic-gabbroic to syenitic-quartz monzonitic plutons in the San Bernardino, Little San Bernardino, and western San Gabriel mountains (Fig. 21) with ages between 156 and 149 Ma and several foliated calc-alkaline bodies, including sheeted bodies, in the Hexie, Pinto, and Little Bernardino mountains (Fig. 21) in the 80 to 74 Ma range; and sparse plutons in the 181 to 167 Ma range.

The 159 ± 7 Ma Corn Springs granodiorite crops out within the Chuckwalla and Little Chuckwalla mountains, part of the eastern Transverse ranges (Fig. 21) and was cut by mylonites, which were in turn intruded by a porphyritic granite and a comingled diorite, with an age of 150 Ma, and a much younger 74 ± 6 Ma granodiorite (Davis et al., 1994). Those authors suggested that the post–159 Ma deformation was widespread within the region, used geobarometry to indicate abrupt postcompressional exhumation following the deformation, and argued for extension during the emplacement of the 150 Ma bodies and the 148 Ma Independence dikes. Just to the north of the Chuckwallas, in the Eagle Mountains, the Eagle Mountain intrusion, a compositionally heterogeneous intrusion ranging from diorite and tonalite to monzogranite was emplaced into Precambrian gneiss and lower Paleozoic metasedimentary rocks at 165 Ma and later cut by Independence dikes (Mayo et al., 1998; James, 1989).

In extreme southeastern California (Fig. 21), within the hanging wall of the Chocolate Mountains detachment, are outcrops of dacitic lavas, 80–100 m thick, intercalated with minor graywacke beds, and overlain by quartz arenite and argillite with minor conglomerate beds (Haxel et al., 1985; Jacobson et al., 2002). Haxel et al. (1985) correlated the volcanic rocks with Jurassic metavolcanic rocks of Slumgullion in the Dome Rock Mountains of southern Arizona, and the overlying sedimentary rocks with the Jurassic–Cretaceous McCoy Mountains Formation.

Southern Arizona

Jurassic volcanic rocks are common within the Arizona-Sonora segment of the Sonora Desert, and like the Mojave region, the area was strongly distended during the Tertiary; so outcrops are widely scattered in the numerous, fault-bounded ranges separated from one another by broad alluvial valleys (Fig. 30). The area is bounded on the north by the WNW-striking Phoenix fault, which separates the extended region from the Transition zone of the Colorado Plateau (Hildebrand, 2009).

In west-central Arizona, Mesozoic volcanic rocks, deformed by both Cretaceous thrusting and Tertiary extension, are exposed in several ranges, either within the upper plates of detachment faults or within thrust stacks (Reynolds et al., 1987, 1989; Richard et al., 1987). The following description comes from these studies. In the Rawhide Mountains (Fig. 30), the 155 Ma Planet volcanics, which sit in fault blocks on the upper plate of the Buckskin-Rawhide detachment, comprise up to 600 m of variably deformed metavolcanic rocks, such as rhyolite and rhyolitic ash-flow tuff intercalated with volcaniclastic rocks and andesitic lavas. In the Granite Wash Mountains (Fig. 30), metamorphosed and complexly deformed hypabyssal porphyries and rhyolite interbedded with volcaniclastic rocks, all apparently unconformably overlain by sedimentary rocks presumed correlative with the McCoy Mountains Formation, are overthrust from the NE by thrust sheets of Proterozoic and Jurassic crystalline rocks. In the Harquahala Mountains (Fig. 30), rhyolitic-rhyodacitic ash-flow tuffs, minor andesitic-dacitic lavas and epiclastic rocks of the 156 ± 10 Ma Black Rock volcanics were intruded by hypabyssal porphyries, overlain by clastics correlated with rocks of the McCoy Mountains Formation, and outcrop within the structurally lowest thrust plate.

In southeasternmost California, southwestern Arizona, and northwestern Sonora, Tosdal et al. (1989) described two sequences that contain metavolcanic rocks: the Early Jurassic (205–170 Ma) Fresnal Canyon sequence, which locally consists of more than 7–8 km of rhyolitic-dacitic ash-flow tuff, lava flows and related breccias, local concentrations of andesitic lavas, and hypabyssal porphyries; and the younger Artesia sequence, a highly variable amalgam of metasedimentary rocks, in part Oxfordian in age, and metavolcanic rocks, such as rhyolitic to basaltic lavas and tuffs, cut by 159–145 Ma plutons of the bimodal, alkaline Ko Vaya plutonic suite. They also divided the Jurassic plutonic rocks (Fig. 31) into three different groups based on composition and age: (1) the extensive Kitt Peak–Trigo Peak super-unit, which consists of a compositionally continuous, and progressively younging (170–160 Ma), suite ranging from hornblende diorite to granite; (2) the Cargo Muchacho super-unit, which was defined as a separate suite because it occurs in thrust sheets structurally above the McCoy Mountains Formation, but elsewise is compositionally and temporally (173–159 Ma) equivalent to the Kitt Peak–Trigo Peak super-unit; and (3) the somewhat bimodal Ko Vaya super-unit, which is largely composed of granite with lesser amounts of diorite, and all with ages in the range 159 to 145 Ma.

Rocks of the >8-km-thick Topawa Group crop out in the Baboquivari Mountains (Fig. 30) and comprise ~170 Ma rhyolitic-dacitic volcanics, epiclastic rocks, minor alkali basalt

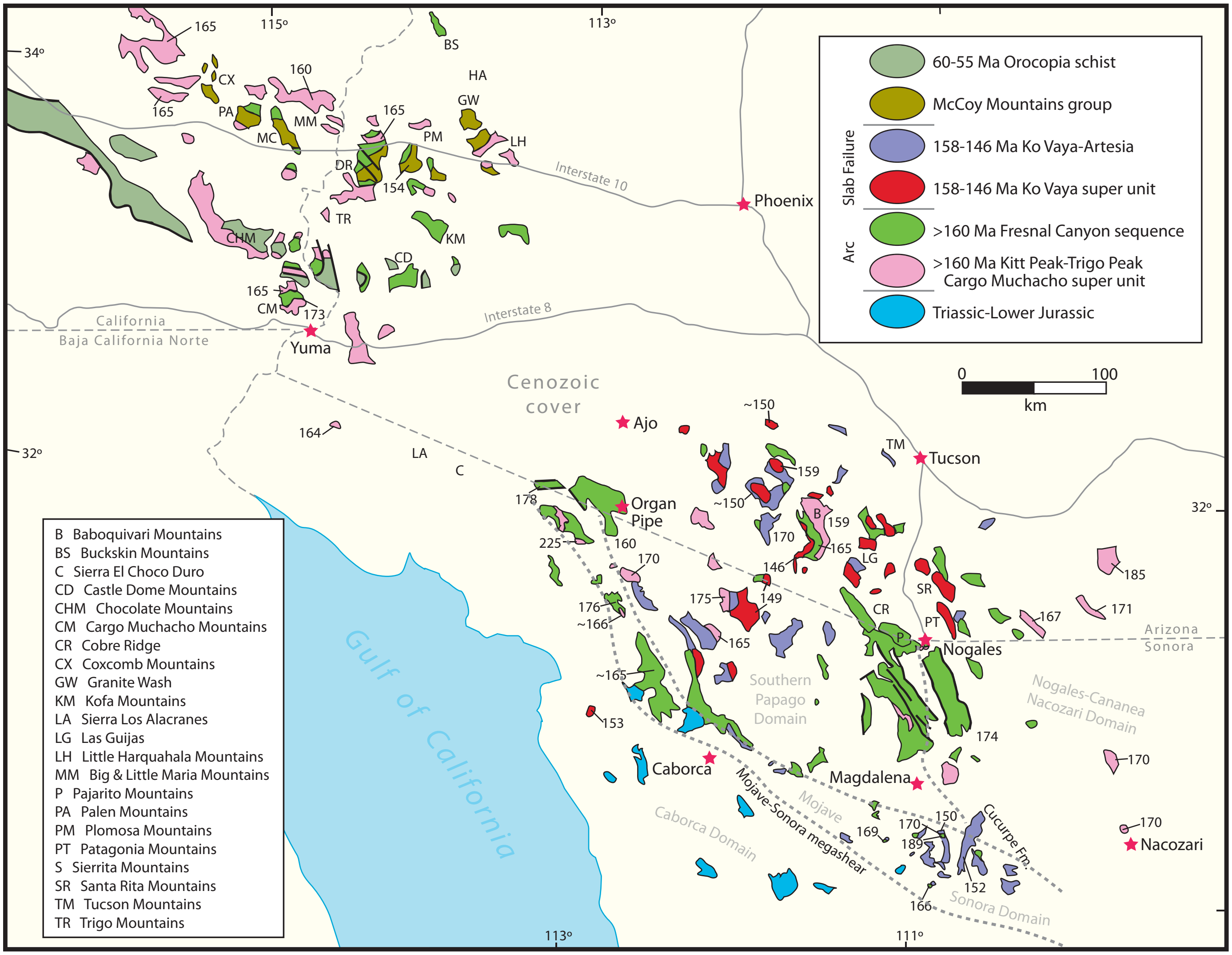


Figure 30. Geological sketch map of southeastern California, southern Arizona, and northern Mexico, showing the various ranges discussed in the text as well as the distribution of various Mesozoic rock units. Also shown are the regional units of Anderson and Silver (2005) in northern Mexico. Modified from Mauel et al. (2011).

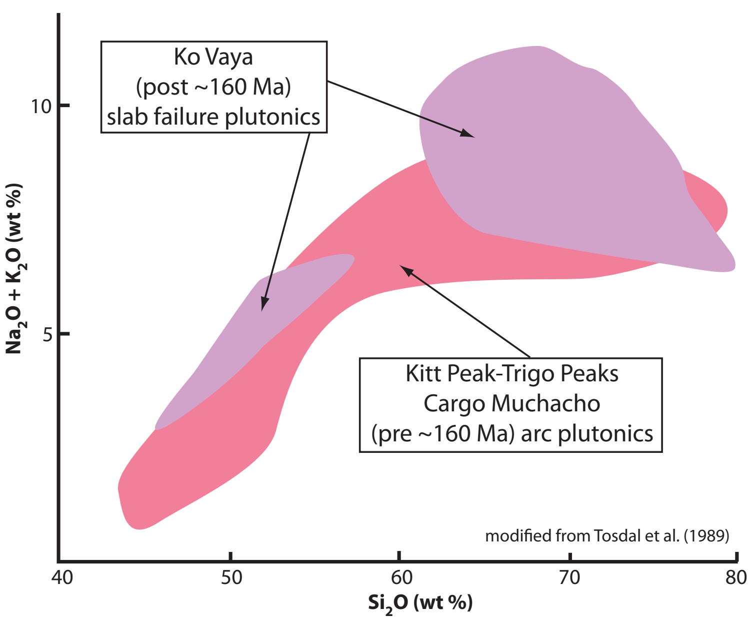


Figure 31. Harker variation diagram modified from Tosdal et al. (1989) illustrating the bimodal nature of the postcollisional Ko Vaya suite, interpreted here to represent slab-failure magmatism, and more continuous compositional nature of the older Kitt Peak–Trigo Peaks Cargo Muchacho plutons, generally accepted as arc related.

and comendite, all cut by hypabyssal intrusives and dioritic-granitic plutons of the 165–159 Ma Kitt Peak suite and by a 146 Ma perthite granite of the Ko Vaya suite (Haxel et al., 1980, 1982, 2005). Similar age rocks are exposed in several surrounding ranges: to the east in the San Luis, Las Guijas, and Pajarito mountains (Fig. 30) are rocks of the 170 ± 5 Ma Cobre Ridge tuff (Riggs et al., 1993); in the Santa Rita Mountains are extensive exposures of the ~183–170 Ma Mount Wrightson Formation, which consists of andesitic-dacitic lavas and breccias overlain mainly by siliceous lavas and tuffs; to the NE within the Sierrita Mountains (Fig. 30) are up to 1.3 km of andesitic lavas, siliceous lavas, and tuffaceous rocks of the Ox Frame volcanics cut by a 175 Ma pluton (Cooper, 1971; Spencer et al., 2003); and to the NW in the Comababi Mountains (Fig. 30) are similar andesites and rhyolites within the Sill Nakya Formation (Haxel et al., 1978), which are overlain by Late Jurassic conglomerates (Bilodeau et al., 1987). Along the international border south-southeast of Tucson (Fig. 30), volcanic rocks, approximately dated at 177 and 169 Ma by K-Ar on biotite, and including thick sequences of probable intracauldron rhyolitic ash-flow tuff, are overlain by Glance conglomerate, the basal unit of the Bisbee basin (Bilodeau, 1979; Bilodeau et al., 1987) that in this area contains sparse basalt-andesite flows and rhyolitic tuff dated at 151–146 Ma (Krebs and Ruiz, 1987).

Mexican Sonora

Within northern Sonora, Anderson et al. (2005) divided the Jurassic rocks into four domains, and what follows comes from their work: From Nogales (Fig. 30) eastward are sparse Jurassic rocks of the Nogales-Cananea-Nacozari domain, which overlie, or are inferred to overlie, Precambrian crystalline basement and comprise 174 Ma rhyolitic-dacitic lavas and tuffs intercalated with various siliciclastic sedimentary rocks, in places cut by hypabyssal porphyries (174 Ma) and a 177–173 Ma pluton. Located just to the west of the Nogales-Cananea-Nacozari domain is the Southern Papago domain, which is a continuation of the Topawa–Cobre Ridge–Wrightson and Kitt Peak–Trigo Peak rocks to the north, and was first recognized by Haxel et al., (1984) as an area of voluminous Jurassic magmatism without exposed Precambrian basement. It contains thick sequences of rhyolitic tuff, lavas, hypabyssal porphyries, and plutons dated in the 176 to 166 Ma range. The Mojave-Sonora domain, which is a NW-SE striking band of strongly deformed Jurassic rocks, including recumbent folds and thrust faults that sit north of the steeply dipping mylonites, is interpreted to represent the Mojave-Sonora megashear. Movement on the megashear deformed the rocks of the more easterly domains after ~160 Ma yet before the earliest Cretaceous. The Caborca domain comprises sedimentary successions sitting upon 1.8–1.7 Ga crystalline basement, with only a few possible, but as yet undated, outcrops of Jurassic volcanic rocks, but near the Gulf of California are three locations with deformed and metamorphosed rocks dated as 164, 153, and 141 Ma. Just to the southeast are Upper Jurassic rocks of the Cucurpe Formation, which unconformably overlies ~700 m of Middle Jurassic arc assemblages containing a dacitic ash-flow tuff dated to be 168 Ma, and comprises marine sedimentary rocks with thin siliceous tuff beds dated as 152–150 Ma (Mauel et al., 2011).

Striking southeastward from near Durango to the Gulf of Mexico is an area of Jurassic metavolcanic rocks (Fig. 5), known generally as the Nazas volcanics (Barboza-Gudiño et al., 1998, 2008; Bartolini, 1998; Bartolini et al., 2003; Godínez-Urban et al., 2011). The volcanic succession, hypothesized to be a possible continuation of the northern Sonoran rocks separated by the Mojave-Sonora megashear (James et al., 1993; Blickwede, 2001; Barboza-Gudiño et al., 2008), is unconformably overlain by Oxfordian sedimentary rocks, and includes up to 3 km of dominantly pyroclastic rocks intercalated with lesser amounts of andesitic-dacitic lavas, ash-fall beds, and volcaniclastic rocks cut by a variably deformed rhyolitic porphyry dated at 158 ± 4 Ma. Others, based largely on paleomagnetism (Molina-Garza and Geissman, 1999), or detrital zircons with similar ages to Pan-African rocks (Godínez-Urban et al., 2011), have argued against this possibility.

Baja California

Jurassic rocks in southwestern California and the Baja Peninsula of Mexico aren't very well known and are relatively scarce, but sufficient remnants exist to suggest that they were an important component of the area prior to Cretaceous magmatism. East and northeast of San Diego, Jurassic plutons in the 170 to 160 Ma range, both peraluminous and metaluminous, but generally gneissic, intrude mostly Late Triassic–Jurassic metasedimentary rocks, now steeply dipping, in a belt 45 km wide by at least 150 km long (Girty et al., 1993b; Shaw et al., 2003). Farther south within the Sierra San Pedro Martir of northern Baja, Schmidt and Paterson (2002) mapped and dated biotite orthogneiss at 164 Ma in two different locations within an extensive tract of orthogneiss in the central and eastern Peninsular Ranges batholith.

About halfway down the peninsula, near the state line at El Arco, ~6 km of exposed greenschist facies Jurassic rocks, including andesitic lava flows, breccias, pyroclastic rocks, and a spectrum of isoclinally folded sedimentary rocks, are cut by a mineralized granodiorite porphyry dated at 165 ± 7 Ma (Valencia et al., 2006; Weber and Martínez, 2006). Additional Jurassic rocks crop out sporadically all the way down the peninsula (D. Kimbrough, 2010, personal commun.).

San Gabriel–Caborca Block

Rocks that lie to the south of the reconstructed band of Pelona-Orocopia schists (Fig. 8) might represent a separate terrane, or group of terranes, because the schists, as discussed earlier, are generally interpreted to represent part of an exhumed subduction complex, and therefore separate this block from rocks to the north. The region is much broken by Cenozoic faults, but several reconstructions help to define the southern area as one coherent block prior to faulting (Powell, 1993; Nourse, 2002). It thus includes the San Gabriel, Orocopia, and Chocolate mountains and regions to the south (Figs. 8 and 20).

Sitting structurally above the Pelona schist (Ehlig, 1982) in the San Gabriel Mountains are polydeformed and metamorphosed 1.7–1.6 Ga Precambrian gneisses intruded by a 1.2 Ga anorthosite-syenite-gabbro complex, the Early Triassic Lowe granodiorite, and variety of Cretaceous plutons (Dibblee, 1968, 1982; Ehlig, 1975, 1981; May and Walker, 1989; Powell, 1993; Nourse, 2002). Hypersthene-bearing tonalitic gneisses gave U-Pb zircon ages of 88–84 Ma, whereas undeformed biotite granite yielded a U-Pb age of 78 ± 8 Ma (May and Walker, 1989). Most biotite K-Ar ages from Cretaceous intrusive units fall between 78 and 57 Ma (Miller and Morton, 1980; Mahaffie and Dokka, 1986). These ages combine to suggest a major deformational event at ~80 Ma followed by rapid exhumation.

To the east around the Sonora-Arizona frontier in the Sierra los Alacranes and Sierra El Choclo Duro (Fig. 30) is a block of 1.7–1.6 Ga Precambrian gneisses, amphibolite-grade metasedimentary rocks cut by 1.45 Ga porphyritic granite and several Late Cretaceous quartz dioritic, granodioritic, and granitic plutons (Nourse et al., 2005). Those authors suggest that these rocks continue southward and correlate with 1.78–1.69 Ga basement within the Caborca block (Iriondo, 2001; Premo et al., 2003; Anderson and Silver, 2005), but are different in age of deformation from rocks in the Mazatzal province farther north. The Caborca block is also known to contain a number of anorthositic complexes dated at ~1100 Ma (Espinoza et al., 2003). During the Early Triassic the Caborca block apparently lay at ~21° ± 4°N with a paleopole rotated clockwise relative to the Early Triassic reference pole, suggesting early sinistral displacement with respect to North America followed by younger dextral movement (Steiner et al., 2005). Using the Early Triassic paleopositions for North America of Kent and Irving (2010), the determined paleolatitude corresponds to a position off the present-day Pacific Northwest.

Guerrero and Other Mexican Terranes

Mexico is composed of many terranes, most of which appear to be exotic with respect to North America (Campa and Coney, 1983). The large Guerrero composite terrane, which occupies a huge chunk of central and western mainland Mexico comprises five separate terranes, Teloloapan, Guanajuato, Arcelia, Tahue, and Zihuatanejo (Fig. 5), all of which have successions of uppermost Jurassic to Cretaceous volcanic rocks, and lie to the west of Oaxaquia and Mixteca terranes (Centeno-García et al., 2008).

Oaxaquia (Ortega-Gutiérrez et al., 1995) contains a Precambrian Grenville-age basement of meta-anorthosite, orthogneiss, and charnockite (Ruiz et al., 1988; Keppie et al., 2001, 2003; Solari et al., 2003; Ortega-Obregón et al., 2003) unconformably overlain by Paleozoic sedimentary rocks with dominantly exotic, non-Laurentian biofacies and Permian volcanic and related sedimentary rocks that were overlain along their western margin by thick successions of turbidites—all strongly deformed prior to eruption and deposition of Kimmeridgian volcanic rocks (Centeno-García and Silva-Romo, 1997; Jones et al., 1995; Barboza-Gudiño et al., 2004). Although the Oaxaca terrane may have originated in eastern North America, based on its NW orientation nearly perpendicular to North American Grenville, its exotic fauna, and extensive Paleozoic deformation, Oaxaquia probably spent most of the Paleozoic some distance from North America (Ortega-Gutiérrez et al., 1995). The western boundary of the terrane appears to mark a pronounced step in crustal thickness from 40 km to 20 km to the west (Delgado-Argote et al., 1992). A small terrane of Permian MORB-like magmatic rocks, known as the Juchatengo terrane, lies along the southwest side of Oaxaquia and is interpreted to represent a short-lived Paleozoic back-arc basin developed prior to emplacement of 290–219 Ma plutons (Grajales-Nishimura et al., 1999).

Mixteca terrane (Fig. 5) contains pre-Mississippian polydeformed metamorphic rocks (Ruiz et al., 1988; Yañez et al., 1991; Ortega-Gutiérrez et al., 1999) unconformably overlain by Permian sedimentary rocks and Middle Jurassic volcanic and sedimentary rocks (García-Díaz et al., 2004). Along its western contact with the composite Guerrero terrane is a sequence of deformed and metamorphosed volcanic and sedimentary rocks dated by U-Pb on zircons as ~130 Ma (Campa Uranga and Iriondo, 2003, 2004) and containing detrital zircons broadly similar to those of the metamorphic basement (Talavera-Mendoza et al., 2007). The metamorphic rocks are unconformably overlain by Albian–Turonian platformal carbonates and an eastward thickening Turonian–Paleocene foredeep that developed contemporaneously with eastward-vergent thrusting (Cerca et al., 2010). Both Oaxaquia and Mixteca terrane were deformed in the Late Cretaceous–early Tertiary in a mostly eastward-vergent fold-thrust belt (Suter, 1984, 1987; Hennings, 1994). A deep trough (Fig. 5) formed to the east in front of the thrust belt and is known as the Tampico-Misantla foredeep (Busch and Gavela, 1978).

The Arcelia, Guanajuato, and Teloloapan terranes (Fig. 5) are tectonic slices containing thick sections of Cretaceous oceanic and arc-like rocks that were thrust eastward over the Oaxaquia and Mixteca terranes (Centeno-García et al., 2008). The Teloloapan terrane—overthrust on the west by the Arcelia terrane—contains ~3000 m of basaltic-andesitic lavas and breccias interbedded with Lower Cretaceous siliciclastic rocks in the lower part and Aptian limestone in the upper, all overlain by Albian–Turonian marine sedimentary rocks (Monod et al., 2000; Talavera-Mendoza et al., 2007; Cerca et al., 2010). Arcelia terrane comprises 2 km of Albian–Cenomanian, tholeiitic pillow lavas and breccias, interbedded and overlain by radiolarian chert and shale with small bodies of serpentinite, collectively interpreted to represent part of an oceanic arc terrane (Delgado-Argote et al., 1992; Elías-Herrera et al., 2000; Mendoza and Suastegui, 2000). Eastward thrusting in the terrane appears to be coincident with deposition of a thick section of mainly Coniacian–Campanian red beds interbedded with 84 Ma lava and a conglomerate holding an andesite clast dated at 74 Ma (Martini et al., 2009; Martini and Ferrari, 2011).

The Guanajuato terrane (Fig. 5) consists of thrust slivers of gabbro, tonalite, and ultramafic rocks thrust north-northeastward over an isoclinally folded, bimodal volcanic suite and flysch package of Lower Cretaceous age, prior to the deposition of unconformably overlying Aptian–Albian carbonates (Lapierre et al., 1992; Ortiz-Hernandez et al., 2003). At San Miguel de Allende, upper Aptian, calc-alkaline basalts and basaltic andesites interbedded with pelagic sedimentary rocks sit structurally on the western margin of Oaxaquia (Ortiz-Hernández et al., 2002).

The two westernmost parts of the Guerrero composite terrane, the Tahue and Zihuatanejo terranes (Fig. 5), apparently shared a common Lower Cretaceous history. The Tahue terrane is the northwesternmost of the two and includes a deformed and metamorphosed Ordovician arc terrane unconformably overlain by deformed Pennsylvanian–Permian turbidites (Centeno-García, 2005). These rocks are overlain, perhaps in places tectonically, by Cretaceous arc volcanic rocks and intruded by related mafic-intermediate plutons (Ortega-Gutiérrez et al., 1979; Henry and Fredrikson, 1987; Centeno-García et al., 2008). The Zihuatanejo terrane comprises Upper Jurassic siliceous lavas and 163–155 Ma plutons that sit atop and intrude a Triassic accretionary complex of flyschoid mélange holding blocks of pillow lava, chert, serpentinite, and limestone, all unconformably overlain by deformed Lower Cretaceous mafic-siliceous lavas, sedimentary rocks, and cut by 105 Ma and younger plutons (Centeno-García et al., 2008, 2011).

Several scientists have suggested that the Mesozoic volcano-sedimentary rocks of the Guerrero composite terrane were deposited on attenuated North American cratonic crust (Cerca et al., 2007; Centeno-García et al., 2008; Martini et al., 2009), but as far as I am aware there is no evidence of such old basement either in outcrop or isotopes. The extensive tracts of Late Jurassic–Early Cretaceous volcanic successions are more typical of rocks within the Rubian superterrane than western North America and likely represent a composite arc (Tardy et al., 1994). Furthermore, detrital zircons recently collected from the Zihuataneyo terrane are dominated by 110–105 Ma zircons with smaller age-distribution peaks of ~1000 Ma and 560–590 Ma (Centeno-García et al., 2011), which suggest closer ties to Rubia than Laurentia.

To the south of the Zihuatanejo terrane, along the coast in the Sierra Madre del Sur lies the Xolapa complex (Fig. 5), which is a 50-km-wide by 650-km-long, high-grade migmatitic orthogneiss terrane, at least in part derived from Grenville-age crust, cut by deformed plutons dated as 160–136 Ma and massive bodies dated to be 66–46 Ma (Campa and Coney, 1983; Herrmann et al., 1994; Ducea et al., 2004). The northern boundary of the complex appears to be marked by extensive mylonites indicating top to the NW and evidence of younger, Eocene southwest-directed thrusting and strike-slip motion (Nieto-Samaniego et al., 2006; Solari et al., 2007).

East of the Xolapa complex and the Mixteca terrane is the Mayan terrane (Fig. 5), which contains a southwest-facing Cretaceous platformal carbonate succession (Cordoba platform) that sits atop Jurassic continental strata and in the Coniacian–Maastrichtian Zongolica fold-thrust belt were overthrust along northeastward-verging thrust faults by rocks of a Kimmeridgian–Eocene basin comprised of lower submarine basalt, sandstone, shale, and conglomerate passing upward into limestone, sandstone, and mudstone topped by flysch (Nieto-Samaniego et al., 2006). The southern part of the Mayan block contains the Chiapas massif (Fig. 5), an elongate terrane comprising Paleozoic plutons metamorphosed during the Permian and cut by Late Cretaceous–early Tertiary plutons (de Cserna, 1989; Burkart, 1994) and the Maya Mountains of Belize, which contain metamorphosed and deformed Pennsylvanian–Permian sediments sitting unconformably on Late Silurian plutonic basement (Steiner and Walker, 1996). Along the western boundary of the terrane is the narrow Cuicateco terrane (Fig. 5), which consists of dominantly Maastrichtian schists, greenstones, gabbros, and serpentinites of ophiolitic character that were thrust eastward over red beds of the Maya terrane during the latest Cretaceous–Paleocene (Pérez-Gutiérrez et al., 2009).

At the southernmost part of North America, just north of the Guatamalan suture complex, a west-facing Cretaceous platform of a passive margin that sat on Mesoproterozoic–Triassic basement of the Maya block was drowned during the uppermost Campanian, buried by orogenic flysch during the Maastrichtian–Danian (Fourcade et al., 1994), and overthrust by ultramafic nappes. Rocks of the lower-plate crystalline basement in the Chuacús complex were metamorphosed to eclogite at 76 Ma, which implies that part of the North America margin was subducted to greater than 60 km depth at about that time, and exhumed to amphibolite grade a million years later (Martens et al., 2012), presumably after slab failure. This deformation and metamorphism are generally attributed to attempted subduction of North America to the west beneath an arc terrane generally known as the Great Arc of the Caribbean (Pindell and Dewey, 1982; Burke, 1988; Pindell et al., 1988; Donnelly et al., 1990; Rosenfeld, 1993; Burkart, 1994).

The Great Arc of the Caribbean

Ever since Wilson (1966) suggested that the Antillean and Scotian arcs came out of the Pacific along transform faults, various authors have suggested that they were part of more extensive arcs located within the Pacific basin during the Mesozoic (Moores, 1970; Malfait and Dinkelman, 1972; Burke, 1988; Pindell, 1990; Pindell and Kennan, 2009; Wright and Wyld, 2011). Burke (1988) coined the term "Great Arc of the Caribbean" to call attention to his idea that the Antillean arc was just one part of a more extensive Mesozoic arc that collided with rocks in Central America, Mexico, and northern South America during the Late Cretaceous. Although some workers (Pindell et al., 2005; García-Casco et al., 2008) have recently suggested that the west-dipping subduction zone already existed by 120–115 Ma, central to Burke's (1988) idea was that the arc was originally constructed above a easterly-dipping subduction zone, but that it flipped to westerly-dipping after it collided at 85–80 Ma with an oceanic plateau formed ~90 Ma above the Galápagos hotspot (Vallejo et al., 2006). Both the arc and the oceanic plateau entered the Atlantic, where they occur today in and around the Caribbean Sea. Remnants of the oceanic plateau within the Caribbean were studied by Sinton et al. (1997, 1998) and Kerr et al. (2003), who also studied fragments of the Cretaceous arc. In the Honduras-Nicaragua border area, a 350-km-long segment of the arc, termed locally the Siuna terrane, was emplaced over the northward-vergent Colon fold-thrust belt of the Chortis block at ~80–75 Ma (Venable, 1994; Rogers et al., 2007).

The Great Arc and its trailing oceanic plateau also collided with the northwestern part of South America, and it apparently arrived in Ecuador during the Late Campanian at ~75 Ma (Jaillard et al., 2004; Luzieux et al., 2006). In Venezuela and Colombia, many strike-slip faults sliced the allochthons and transported them north and eastward during and after the collision (Altamira-Areyán, 2009).

Peninsular Ranges Batholith

The Cretaceous Peninsular Ranges batholith (Figs. 5 and 8) extends 800 km from southern California at least halfway down the Baja Peninsula of Mexico (Gastil et al., 1975) and comprises two petrographically, spatially, and temporally distinct magmatic suites, each with different basement rocks (Gastil, 1975; Gromet and Silver, 1987; Silver and Chappell, 1988; Walawender et al., 1990; Gastil et al., 1990). The older plutons, which are deformed, range in age from ~140 to 105 Ma; include many ring complexes; occur in the west; intrude crudely coeval volcanic and volcaniclastic rocks of the 127–116 Ma Santiago Peak volcanics in the north and 120–110 Ma Alisitos group at shallow-moderate depths in Baja; and range compositionally from gabbro to monzogranite (Gastil, 1975; Johnson et al., 2002; Wetmore et al., 2003; Busby et al., 2006; Kimbrough et al., 2001; Gray et al., 2002). The eastern group, known as the La Posta suite, ranges in age from 99 to 92 Ma, and comprises inwardly zoned bodies of hornblende-bearing tonalite to muscovite-biotite granodiorite and monzogranite (Clinkenbeard and Walawender, 1989; Walawender et al., 1990) emplaced at depths of 5–20 km into upper greenschist to amphibolite grade wall rocks that are in many places migmatitic (Gastil et al., 1975; Todd et al., 1988, 2003; Grove, 1993; Rothstein, 1997, 2003). Prior to emplacement of the La Posta suite but after the emplacement of the western plutonic series, the plutons and their respective basements were juxtaposed along a group of thrust faults, which in the Sierra San Pedro Martir form a doubly-vergent fan structure, and led some researchers to hypothesize various collisional models based on an exotic Alisitos arc (Johnson et al., 1999a; Schmidt and Paterson, 2002; Schmidt et al., 2002; Wetmore et al., 2002; Alsleben et al., 2008). Other workers (Gastil, 1993; Busby et al., 1998) argued for back-arc extension within the arc and the creation of a "fringing arc" or, based on continuous plutonism in the U.S. sector, saw no need for any break at all (Todd et al., 2003). However, paleomagnetic data suggest that the western belt was far traveled (11° ± 4° northward) with respect to North American paleopoles, whereas the La Posta suite yielded proximate poles (Symons et al., 2003).

The La Posta plutons were emplaced from 99 to 92 Ma immediately following this deformational event and, while situated mostly to the east of the suture zone, they locally cut across it and intrude rocks of the Alisitos arc to the west (Silver and Chappell, 1988; Kimbrough et al., 2001). Exhumation of the La Posta plutons and their wall rocks occurred in two discrete phases: (1) a Cenomanian-Turonian phase; and (2) a Late Cretaceous Campanian-Maastrichtian phase (Grove, 1993; Lovera et al., 1999; Kimbrough et al., 2001; Grove et al., 2003a). During the earlier phase, which likely overlapped with emplacement of the plutons, rocks at depths of 10 km were brought much closer to the surface by detachment faulting and collapse coincident with a pulse of early Cenomanian to Turonian coarse clastic sedimentation in basins, located to the west and containing 100–90 Ma detrital zircons (George and Dokka, 1994; Lovera et al., 1999; Kimbrough et al., 2001). Based on their age, Grove et al. (2003) related the younger cooling ages to the Laramide event.

Cretaceous Magmatism in the Mojave and Sonoran Deserts

In the Mojave region (Fig. 21), magmatism in the 125 to 80 Ma range is generally scarce, but many plutons remain to be dated. In the Clark–Mescal Range–Ivanpah mountains, located near the California-Arizona border south of Las Vegas, 147–142 Ma plutons were deformed and transported on thrust faults, in part over the 100 Ma Delfonte volcanics, all possibly prior to the emplacement of phases of the Teutonia batholith at 93 Ma (Fleck et al., 1994; Walker et al., 1995; Beckerman et al., 1982).

Latest Cretaceous–early Tertiary plutons coincide with the region of thrusting formed during the 80–75 Ma deformation characteristic of the Sonora block. A well-studied example within eastern California is the Old Woman–Piute Range batholith (Fig. 21), which postdates peak metamorphism and deformation; comprises metaluminous and peraluminous granites dated at 71 ± 1 Ma by U-Pb on zircons; and was unroofed from mid-crustal levels during and shortly after emplacement, as evidenced by $^{40}Ar/^{39}Ar$ ages on hornblende of 73 ± 2 Ma and 70 ± 2 Ma from biotite (Foster et al., 1989; Miller et al., 1990). Other similar plutonic complexes in the immediate area include the ~70 Ma Chemehuevi Mountains plutonic suite (Fig. 21), which is a compositionally zoned complex of biotite granite and garnet–two mica granite (John and Wooden, 1990), and the 66.5 ± 2.5 Ma Ireteba pluton, a garnet–two mica peraluminous granite with adakitic-type concentrations of Sr, Eu, and heavy rare earth elements (Kapp et al., 2002). Within the Whipple Mountains (Fig. 21) metamorphic core complex, the 73 Ma Axtel quartz diorite also has adakitic characteristics (Anderson and Cullers, 1990).

The ~70 Ma Coxcomb intrusive suite (Howard, 2002) includes a number of quartz monzodioritic, granodioritic, and granitic plutons exposed within the Kilbeck Hills and the Coxcomb, Sheephole, Calumet, and Bullion Mountains (Fig. 21). They tend to be more quartz-rich than the older predeformational, 80 Ma plutons (John, 1981) and show some adakitic tendencies (R. Economos, 2011, personal commun.).

Exposed within the eastern Transverse Ranges (Fig. 21) is an oblique cross section representing as many as 22 km of paleodepth (Barth et al., 2008b). There, more than twenty 82–73 Ma plutons, including a complex of northwest-striking, moderately northeastward-dipping sheeted bodies that may extend for 150 km along strike (Powell, 1993), are dominantly tonalite—granodiorite and biotite + muscovite ± garnet granite, and are only weakly deformed (Needy et al., 2009).

Late Cretaceous–early Tertiary magmatism also occurs in southwest Texas, just to the northwest of the Big Bend in the Rio Grande, where the 64 Ma Red Hills pluton hosts a copper-molybdenum porphyry system (Gilmer et al., 2003). Thus, within the southwestern United States, the overall trend of magmatism of this age is nearly E-W from California to West Texas.

The Sonora segment, south of the Colorado Plateau, contains another magmatic belt that was interpreted to represent slab break-off magmatism (Hildebrand, 2009). These include the widespread Laramide intrusions of southern Arizona and New Mexico, as well as a linear belt of 76–55 Ma plutons that continue southward through much of western Mexico (Anderson et al., 1980; Damon et al., 1983; Zimmermann et al., 1988; Titley and Anthony, 1989; Barton et al., 1995; McDowell et al., 2001; Henry et al., 2003; Valencia-Moreno et al., 2006, 2007; Ramos-Velázquez et al., 2008; González-León et al., 2010). Recent U-Pb zircon dating in northern Sonora revealed that the Tarahumara Formation, a greater than 2-km-thick, mixed clastic-volcanic unit that unconformably overlies thrusted and folded rocks ranging in age from Proterozoic to Late Cretaceous, was deposited and erupted between 76 and 70 Ma, and was intruded by plutons dated between 70 and 50 Ma (McDowell et al., 2001; González-León et al., 2010, 2011).

Subduction and Fore-Arc Complexes

Within the Sonoran sector, equivalents to the Franciscan complex are well exposed on Catalina Island (Figs. 5 and 28), where a stack of shallowly-dipping thrust sheets contain rocks that are progressively younger and lower grade structurally downwards from amphibolite to blueschist (Platt, 1975, 1976). The structurally highest unit is dominantly hydrated and metasomatized amphibolite, of which the upper part consists of a mélange with blocks of metabasite in serpentinite and ultramafic paragneiss, and a lower part of amphibolite gneiss and pelitic schist (Sorensen, 1988). Down structure, Platt (1975) identified a greenschist unit comprising pelitic schist intercalated with mafic schist, probably metavolcanic rocks, and sparse serpentinite bodies. Subsequently, Grove and Bebout (1995) subdivided the unit into epidote-amphibolite, epidote-blueschist, and lawsonite-bearing blueschist units. The structurally lowest unit is a diverse amalgamation of blueschist-grade metagraywacke, massive metabasalt, mafic tuff or sandstone, metachert, carbonaceous pelitic schist, and conglomerate containing clasts of metachert, metavolcanic rocks, ultramafic pebbles, now tremolite-fuchsite rock, and metadiorite-metagabbro (Platt, 1975). Grove et al. (2008a) presented U-Pb analyses of detrital zircons from rocks of the thrust stack, which demonstrated that the upper amphibolite unit was no older than 122 ± 3 Ma; the epidote amphibolite unit no older than 113 ± 3 Ma; the epidote blueschist unit, 100 ± 3 Ma, the lawsonite blueschist unit, 97 ± 3 Ma, and garnetiferous blueschist blocks in mélange, 135 Ma.

Even farther south, along the western side of the Baja Peninsula of Mexico, Franciscan-type rocks occur in two areas: the Vizcaino Peninsula–Cedros Island area and on islands along the western side of Bahia Magdalena (Fig. 5). There, a sequence of volcanic and sedimentary rocks, generally interpreted to represent a magmatic arc, active from at least 166 ± 3 Ma until 160 Ma and again during the Cretaceous, sits atop a Late Triassic and mid-Jurassic suprasubduction ophiolitic basement, which in turn, sits structurally above a blueschist-bearing accretionary complex and is separated from it by a serpentinite mélange containing high-grade blocks (Kimbrough, 1985; Moore, 1985, 1986; Kimbrough and Moore, 2003; Sedlock, 2003). The blocks in the mélange are typically eclogite and amphibolite with ages in the 170–160 Ma range, and blueschist blocks ranging in age from 115 to 95 Ma (Baldwin and Harrison, 1989, 1992). The upper contacts of the blueschist belts are interpreted as normal faults and reflect their probable Late Cretaceous–Paleogene exhumation (Sedlock, 1996, 1999).

Two small areas of likely correlative fore-arc rocks occur along the western margin of the Baja Peninsula in the Vizcaino Peninsula–Cedros Island area and farther south on islands along the western side of Magdalena and Almejas bays (Fig. 5). In the northern areas, perhaps 10 km of upper Albian–Cenomanian to Coniacian–Maastrichtian siliciclastic turbidites sit unconformably on older arc-ophiolite-accretionary complex rocks (Minch et al., 1976; Boles, 1986; Busby-Spera and Boles, 1986; Sedlock, 1993). To the south, lesser quantities of similar rocks occur on the islands of Santa Margarita and Magdalena (Rangin, 1978; Blake et al., 1984b; Sedlock, 1993). Paleomagnetic data from Cedros Island and the Vizcaino Peninsula suggest that rocks there were deposited at 20° ± 6° and at 16° ± 7°, respectively (Smith and Busby-Spera, 1993).

Late Cretaceous Deformation and Metamorphism

Although it is much broken and separated by Cenozoic strike-slip and normal faults, there is a broad swarth of Cretaceous deformation and metamorphism that extends from Sonora westward across the Mojave Desert into the Transverse Ranges and the Salinian block. The best studied area is located within west central Arizona and eastern California, where an arcuate region of highly tectonized and metamorphosed crystalline basement and overlying metasedimentary veneer known as the Maria fold and thrust belt (Fig. 8)—with similar lithologies and stratigraphy to those of the North American platform (Stone et al., 1983), but not necessarily connected to it, or in its current location, during the Cretaceous—were metamorphosed to amphibolite grade, locally thinned to 1% of their original thickness, and recumbently folded between 84 Ma and 73 Ma (Brown, 1980; Hamilton, 1982; Hoisch et al., 1988; Fletcher and Karlstrom, 1990; Spencer and Reynolds, 1990; Tosdal, 1990; Knapp and Heizler, 1990; Howard et al., 1997; Boettcher et al., 2002; Barth et al., 2004; Salem, 2009). The belt (Fig. 8) appears to be truncated to the north by the Phoenix fault (Hildebrand, 2009). Just to the south, and overthrust on both the north and south, is the linear thrust-bounded band of the McCoy Mountains Formation (Tosdal, 1990; Tosdal and Stone, 1994; Barth et al., 2004).

The McCoy Mountains Formation (Fig. 8), which outcrops in faulted ranges in southern California and Arizona, has remarkably similar ages to the Great Valley group, but is more commonly considered to be of back-arc (Barth et al., 2004; Jacobson et al., 2011) or rift provenance (Spencer et al., 2011). The formation is a 7-km-thick sedimentary succession, deposited in alluvial-fan, fluvial, and shallow lacustrine settings, that outcrops in an east-west strip across the California-Arizona border just to the north of the band of Orocopia schist (Pelka, 1973; Harding and Coney, 1985; Tosdal and Stone, 1994; Jacobson et al., 2011). Rocks of the formation sit unconformably upon Middle-Late Jurassic volcanic and volcaniclastic rocks ranging in age from 174 to 155 Ma (Fackler-Adams et al., 1997; Barth et al., 2004) with an unknown upper limit, as the rocks were overthrust on both the northern and southern margins (Tosdal, 1990). A detailed U-Pb detrital zircon study showed that the basal sandstone member could be as old as Callovian as it had zircons no younger than 165 Ma; whereas samples just above the lower sandstone yielded zircons as young as 109 Ma, with 84 Ma zircons found near the exposed top, which was cut by a granitic pluton dated at 73.5 ± 1.3 Ma (Barth et al., 2004).

South of Tucson, and extending westward to Ajo and Organ Pipe National Monument (Fig. 30), is a band of Late Cretaceous–early Tertiary thrusts that place Precambrian crystalline basement over metamorphosed Jurassic and Cretaceous sedimentary, volcanic, and plutonic rocks with west-dipping foliations and SW-plunging lineations (Haxel et al., 1984). The thrusts are cut by metaluminous plutons in the age range 74 to 64 Ma and peraluminous bodies in the range 58 to 53 Ma (D.M. Miller et al., 1992). Hildebrand (2009) suggested that the plutons in the age range 74 to 64 Ma might be part of a slab break-off suite that continues southward through western Mexico, whereas the younger group is part of the postcollisional arc that also extends down through western Mexico.

To the southeast, rocks of the Cretaceous Bisbee basin were folded about NW-trending axes and broken by reverse faults (Davis, 1979). Folds of this age, along with reverse and thrust faults, carry along strike into southwestern New Mexico and throughout the Chihuahua trough (Corbitt and Woodward, 1970, 1973; Gries and Haenggi, 1970; Haenggi and Gries, 1970; Brown and Clemons, 1983; Lehman, 1991; Hennings, 1994; Clemens, 1998; Hildebrand et al., 2008).

A fold-thrust belt extends from northern Mexico to Guatemala. It forms the eastern margin of Oaxaquia (Fig. 5) and, as described earlier, is also well exposed in Mexican Chiapas, Guatemala, Honduras, and Colombia.

To the west within the north-central Mojave region of California, and exposed within the lower plate of the Central Mojave metamorphic core complex (Fig. 21), Fletcher et al. (2002) studied complexly deformed 105–85 Ma Cretaceous migmatites and plutons. They suggested that a leucogranite body with reversely concordant monazites (86–84 Ma) might be syn-deformational, but as it contains the two main deformational fabrics and is clearly folded, it probably predates the deformation.

Farther to the west at the NW end of the San Bernardino Mountains (Fig. 21), the Cajon Pass drill hole penetrated a number of gently foliated, 81–75 Ma plutons, typically separated by shallowly inclined to horizontal faults (Silver and James, 1988; Silver et al., 1988). The Transverse Ranges also contain a number of terranes or fault slices of apparent Cretaceous migmatites and mylonitic thrust zones, including the Cucamonga and San Antonio slices in the San Gabriel Mountains (Fig. 21), where an 84 Ma tonalitic body is deformed but a 78 ± 8 Ma biotite granite that intrudes it is undeformed (May, 1989; May and Walker, 1989; Powell, 1993). A regional potassium-argon study of the eastern Transverse Ranges and southern Mojave Desert area showed that most rocks, including Precambrian gneisses, yield biotite ages in the range 70 to 57 Ma (Miller and Morton, 1980) suggesting cooling during this period. Rocks of the Peninsular Ranges to the south were apparently also affected by this deformational event in that they show evidence of rapid exhumation between 80 Ma and 68 Ma (Grove et al., 2003a).

CANADIAN SECTOR

Passive Margin

As discussed early in this publication, within Canada the North American west-facing passive margin formed in the latest Proterozoic (Bond and Kominz, 1984) and consists of a shallow water platformal sequence consisting of ~1 km of lower Paleozoic carbonate rocks sitting atop widespread clastic rocks of the Gog Group (Fig. 9), which locally sits on uppermost Neoproterozoic Miette Formation (Wolberg, 1986; Aitken, 1989). The lower Paleozoic platform margin is well exposed in the Main Ranges and is known as the Kicking Horse Rim (Cook, 1970; Aitken, 1971). Fine-grained, deep-water units of obscure provenance occur to the west of the reefal rim and include the famous Burgess shale (Fig. 10). During the Devonian the continental interior was broadly arched such that Upper Devonian platformal carbonates unconformably overlap Silurian, Ordovician, and Upper Cambrian rocks to the east (Price, 1981). Carboniferous strata comprise platform carbonates and sandstone dominated siliciclastic facies (Beauchamp et al., 1986; Mamet et al., 1986). Permian strata are absent from most of the margin, apparently due to widespread younger erosion (Henderson, 1989). Other than in the extreme northwest, Triassic rocks of the platform occur mainly east of the Sweetgrass arch, a northeast-trending broad uplift near the international border (Fig. 12), and in the Rocky Mountain front ranges, where they are known as the Spray River Group (Porter et al., 1982). Overall, Triassic rocks are marine to marginal-marine carbonate and siliciclastic rocks with minor evaporite (Gibson and Barclay, 1989; Poulton, 1989).

Foredeep

The Western Canada basin contains rocks of the Cordilleran foredeep. The foredeep is generally considered to have formed sometime in the Middle Jurassic, at ~170–160 Ma, with the deposition of sedimentary rocks of the Fernie Formation, called Cycle 1 by Leckie and Smith (1992), but there are problems with this interpretation. The lowermost part of the formation—which sits unconformably upon progressively older platformal sedimentary rocks to the east and northeast—contains phosphorite units, with up to 30% P_2O_5, and which typically form along the eastern margins of open oceans as cold, nutrient-rich waters upwell from cold, southward-flowing currents, often due to trade winds or possibly due to deflections of currents (Poulton and Aitken, 1989; Parrish and Curtis, 1982). Disconformably overlying this phosphatic unit, and perhaps signifying the end of open ocean access, are black shales, typically radioactive, green sands, and fine-grained siliciclastic rocks, with bentonites entirely absent from their lower parts, and which are overlain by coarser-grained siliciclastics, peat and coal beds of the Kootenay Group (Jansa, 1972; Stronach, 1984). These rocks were deposited by northerly paleoflow in a narrow, northwest-trending basin, whose westernmost exposed and thickest parts occur ~70 km east of the miogeoclinal platform edge (Hamblin and Walker, 1979; Price and Fermor, 1985). Gibson (1985) studied the petrography of the Kootenay Group and found that most, if not all, of the detritus could be accounted for by erosion of rocks of the miogeocline. Similarly, Ross et al. (2005) studied the Nd isotopes and detrital zircons of the Fernie and Kootenay and concluded that both are a good match for the Devonian through Triassic miogeocline. Given that plutonism, recumbent folding, and 4 kb of exhumation of rocks within Kootenay terrane, which was hypothesized to have collided with the margin at this time (Monger et al., 1982) and produced the subsidence and sedimentation within the Fernie-Kootenay, workers have had to conjure a drainage divide to explain the lack of metamorphic and plutonic debris within the Fernie-Kootenay basin (Ross et al., 2005; Evenchick et al., 2007). Thrust faults do not appear to cut the Fernie-Kootenay until the Santonian–Campanian (Larson et al., 2006), or ~80–90 Myr after initiation of sedimentation within the basin, which seems a particularly long time from basin inception to thrusting for a foredeep.

Sitting unconformably on the Fernie-Kootenay—and perhaps as many as 25 Myr younger—are coarse sediments and sandstones of the Cadomin Formation, which was deposited after 114 Ma, may be younger than 95 Ma (Leier and Gehrels, 2011), and contains gravels and conglomerates holding chert and quartzite clasts of miogeoclinal provenance (McLean, 1977; Leckie and Smith, 1992; Leckie and Cheel, 1997). As discussed in a much earlier section, Hildebrand (2009) suggested that these intra-platformal conglomerates, which occur over much of the Canadian and Great Basin sectors, formed when the passive margin was uplifted and exposed as it was transported over the outer swell to the trench and that foredeep sedimentation within the Cordillera started with sedimentation on top of the conglomerates as the margin was pulled down into the trench. The conglomerates and sandstones are overlain by a thick clastic wedge of siliciclastic sedimentary rocks contained within the Blairmore and Manville Groups, which contain volcanic and plutonic clasts, mid-Cretaceous detrital zircons, and with much higher ε_{Nd} than rocks of the Cadomin Formation (Ross et al., 2005), consistent with uplift and erosion of young volcano-plutonic material to the west.

As part of a regional unconformity, paleosols formed within, and an incised valley system was cut into, middle Albian(?) rocks of the Blairmore and Manville Groups (Leckie et al., 1989). Other than local conglomeratic and sandy zones, Cenomanian–Campanian shales, which contain evidence for periodic anoxic conditions and generally deep water, dominate the section (Leckie and Smith, 1992).

During the Campanian–Paleocene, another thick clastic wedge with sandstones, containing andesitic-dacitic volcanic fragments, low-grade metamorphic fragments and pelites, carbonates, and arenites, with northwestern paleocurrents, was deposited within the basin (Mack and Jerzykiewicz, 1989). This pulse of sedimentation appears to correlate with the main pulse of thrusting within the miogeoclinal platform (Larson et al., 2006), which ended at ~58 Ma with uplift and erosion (Price and Mountjoy, 1970; Sears, 2001; Ross et al., 2005). On the North American platform, the ~95 Ma Crowsnest volcanics (Peterson et al., 1997) are cut by thrusts. Also, the Bourgeau thrust, which is the westernmost of the major thrusts that cut the platformal rocks, overthrusts the Santonian–Campanian Wapiabi Formation (Price, in press; Larson et al., 2006). Following a major erosional period, Tertiary gravels and conglomerates, ranging in age from Eocene to Pliocene, and containing clasts up to 0.5 m across were deposited over Alberta and Saskatchewan (Leckie and Smith, 1992).

Canadian Terranes and Superterranes

Thirty years ago, Monger et al. (1982) published a provocative paper in which they presented compelling evidence that many of the terranes within the Canadian Cordillera could be lumped into two superterranes: Terrane I and Terrane II, which were each amalgamated offshore prior to their collisions with North America. One of their fundamental ideas was that farther outboard terranes were sequentially added to North America through time. Subsequently, these terranes became widely known as the Intermontane and Insular superterranes. As generally understood today, the Intermontane terrane comprises Kootenay, Slide Mountain, Quesnellia, Cache Creek, and Stikinia, whereas the Insular terrane consists of Alexander and Wrangellia (Fig. 5). They argued that Intermontane terrane was stitched together during the Triassic, whereas the Insular during the Cretaceous. They also considered the western margin of North America to be located at the western margin of the Omineca belt (Fig. 5). It is now recognized that Wrangellia and Alexander terranes were stitched in the Pennsylvanian at ~309 Ma (Gardner et al., 1988). In some terminology, the Alexander, Wrangellia, and Peninsular terranes of Alaska are lumped together as the Wrangellian superterrane (see Nokleberg et al., 2000).

Belt-Purcell-Windermere Supergroups

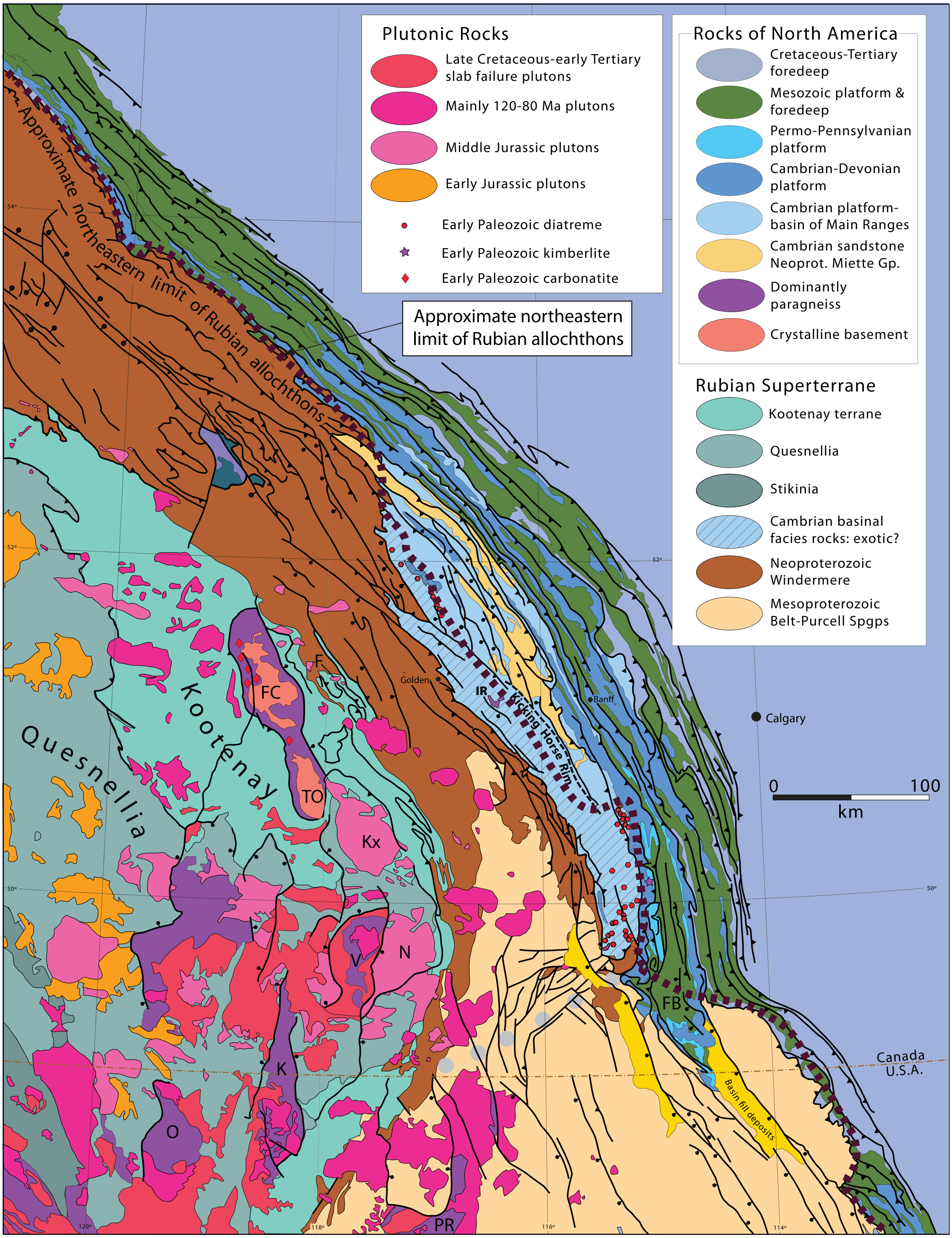
From Idaho north to British Columbia, sedimentary rocks of the Neoproterozoic Windermere group sit unconformably upon nearly 30 km of metasedimentary rocks and mafic sills of the Mesoproterozoic Belt–Purcell supergroups (Gabrielse, 1972)—all contained within huge thrust sheets (Figs. 5 and 8). Crystalline basement is not known from the thrusts sheets. In Montana and southernmost Canada, a huge slab of Belt Supergroup rocks, 70–110 km wide, ~450 km long, and as much as 14–16 km thick, was transported over and on top of the North American platform—even sitting well east upon Cretaceous shales of the foredeep basin (Figs. 32 and 33)—along the Lewis-Eldorado-Hoadley-Steinbach thrust system (Mudge and Earhart, 1980; Mudge, 1982; Sears, 1988, 2001; Cook and van der Velden, 1995; Fuentes et al., 2012). While rocks of the Belt-Purcell supergroups traditionally are considered to have been deposited on the western margin of North America more or less in their current location (Price and Sears, 2000), Hildebrand (2009) argued that they were exotic relative to North America. Subsequent studies that produced data supporting the exotic model include (1) the recognition of a suite of 664–486 Ma alkaline plutons intruding rocks of the Belt Supergroup and its miogeoclinal Paleozoic cover in central Idaho (Lund et al., 2010; Gillerman et al., 2008); (2) the realization that at least one of the plutons was likely deroofed during the Upper Cambrian (Link and Thomas, 2009; Link and Janecke, 2009), a peculiar occurrence for the outer part of a miogeocline; and (3) 1.2–1.0 Ga metamorphism and deformation found in Belt-Purcell metasedimentary rocks (Nesheim et al., 2009; Zirakparvar et al., 2010) are unknown in cratonic northwestern North America.

The Windermere supergroup (Fig. 32) includes a wide variety of coarse clastic rocks, in part glaciogenic, sparse volcanic rocks, shales, arkosic grits, and minor deep-water carbonate (Ross, 1991). Some workers (Stewart, 1972; Burchfiel and Davis, 1975; Lund, 2008) argued that rocks of the Windermere Supergroup, and equivalents, or even older rocks (Dehler et al., 2010), represent rift deposits on the western margin of North America, but as they don't contain extensive tracts of volcanic rocks and are some 85–100 Myr older (Lund et al., 2003; Fanning and Link, 2004) than the development of the passive margin, they wouldn't have retained enough heat to match the rate of early Paleozoic subsidence (Bond and Kominz, 1984; Devlin and Bond, 1988).

Selwyn Basin

A huge composite allochthon, some 700 km long and up to 200 km across, of Neoproterozoic to Paleozoic rocks of the Earn, Road River, and Hyland groups within Selwyn basin (Fig. 5) was thrust over the North American miogeocline along the Dawson and Broken Skull faults, and has no rocks in common with those of their footwall (Gordey and Anderson, 1993). The rocks of the allochthons comprise fine-grained sedimentary rocks, chert, limy turbidites, and graptolitic shale with alkaline basalts, barite beds, and sedimentary exhalative Ag-Pb-Zn ore deposits (Goodfellow et al., 1995; Mair et al., 2006). These features plus the presence of intermittent basalts throughout the section (Goodfellow et al., 1995; Cecile, 2010) are very similar to features within the Roberts Mountain allochthon as first noted by Turner et al. (1989) and are more typical of sedimentation, magmatism, and alteration within a restricted marginal basin, such as the South China Sea–Taiwan Strait (Teng and Lin, 2004; Koski and Hein, 2004), than a passive margin. Furthermore, recent work (McLeish et al., 2010; McLeish and Johnston, 2011) showed that rocks of the Kechika Trough, the southernmost part of the Selwyn basin, were recumbently folded during the Devonian, which makes it difficult to correlate these rocks with North American strata just to the east where similar age rocks do not show this deformation.

Figure 32. Geological sketch map illustrating the relationships within the Rocky Mountain thrust-fold belt of southern Canada. The Windermere-Belt-Purcell supergroups are interpreted here to constitute exotic allochthons sitting atop rocks of the North American passive margin. North American basement is exposed in Frenchman Cap (FC) and Thor-Odin dome (TO) of the Monashee complex, an erosional duplex structure, but was probably transported from well to the south during the Laramide event. Note the great numbers of small ultramafic intrusions that cut rocks interpreted here to be part of the Rubian superterrane just west of the Kicking Horse Rim, which is generally considered to represent part of the westward-facing North American Cambrian shelf edge. They probably mark the Rubian–North American suture as suggested in the general model of Burke et al. (2003) and more specifically by Johnston et al. (2003). To both the north and south the Windermere-Belt allochthons, derived from farther outboard, sit atop this zone, and were thrust over it such that their eastern limit marks the Rubian–North American suture. F—Fang stock; FB—Fernie basin; IR—Ice River complex; K—Kettle–Grand Forks dome; Kx—Kuskanax batholith; N—Nelson batholith; O—Okanagan dome; PR—Priest River complex; V—Valhalla dome. Geology from Wheeler and McFeely (1991) and Pell (1994).



Plutonic Rocks
Late Cretaceous-early Tertiary slab failure plutons
Mainly 120-80 Ma plutons
Middle Jurassic plutons
Early Jurassic plutons
Early Paleozoic diatreme
Early Paleozoic kimberlite
Early Paleozoic carbonatite
Rocks of North America
Cretaceous-Tertiary foredeep
Mesozoic platform & foredeep
Permo-Pennsylvanian platform
Cambrian-Devonian platform
Cambrian platform-basin of Main Ranges
Cambrian sandstone Neoprot. Miette Gp.
Dominantly paragneiss
Crystalline basement
Rubian Superterrane
Kootenay terrane
Quesnellia
Stikinia
Cambrian basinal facies rocks: exotic?
Neoproterozoic Windermere
Mesoproterozoic Belt-Purcell Spgps
Approximate northeastern limit of Rubian allochthons
Approximate northeastern limit of Rubian allochthons
Kootenay
Quesnellia
Golden
Banff
Calgary
IR
Kicking Horse Rim
0
100
km
F
FC
TO
Kx
N
V
K
O
PR
FB
Canada
U.S.A.
Basin fill deposits

Figure 33. Annotated photograph showing the Lewis thrust, along which rocks of the Belt-Purcell supergroups were transported eastward to now sit atop Cretaceous foredeep fill, Glacier National Park. In the model presented here, the Lewis thrust represents part of the suture between North America and the Rubian superterrane.

A Lower Cambrian Archeocyathid-bearing carbonate platform, known as the Cassiar platform (Fig. 5), occurs to the west of Selwyn basin. Rocks of the Cassiar platform are linked to those of the Selwyn basin to the east by a suite of 110–90 Ma plutons (Johnston, 2008). Both Johnston (2008) and Hildebrand (2009) argued that the Cassiar platform is a different, and older, platform than the North American platform, which contains a dominantly Lower Cambrian siliciclastic wedge overlain by a Middle Cambrian carbonate bank. In their models it is equivalent to the Antler shelf discussed earlier. Pope and Sears (1997) noted that miogeoclinal rocks found to the south were missing in the Idaho-Montana area and so suggested that rocks of the Cassiar platform in Canada escaped northward along the Tintina–Northern Rocky Mountain trench fault.

Omineca Belt Magmatism

A band of Cretaceous plutons (Figs. 5 and 22) occurs in the Canadian Cordillera from Yukon-Tanana terrane and Selwyn basin southward to the Lewis and Clark lineament (Monger et al., 1982). As to the south within the western United States, there was a magmatic lull in the northern Cordillera during the Early Cretaceous from ~140–135 Ma to 120–115 Ma (Armstrong, 1988). In the north, major plutonism flared up from ~120 Ma until 96 Ma, is progressively younger to the east, and includes both metaluminous and peraluminous bodies (Hart et al., 2004; Johnston, 2008). The Anvil-Hyland-Cassiar sub-belt intruded dominantly sedimentary rocks of the Selwyn basin and Cassiar platform for ~900 km along strike, range in age from 110 to 96 Ma, and are both peraluminous and metaluminous (Driver et al., 2000; Hart et al., 2004). Another linear band of plutons that intrudes sedimentary rocks of the Selwyn basin just to the east of the Anvil-Hyland-Cassiar sub-belt, is known as the Tombstone-Tungsten belt, and comprises generally small 96–90 Ma sub-alkalic to alkalic bodies with varied Au, Cu, Bi, W, Zn, Sn, Mo, and Sb mineralization (Hart et al., 2005). The younger suites extend along strike for over 1000 km into Alaska where they are known as the Livengood and Fairbanks-Salcha suites (Reifenstuhl et al., 1997a, 1997b; Newberry et al., 1990, 1996). Thus, there are major differences between the plutons in this region than other sectors. First, the Cordilleran-type magmatism ended at ~96 Ma, whereas elsewhere it ended around 82 Ma. The linear bands of 96–90 Ma metalliferous plutons have no obvious equivalents in the other sectors.

Kootenay Terrane

Rocks of this terrane (Figs. 5 and 32), which are metamorphosed and sit structurally upon more easterly allochthons of the Windermere and Purcell supergroups, are varied lower to middle Paleozoic sedimentary rocks, including Archeocyathid-bearing marbles—typical of the Cassiar platform, not North American cratonic sections—intruded by Ordovician–Devonian plutons. One group of rocks, the Lardeau group, was metamorphosed to quartzite, schist, and gneiss, and folded prior to deposition of sediments of the Mississippian basalt-bearing Milford Group (Read and Wheeler, 1975; Klepacki, 1985; Klepacki and Wheeler, 1985; Roback, 1993; Paradis et al., 2006). Smith and Gehrels (1992b) saw similarities of the Lardeau group to rocks of the Roberts Mountain allochthon of Nevada. Paradis et al. (2006) described Devonian–Mississippian volcanic and plutonic rocks of the Eagle Bay assemblage as an arc terrane built on the western edge of North America. Other packages of rocks within this terrane, such as the Neoproterozoic Horsethief Creek Group, the Eocambrian Hamill Group, the Archeocyathid-bearing Badshot Formation all lie west of a major fault named the Purcell fault, are not known to sit on North American basement, contain older deformations including huge westerly-vergent recumbent folds (Ross et al., 1985; Brown and Lane, 1988; Simony, 1992; Ferri and Schiarizza, 2006), which do not occur in rocks east of the Purcell fault, were transported a minimum of 200 km eastward (Price and Mountjoy, 1970), and were variously intruded by plutons ranging in age from late Paleozoic to Cretaceous (Okulitch et al., 1975; Parrish, 1992; Crowley and Brown, 1994; Colpron et al., 1998).

In the Selkirk fan structure, located on the eastern flank of the Monashee complex (Figs. 5 and 32, which is a erosional window that exposes probably duplexed North American basement and cover rocks beneath the Kootenay terrane, 187–173 Ma plutons were intruded before and/or during deformation and before a 173–168 Ma period of rapid exhumation of rocks from 7 kb to 3 kb (Colpron et al., 1996). Similarly, the Scrip nappe, a west-verging isoclinal structure located just to the north, has an overturned limb as wide as 50–60 km across strike and probably formed at about the same time (Raeside and Simony, 1983). As discussed earlier, similar structures farther south were documented and reported by Höy (1977) and, while their age is poorly constrained, appear to also have formed between 178 and 164 Ma (Read and Wheeler, 1975). Colpron et al. (1996, 1998) argued that these events took place in strata of the outer and proximal North American miogeocline (Colpron and Price, 1995), but there is simply no record, either deformational or sedimentological, of major plutonism, folding, thickening, and exhumation at this time on the North American cratonic terrace.

Yukon-Tanana Terrane

Colpron et al. (2006) divided rocks of this terrane (Fig. 5), which is plagued by poor bedrock exposure and extensive felsenmeer, into four assemblages: (1) the Snowcap assemblage, representing the oldest recognized rocks in the terrane and consisting of polydeformed and amphibolite-grade metasedimentary rocks intruded by Late Devonian–Early Mississippian plutons; (2) upper Devonian–Early Mississippian metasedimentary and metavolcanic rocks of the Finlayson assemblage; (3) mid-Mississippian–early Permian intermediate-mafic volcanic and volcaniclastic rocks that unconformably overlie the Snowcap and Finlayson assemblages; and (4) middle-late Permian calc-alkaline volcanic rocks and associated plutons of the Klondike assemblage. Despite the creation of complex models—based largely on geochemistry from strongly deformed and metamorphosed volcanic rocks and non-unique detrital zircon profiles from metasedimentary rocks—that relate the magmatism of the first three assemblages to an arc built on the western margin of North America above an eastward-dipping subduction zone (Piercey et al., 2006; Piercey and Colpron, 2009), cratonic basement within the terrane is unknown, as are sedimentological, magmatic, and deformational links with the North American passive margin.

Slide Mountain Terrane

Rocks of this terrane occur as structural slices separating rocks of Quesnellia and Yukon-Tanana terranes from Cassiar platform in the north and rocks of Kootenay terrane to the south (Fig. 5). They occur over the length of the Canadian sector and have different names depending on location (Harms, 1986; Struik, 1987; Schiarizza and Preto, 1987; Ferri, 1997). In the north they were thrust eastward over rocks of Cassiar platform along the Inconnu thrust (Murphy et al., 2006; Piercey et al., 2012). The allochthons are internally complex but include Carboniferous to Lower Permian dismembered basinal facies sedimentary rocks, including basalt, along with Upper Devonian to Permian ophiolitic material (Nelson, 1993; Roback et al., 1994; Murphy et al., 2006). Nelson (1993) demonstrated that rocks of the thrust belt represent a cross section of a collapsed oceanic basin between the lower Cassiar plate and the Yukon-Tanana arc located atop the overriding plate. Recent work by Beranek and Mortensen (2011) better documented the age of the collision to be 260–253 Ma and confirmed Permian ages for several plutons with the Yukon-Tanana arc.

Stikinia and Quesnellia

These two terranes (Fig. 5) are generally interpreted as arcs and are separated by the oceanic Cache Creek terrane (Monger et al., 1982). Stikinia is composed of Carboniferous to mid-Jurassic volcanic, plutonic, and sedimentary rocks, which based on Tethyan ammonites, developed well offshore from North America (Smith and Tipper, 1986; Schiarizza and MacIntyre, 1999). The terrane contains at least two ages of plutonism—one ranging from 220 to 193 Ma, another ranging from 181 to 165 Ma—and a thick succession of andesitic lavas, breccias, and rhyolitic ash-flow tuffs with ages between 185 and 174 (Whalen et al., 2001; MacIntyre et al., 2001).

In southern British Columbia, the Quesnel terrane comprises Upper Triassic–Lower Jurassic volcanic and sedimentary rocks sitting upon Triassic and upper Paleozoic rocks that themselves sit locally on 372 Ma gneisses (Beatty et al., 2006; Simony et al., 2006). Within this region, the upper Paleozoic Harper Ranch Group comprises a Late Devonian–Late Mississippian arc suite overlain by a Permian carbonate platform containing McCloud fauna (Beatty et al., 2006). The Late Triassic–Early Jurassic (227–210 Ma) Nicola Group sits atop the older rocks and contains a variety of augite porphyritic alkaline lavas, related breccias, pyroclastic rocks, and intercalated sedimentary rocks, which are of similar composition and age to rocks occurring locally from the Mojave Desert to the Yukon Territory (Mortimer, 1986, 1987; Monger, 1989; Monger and McMillan, 1989). Miller (1978) recognized the similarity of Triassic alkalic plutons in Quesnellia to those of the western United States. The upper part of the supracrustal sequence within Quesnellia is contained within the 204–187 Ma Rossland Group, which contains a diverse grouping of sparse carbonates, fine to coarse clastics, and a thick accumulation of dominantly calc-alkaline mafic to intermediate lava flows and breccias, pyroclastic flows, related epiclastic rocks and associated mafic-intermediate composition plutons (Tipper, 1984; Andrew and Höy, 1990, 1991; Höy and Dunne, 1997).

Farther north in British Columbia, both Stikine and Quesnel terranes contain Upper Triassic basaltic-andesitic volcanic and sedimentary rocks of the Takla Group, overlain by coarse sedimentary rocks, dacitic and andesitic lavas, breccias, minor basalt and rhyolite of the Lower Jurassic Hazleton Group (Monger and Church, 1977), interpreted to represent an extensional basin within the arc (Thorkelson et al., 1995). Dostal et al. (1999) showed that rocks of the Takla Group on both Stikina and Quesnellia are similar in terms of age, composition, and lithology and argued that they formed a continuous arc terrane. Earlier, other workers also recognized the similarities of the two terranes and suggested that Stikinia and Quesnellia were folded, along with the Yukon-Tanana terrane, during the Early to Middle Jurassic, around the Cache Creek terrane, which lies between the two (Nelson and Mihalynuk, 1993; Mihalynuk et al., 1994); or that Stikinia was forced northward from the area of the Columbia embayment during the Middle to Late Jurassic (Wernicke and Klepacki, 1988).

Cache Creek Terrane

The Cache Creek terrane (Fig. 5) is in fault contact with, and lies geographically between Stikinia and Quesnellia. It is interpreted to represent a Mississippian to early Jurassic accretionary complex holding fragments of the uppermost parts of basaltic seamounts capped with sedimentary rocks characterized by Upper Permian Tethyan fauna (Monger and Ross, 1971), mélange belts, radiolarian chert with tuffaceous veneer, graywackes, conglomerate, shale, mafic-ultramafic magmatic belts and ophiolitic assemblages, and pieces of reefal carbonate up to 75 km long and 40 km wide (Gabrielse, 1991; Johnston and Borel, 2007). Cherts as young as Toarcian are overprinted by blueschist-grade metamorphism that yielded $^{40}Ar/^{39}Ar$ of 174 Ma and were exhumed to provide debris to the adjacent Whitehorse trough by 171 Ma (Thorkelson et al., 1995; Mihalynuk et al., 2004). Cache Creek rocks were deformed by southwest-vergent folds and northeast-dipping thrusts, locally by northeast-vergent folds, and at their base is an east-dipping thrust that placed Cache Creek rocks upon those of Stikinia (Struik et al., 2001). A deformed pluton near the Stikinia–Cache Creek contact is 219 Ma; a pluton that cuts the thrust gave an age of 161 Ma; another pluton that cuts another thrust fault at the base of an ultramafic slab yielded a U-Pb zircon age of 166 ± 2 Ma; and a postdeformational pluton cutting deformed Cache Creek rocks yielded a U-Pb age of 172 Ma (Mihalynuk et al., 1992; Ash et al., 1993; Struik et al., 2001). These ages fit well with the youngest Jurassic plutonic suite in Stikinia (MacIntyre et al., 2001); the presence on Stikinia of Bajocian chert pebble conglomerates apparently shed from Cache Creek terrane (English and Johnston, 2005); and the presence of easterly derived debris of Bajocian age in the northern Bowser basin, which sits entirely on Stikinia (Ricketts et al., 1992). The eastern boundary of the Cache Creek terrane is a series of Cretaceous–Tertiary dextral strike-slip or oblique-slip faults (Gabrielse, 1985; Struik et al., 2001).

Sitting unconformably upon rocks of Stikinia, and locally Cache Creek terrane, are sedimentary rocks of two basins, Bowser and Sustut (Evenchick and Thorkelson, 2005; Ricketts, 2008). The Bowser basin (Fig. 5) is filled with >6000 m of Middle Jurassic to mid-Cretaceous marine and nonmarine clastics, while the Sustut basin contains >2000 m of nonmarine clastics and ranges in age from Aptian–Albian to Campanian (Evenchick et al., 2007).

Triassic Overlap Sequence

Several terranes within the Canadian Cordillera have conglomerate beds that overlap contacts with adjacent terranes and/or contain debris obviously derived from them, indicating that they were in close proximity by that time. Rocks of the Cassiar platform were overthrust by Permian rocks of Yukon-Tanana terrane prior to deposition of Triassic conglomerates containing clasts of blueschist and eclogite, which cover the suture zone (Murphy et al., 2006; Johnston and Borel, 2007). These relations and U-Pb geochronology demonstrate that Yukon-Tanana, Selwyn Basin, Cassiar terrane, and the Slide Mountain oceanic tract, were amalgamated by the Triassic (Beranek and Mortensen, 2007, 2011).

Coast Range Plutonic Complex

The Coast plutonic complex (Fig. 5) is a 50–175-km-wide belt of discrete and composite plutons ranging from gabbro to granite that extends from Washington to Alaska (Brew and Morrell, 1983; Armstrong, 1988; Mahoney et al., 2009), a distance of ~1700 km. There, much like the Peninsular Ranges batholith, two different age plutonic belts are apparently juxtaposed: a western belt, which sits on the previously amalgamated Alexander-Wrangellia composite terranes; the eastern belt, which sits on the joined Stikinia-Yukon Tanana terranes; and between the two, the Gravina-Dezadeash-Nutzotin-Gambier belt (Fig. 34), the remnants of a Middle Jurassic–mid-Cretaceous turbidite basin with local rhyolitic (177–168 Ma), andesitic and

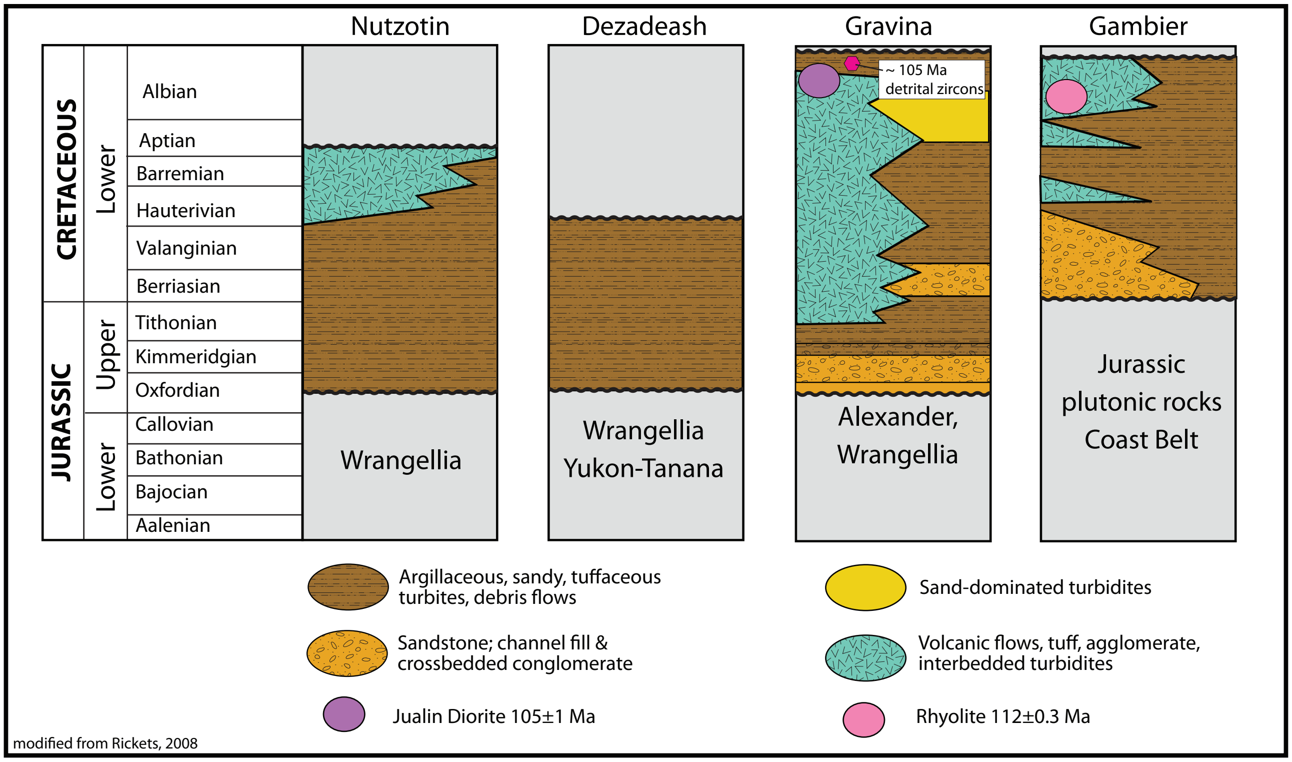


Figure 34. Schematic stratigraphic sections of the Nutzotin-Dezadeash-Gravina-Gambier basins showing the similarities of the units and ages. These basins occur between Insular and Intermontane superterranes and were deformed during the 105–100 Ma transpressional event. Modified after Ricketts (2008).

basaltic lavas (Berg et al., 1972; Rubin and Saleeby, 1991, 1992; Haeussler, 1992; Journeay and Friedman, 1993: Crawford et al., 2000; Manuszak et al., 2007; Gehrels et al., 2009), and in Alaska, turbiditic rocks of the Upper Jurassic–Upper Cretaceous Kahiltna basin (Ridgway et al., 2002; Kalbas et al., 2007). The western belt comprises plutons that appear to have been emplaced at discrete intervals: 177–162 Ma, 157–142 Ma, and 118–100 Ma; whereas the plutons of the eastern sector form a continuum from 180 to 110 Ma (Gehrels et al., 2009). A west-verging thrust belt, which developed between ~100–90 Ma (Haeussler, 1992; Rubin et al., 1990), places high-grade rocks of the eastern belt over lower-grade rocks of the Gravina-Dezadeash-Nutzotin belt, and possible equivalents farther south, to form a thrust stack that is of higher metamorphic grade upwards (Journeay and Friedman, 1993; Crawford et al., 2000; McClelland and Mattinson, 2000). A distinctive suite of 100–90 Ma plutons, containing large crystals of epidote and titanite, was intruded at >25 km depth during the deformation (Gehrels et al., 2009). In the southern part of the belt near latitude 49° 30′ N, pre–92 Ma plutons are clearly metamorphosed and folded (Figs. 35 and 36); whereas 90–84 Ma bodies are not obviously metamorphosed (Brown et al., 2000; Brown and McClelland, 2000), but are clearly deformed and could be folded (Fig. 37). Late Cretaceous–Paleocene magmatism was generally focused into a narrow, but linear, belt at least 1000 km long, just to the east of the Coast shear zone–Coast Range mega-lineament, where steeply-dipping sills of foliated tonalite were emplaced at ~70 Ma in the north and 60–50 in the southeast, along with 76–67 Ma zoned bodies of marginal diorite that grade inward to tonalite and granodiorite (Crawford et al., 1987; Rusmore et al., 2001; Gehrels et al., 2009; Mahoney et al., 2009). An eastward-vergent thrust belt occurs within the south-central part of the area and was active from ~87 to 68 Ma (Rusmore and Woodsworth, 1991; Evenchick et al., 2007).

Cretaceous Overlap Sequence

In south-central British Columbia, the Insular and Intermontane superterranes were apparently joined by the early Upper Cretaceous (95–85 Ma), as documented by two sedimentary successions known as the Powell Creek and Silverquick formations (Fig. 5), which constitute a volcano-sedimentary overlap succession sitting unconformably atop the two superterranes (Garver, 1992; Mahoney et al., 1992; Schiarizza et al., 1997; Riesterer et al., 2001; Haskin et al., 2003). Paleomagnetic study of the overlap succession, including 26 sites in volcanic lava flows and 54 sedimentary sites, along with several positive contact, conglomerate, and tilt tests combine to provide a reliable and robust record of the geomagnetic field inclination for the area and yield a paleolatitude of 39.5° ± 2.2°, which is ~20.3° ± 2.7° south of the expected paleolatitude of North America at that time (Enkin

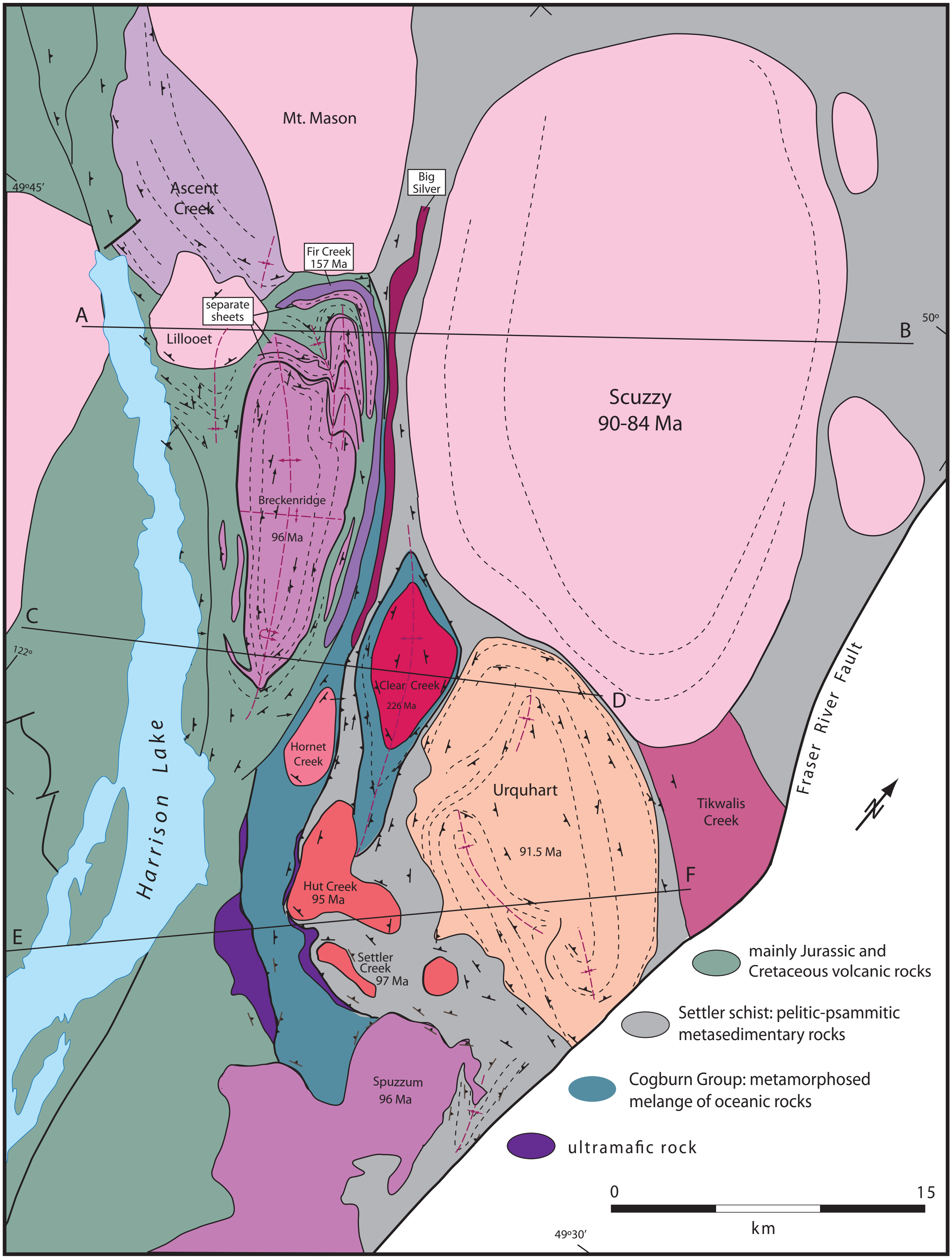


Figure 35. Folded plutons of the Coast Range plutonic complex, southern British Columbia. Modified after Brown and McClelland (2000) and Brown et al. (2000).

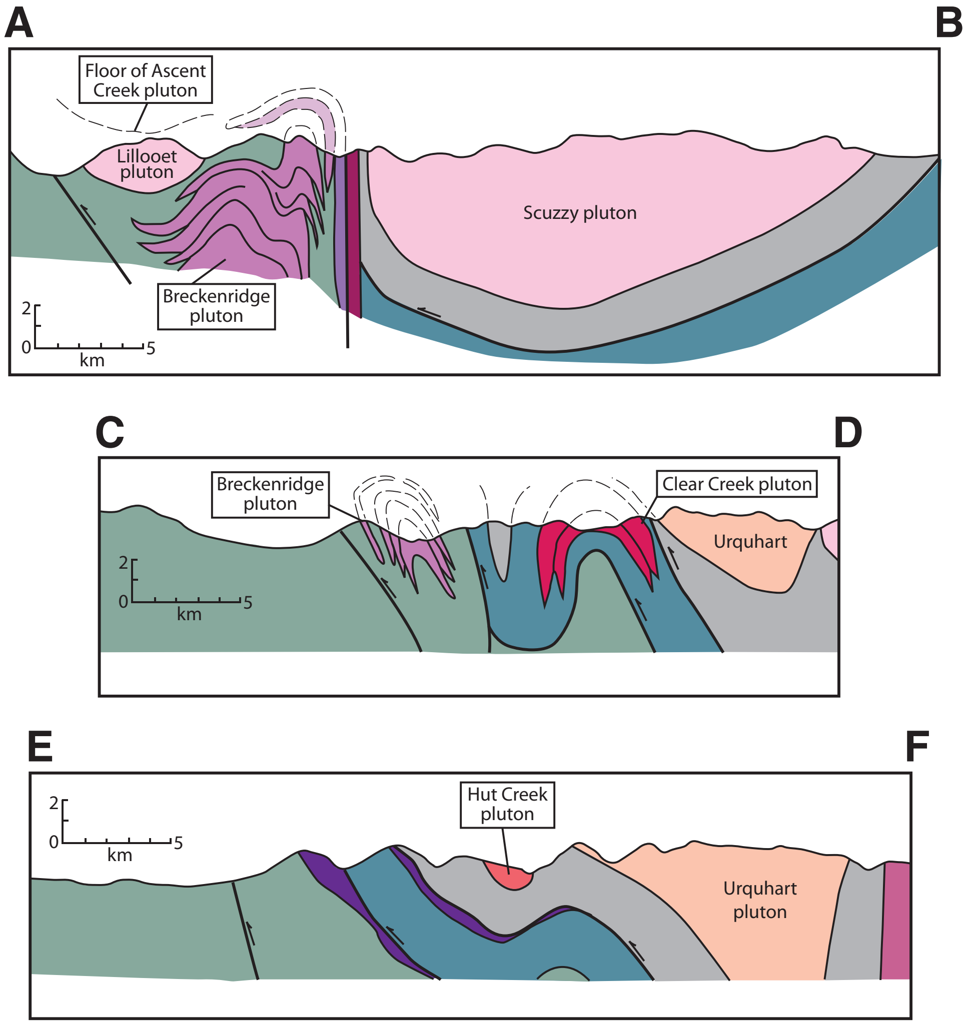


Figure 36. Cross sections from Figure 34 modified from Brown et al. (2000), showing the folded plutons.

Figure 37. Inclined sills at the margin of the Skuzzum pluton. Both the internal fabric and the inclined nature of the sheets and contacts of this pluton suggest that it, like older bodies in the same area of the Coast plutonic complex, is folded. Photo courtesy of Ned Brown.

et al., 2003; Enkin et al., 2006b; Enkin, 2006; Kent and Irving, 2010). These results are consistent with the 70 Ma paleopoles obtained from the Carmacks volcanics, located at the northern end of the Coast plutonic complex in the Yukon Territory (Fig. 5) and discussed earlier (Wynne et al., 1998; Johnston et al., 1996; Enkin et al., 2006a).

Late Cretaceous–Early Tertiary Magmatism

Within the Coast Mountains of western Canada, Jurassic to Late Cretaceous tonalitic-granodioritic plutons, deformed and metamorphosed to gneiss under amphibolite-granulite conditions, and generally considered to constitute the lower and middle crust of a Cordilleran-type magmatic arc, were rapidly exhumed by ~65–60 Ma (Armstrong, 1988; Hollister, 1982; van der Heyden, 1992; Crawford et al., 1999). Extension, which took place mainly prior to 60 Ma and involved at least 15 km of tectonic exhumation, was accompanied by a voluminous Late Cretaceous–early Tertiary intrusive bloom (Fig. 22) derived from multiple sources (Hollister and Andronicos, 2000, 2006; Hollister et al., 2008; Mahoney et al., 2009; Andronicos et al., 2003). The strongly linear locus of uplift and plutonism coincides with a major fault or deep crustal edge, which was imaged on two LITHOPROBE deep reflection seismic lines (Cook et al., 1992, 2004). The shutdown of arc magmatism just before the Late Cretaceous, the rapid exhumation of the central gneiss complex by ~60 Ma, and the voluminous Late Cretaceous–early Tertiary magmatism led Hildebrand (2009) to identify these rocks as slab-failure magmatism. He argued that magmas streaming through the narrow tear in the subducting plate might explain the highly focused and elongate nature of the Late Cretaceous tonalitic bodies along the western margin of the belt (e.g., Barker and Arth, 1990).

North Cascades

In the North Cascades of Washington (Figs. 5 and 38), a south-southeastward continuation of the Coast plutonic complex, there is also an ~100 Ma west-vergent thrust belt with associated metamorphism (Misch, 1966; Mattinson, 1972; McGroder, 1991; Miller et al., 2009), and younger Paleocene plutons emplaced between ~68 and 59 Ma (Miller et al., 1989). The 96–91 Ma Mount Stuart batholith (Fig. 38) postdates the older thrusts but appears to predate another two periods of deformation as it wraps around fold noses with foliations and lineations that cut across internal contacts and compositional heterogeneities, such as mingled areas (Paterson and Miller, 1998; Matzel et al., 2006). Other intrusive suites in the area, such as the Napeequa complex and Cascade River suite (Fig. 38) were deformed at ~88–76 Ma (Matzel et al., 2004). Also located within the North Cascades are rocks known as the Swakane gneiss (Figs. 8, 22, 29, and 38), which are compositionally, texturally, and temporally similar to the Orocopia-Pelona schists of southern California in that they contain detrital zircons as young as ~73 Ma, and were metamorphosed under P-T conditions of 9–12 kbar and 640–740 °C within 5 Myr of deposition (Matzel et al., 2004). A foliated granodiorite at the Swakane-Napeequa contact was dated to be 84 ± 1 Ma (Hurlow, 1992).

Within and west of the crystalline Cascades core, spanning the Straight Creek–Fraser River fault and to the west in the San Juan Islands is an imbricate stack (Fig. 38) of varying units that was assembled after 110 Ma, and seemingly before intrusion of 96–90 Ma plutons (Brandon et al., 1988; Brown and Dragovich, 2003; Brown and Gehrels, 2007; Brown et al., 2007). Units within the stack include schists and numerous mélange belts, some of which hold blueschist-grade metamorphism, the ophiolitic 170–160 Ma Fidalgo complex (Garver, 1988), which may correlate with the ophiolitic 161 Ma Ingalls complex to the east (Miller, 1985; Miller et al., 2003; MacDonald et al., 2008), the Lookout Mountain Formation (MacDonald, 2006), which contains detrital zircons as young as 160 Ma (MacDonald et al., 2003, 2008), the high-P Permo-Triassic Vedder complex (Armstrong et al., 1983), and the pyroxene-bearing and gneissic Yellow Aster complex (Misch, 1966; Mattinson, 1972), which has a similar history to rocks within the Yukon-Tanana and Shoo Fly terranes (Brown and Gehrels, 2007). Just to the west the rocks have many similar attributes and ages to the Franciscan (Brown, 1986) in that they were deformed between 110 and 80 Ma, contain older high-grade blocks ranging in age from 160 to 144 Ma, blueschist slabs of 130–120 Ma, gabbro-tonalites of 164–163 Ma, ophiolite dated at 167 ± 5 Ma, and mélange matrices ranging variously in age from 114 to 90 Ma (Brown and Gehrels, 2007). Both paleomagnetic and faunal evidence suggest that these rocks were located well to the south at 75 Ma (Brown et al., 2007).

Idaho Batholith

Rocks more or less correlative with those in the Cascades occur to the southeast across the sinistral Lewis and Clark lineament and Orofino fault (Armstrong et al., 1977; Wallace et al., 1990; McClelland and Oldow, 2007). There (Figs. 5 and 8), the mid-Cretaceous Atlanta and Bitterroot lobes of the Idaho batholith intrude a variety of rocks including the Mesoproterozoic Belt supergroup and sit along the eastern side of the Salmon River suture, a westerly-vergent thrust belt, and the younger western Idaho shear zone, which is generally interpreted as a locus of continental scale transcurrent motion (McClelland et al., 2000). Along the western margin of the batholith are variably deformed, and in places epidote-bearing, tonalitic to granitic orthogneisses in the age range 118 ± 5 to 105 ± 1.5 Ma cut by epidote-bearing tonalitic sheets in the age range 92 to 90 Ma (Taubeneck, 1971; Hyndman, 1983; Manduca et al., 1993; Giorgis et al., 2008). To the north, just south of the Orofino fault—in what is a probable erosional duplex structure—a window through westerly-vergent thrust faults carrying Neo- and Mesoproterozoic rocks, exposes a group of 94–86 Ma paragneisses and 94–73 Ma orthogneiss, neither of which appear to occur to the west across the suture and fault, and small slivers of ultramafic rocks, that were deformed as late as 68–61 Ma (Lund et al., 2008). Elsewhere in the batholith

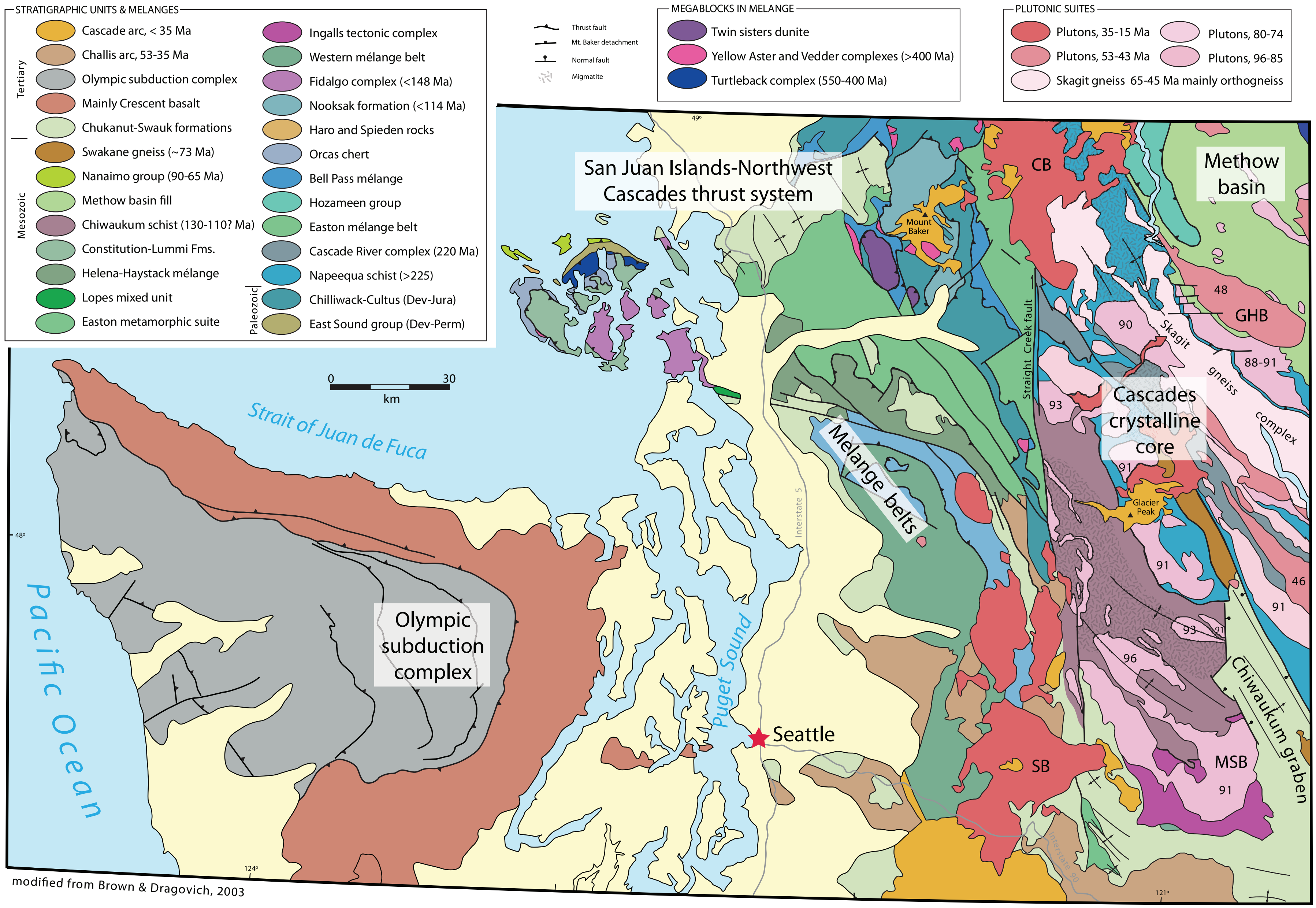


Figure 38. Geological sketch map of northwestern Washington, illustrating the basic geology of the San Juan Islands–Northwest Cascades thrust system, the western Cascade crystalline core, and surrounding area. CB—Chilliwack batholith; GHB—Golden Horn batholith; MSB—Mount Stuart batholith; SB—Snoqualmie batholith. Modified from Brown and Dragovich (2003) with ages of plutons from Miller et al. (2009).

are early metaluminous tonalite, biotite granodiorite, K-feldspar poikolitic granodiorites and granites that fall between 98 and 80 Ma (Kiilsgaard et al., 2001; Gaschnig et al., 2010). The most voluminous phases of the Atlanta lobe are slightly peraluminous, 83–67 Ma, plutonic complexes that comprise biotite granodiorite cored by muscovite-biotite granite, all cut by dikes of leucogranite (Kiilsgaard and Lewis, 1985; Lewis et al., 1987; Gaschnig et al., 2010). While the Bitterroot lobe doesn't appear to contain the older metaluminous rocks found in the Atlanta lobe, it does contain younger metaluminous bodies with ages of 74–69 Ma; however, the bulk of the lobe comprises peraluminous bodies, which range in age from 66 to 53 Ma, are compositionally more heterogeneous than their more southern counterparts, and have associated mafic dikes, sills, and small plutons, which are notably absent in the Atlanta lobe (Hyndman, 1984; Hyndman and Foster, 1988; Foster and Hyndman, 1990; Foster and Fanning, 1997; Gaschnig et al., 2010).

The Boulder batholith, located just to the east and associated with the Elkhorn volcanics (Fig. 8), is a composite batholith comprising small mafic-intermediate plutons, voluminous granites, and late-stage leucocratic bodies, both a potassic and a sodic series, and overall ages in the range 78 to 66 Ma (Hamilton and Myers, 1967; Smedes et al., 1973; Tilling, 1973, 1974; Johnson et al., 2004). Porphyry copper mineralization is associated with this magmatism at Butte (Lund et al., 2002; Dilles et al., 2003). The area was subsequently the site of both volcanism and plutonism of the 53–43 Ma Challis magmatic field (McIntyre et al., 1982; Johnson et al., 1988; Moye et al., 1988).

Blue Mountains Terranes

This assemblage of three different terranes (Figs. 5 and 8) and an overlap sequence lies just west of the Salmon River suture in eastern Oregon–western Idaho and is commonly correlated with rocks to the north as it has geological and faunal similarities to them. The terranes are: the Olds Ferry terrane, which is the southeasternmost of the group, and consists of Middle to Late Triassic mafic to intermediate volcanic rocks in the lower part and rhyodacitic-rhyolitic flows and breccia, volcaniclastic sandstone and conglomerate in the upper; the Baker terrane, comprising sheared Permian to Early Jurassic chert and argillite holding blocks of Devonian–Triassic limestone, serpentinite, mafic and ultramafic rocks, with local blueschist-grade metamorphism; the Wallowa terrane, dominated by Permian to Early Jurassic volcanic and sedimentary rocks; and the Izee overlap sequence, which is a thick overlap succession of Triassic–Jurassic sedimentary rocks that sit unconformably upon the other three terranes (Avé Lallemant, 1995; Dorsey and LaMaskin, 2008). Jurassic–Early Cretaceous plutons stitch the terranes. Originally, the terranes were oriented N-S but were rotated to their current NE-SW orientation after they were assembled (Housen, 2007). The Baker terrane, which sits between the Olds Ferry and Wallowa terranes—both readily interpreted to represent arcs—is bound on both sides by thrusts that dip away from it to form a doubly-vergent thrust belt (Avé Lallemant, 1995), which led Dorsey and LaMaskin (2007) to suggest a Moluccan sea–type arc-arc collision.

Several workers recognized the similarities of arcs of the Blue Mountains to Stikinia and Quesnellia and so suggested that they correlate with the two arcs, Wallowa and Olds Ferry, and that the accretionary rocks of the Cache Creek are similar to those of the Baker terrane (Mortimer, 1986; Stanley and Senowbari-Daryan, 1986). However, there are two subterranes within the Baker terrane: Greenhorn subterrane, which is a serpentinite matrix mélange that contains only fusulinids of McCloud affinity, and Bourne subterrane, which is dominated by fine-grained argillite and contains both Tethyan and McCloud fusulinids (Ferns and Brooks, 1995; Schwartz et al., 2011). Paleontological data from the Izee terrane indicate that it originated at low paleolatitudes during the Late Triassic and migrated to higher paleolatitudes by the Middle Jurassic (Pessagno, 2006).

A basinal remnant, the Ochoco, overlaps the westernmost outcrops of the Baker and Izee terranes, where it comprises a thick succession of middle to Late Cretaceous marine sedimentary rocks, mainly mudstone, sandstone, and conglomerate (Dorsey and Lenegan, 2007). Detrital zircons collected from the basin overlap in age with those of the Hornbrook basin and document that the basin is no older than Cenomanian (Kochelek and Surpless, 2009) and was probably part of the same basin as the Hornbrook (Surpless et al., 2009). Paleolatitudes, determined by paleomagnetic analysis, show that the rocks of the basin—and those of the Blue Mountains in general—were located 1200–1700 km farther south (32° ± 7°) during the Cenomanian (Housen and Dorsey, 2005), but were in their present position by the Eocene, when they were onlapped by the Clarno volcanics (Grommé et al., 1986). Tilt-corrected directions for these rocks are essentially identical to rocks of the Hornbrook basin, which supports a link between the two basins (Housen and Dorsey, 2005).

ALASKAN SECTOR

Brooks Range, North Slope, and Alaskan Orocline

The geology of the Brooks Range and Alaska's North Slope is quite complex but in the most general sense is divided into the Arctic Alaska–North Slope terrane and the Angayucham terrane (Fig. 5), which were both deformed and metamorphosed when rocks of the Arctic Alaska terrane and its passive margin were partially subducted beneath the Angayucham terrane during the Late Jurassic–Early Cretaceous Brookian orogeny (Moore et al., 1994). The stratigraphy of the Arctic Alaska terrane is broken into three main groups: pre-Mississippian rocks, the Ellesmerian, and the Brookian (Lerand, 1973; Moore et al., 1994; Handschy, 1998). The pre-Mississippian sequence comprises variable sections of Neoproterozoic rocks—some of which are quite similar lithologically to those of the Windermere Supergroup (Moore and Bird, 2010), but different in age from rocks of northern

Canada's Amundsen Basin (Macdonald et al., 2009)—overlain by lower Paleozoic siliceous and clastic sedimentary and volcanic rocks. The Paleozoic and older rocks were intruded by Devonian plutons, then deformed and metamorphosed in the north before being covered by rocks of the Ellesmerian sequence, which collectively form a southward-facing (present coordinates) passive margin consisting of a basal Mississippian nonmarine conglomerate overlain by Mississippian–Triassic clastic-carbonate sedimentary rocks (Brosgé et al., 1962; Martin, 1970; Brosgé and Tailleur, 1971; Mull, 1982; Hubbard et al., 1987; Mayfield et al., 1988; Moore et al., 1994). The rocks of both the Franklinian and Ellesmerian sequences were detached from their basement, folded, and thrust to the north during the Brookian orogeny.

Like the other parts of the Cordilleran orogen, a foreland basin, there termed the Colville basin, developed upon the passive margin during the Mesozoic–Tertiary, and its rocks are collectively included within the Brookian sequence. This sequence contains sedimentary rocks related to two phases of deformation: an early phase, termed the Brookian orogeny, and characterized by the emplacement of far-traveled north-vergent thrust sheets holding rocks as young as Aptian, and a dominantly Maastrichtian–Cenomanian phase involving thrusts of little displacement and folds (Mull, 1985; Moore et al., 1994).

The Endicott Mountains and DeLong Mountains allochthons (Fig. 5) are the structurally lowest allochthons and were emplaced during the Brookian orogeny (Moore et al., 1994). The internal stratigraphy of the allochthons, although much disrupted by internal thrust faults and folds, are similar to that of the North Slope except that they don't contain rocks older than uppermost Devonian, as the sections are truncated by the basal detachment, and there is no apparent evidence for the sub-Mississippian unconformity (Moore et al., 1994; Handschy, 1998).

The southern half of the Brooks Range is dominated by the Hammond and Coldfoot subterranes (Fig. 5), both structurally imbricated, but variably deformed, probable Proterozoic to lower Paleozoic assemblages of carbonate, schist, and phyllite variably metamorphosed to greenschist, blueschist, and amphibolite intruded by Late Proterozoic and Devonian plutons (Dillon et al., 1980, 1987; Karl et al., 1989; Till, 1989). A narrow band of Paleozoic phyllite, slate, and sandstone, commonly at blueschist grade, along with spotty mafic intrusions and mélange sit along the southern margin of the Brooks Range, and were collectively termed the Slate Creek subterrane by Moore et al. (1994). Similar rocks, but not necessarily correlative, occur on the Seward Peninsula, where they are known as the Nome complex (Till et al., 2010). There, U-Pb dating of detrital zircons found extensive Neoproterozoic peaks as well as Paleozoic peaks, which serve to demonstrate that rocks of the Seward Peninsula are exotic with respect to North America (Amato et al., 2009).

The structurally highest allochthons of the Brooks Range occur within the Angayucham terrane, which crops out to the south of the Slate Creek belt and as klippen atop the range itself (Mull, 1982; Patton et al., 1994; Moore et al., 1994). The rocks are generally divided into two structural packages: a lower assemblage of greenstones, mainly pillow basalt with subordinate Late Devonian to Early Jurassic chert; and an upper assemblage of Middle Jurassic gabbro and ultramafic rocks (Zimmerman and Soustek, 1979; Pallister and Carlson, 1988; Patton and Box, 1989) cut by plagiogranite dated at 170 Ma and 163 Ma aplite dikes (Moore et al., 1994). The disparate ages for the two assemblages led to models in which the lower volcanic assemblage was thrust beneath the upper ophiolitic assemblage during the Jurassic at ~154 Ma prior to the emplacement of Arctic Alaska below the composite allochthon during the Valanginian (Boak et al., 1987; Mayfield et al., 1988).

The Tozitna terrane (Fig. 5), located to the south and thrust upon the Ruby terrane, was considered a separate terrane by Silberling et al. (1992); however, it contains the same age rocks in nearly identical thrust sequence assembled in the same structural order as the Angayucham terrane. This led some workers (Patton et al., 1989, 1994) to suggest that the two were formerly part of the same terrane.

In similar fashion to the Western Interior Basin of Canada, the Jurassic and early Cretaceous fill, termed the proto–Colville basin by Moore et al. (1994), is relatively thin and consists of mainly turbidites, whereas the younger Colville basin fill is much thicker and coarser (Bird and Molenaar, 1992). The Colville basin—an orogenic foredeep mainly developed on top of another sedimentary sequence, the Beaufortian, itself derived from the Barrow arch, the elevated rift shoulder of the Amerasian basin, which was open to the north by the Early Cretaceous (Grantz and May, 1982; Grantz et al., 1990)—was the locus of sedimentation from the collisional orogen within the Brooks Range to the south (present-day coordinates).

Highly deformed Berriasian to Valenginian turbidites with local olistostromes, conglomerate lenses, and thin Buchia coquinoid limestone beds known as the Okpikruak Formation sit atop the Endicott and DeLong Mountain allochthons in the central and western Brooks Range, and are generally considered to represent debris shed northward during the earliest part of the collision (Bird and Molenaar, 1987, 1992). Correlative rocks in the more northern portions are a unit of pebbly shale holding chert and quartzite pebbles, ironstone concretions, and frosted quartz grains, apparently deposited during Jurassic–Hauterivian uplift and erosion of the outer part of the south-facing platform during the opening of the Amerasian basin (Macquaker et al., 1999). Rocks of the Okpikruak Formation don't contain a detrital zircon profile compatible with allochthonous rocks of the Brooks Range so may have been deposited atop the margin prior to thrusting, or if younger, composed of debris transported longitudinally from a considerable distance (T. Moore, 2011, personal commun). This fits with evidence for dramatic downward flexure of the Colville Basin during the Barremian–Aptian (Cole et al., 1997) and appears to coincide with thrusting to the west in the Lisburne Hills (Moore et al., 2002). Sitting unconformably upon rocks of the Okpikruak Formation and eroded rocks of the Endicott and DeLong Mountain allochthons, are up to 3400 m of Aptian–Cenomanian sandstone, conglomerate, mudstone, and coal beds, collectively termed the Fortress Formation (Mull, 1985; Siok, 1989; Crowder, 1989; Bird and Molenaar, 1992).

Along the south side of the Angayuchan terrane are rocks of the Kobuk-Koyukuk basin (Fig. 5), which is a U-shaped basin consisting of 5–8 km of middle to Upper Cretaceous terrigenous clastic rocks (Nilsen, 1989) bordered on three sides by, and probably derived from, similar metamorphosed Proterozoic and Paleozoic continental terranes: Seward Peninsula on the west, Brooks Range on the north, and Ruby geanticline on the east (Patton and Box, 1989; Patton et al., 1994). Occupying the central portion of the basin is the Koyukuk terrane, which comprises a lower sequence of poorly dated basaltic-andesitic lavas and ultramafic rocks intruded by tonalitic-trondhjemitic plutons, all unconformably overlain by dominantly Berriasian–Valanginian—but locally as young as Aptian—shallow to deep marine volcaniclastic rocks, hyaloclastites, pillowed plagioclase-phyric basalts, and dacitic lavas that collectively range from arc tholeiites to shoshonites (Patton and Box, 1989; Patton et al., 1994, 2009).

To the east, and structurally beneath the Angayucham-Tozitna terranes, lies the Ruby terrane (Patton et al., 1989). The terrane (Fig. 5) is a NE-trending linear belt of Precambrian–Paleozoic pelitic schist, quartzite, metabasite, marble, and orthogneiss, with varying metamorphic grades such as greenschist, blueschist, and amphibolite, cut by Devonian plutons, that collectively form the pre-mid Cretaceous core of a northeasterly trending uplift (Patton et al., 1994). Based on the similarities of rock types, metamorphic grades, presence of Devonian plutons, and tectonic setting, most researchers favor the concept that the Ruby and Arctic Alaska terranes once formed a continuous hinterland (for example, Carey, 1958; Tailleur, 1980; Box, 1984; Grantz et al., 1991; Roeske et al., 1995; Johnston, 2001).

Mid to Late Cretaceous plutons occur in two regional bands (Fig. 5): one a westerly trending band that occurs within the Yukon-Kubak-Koyukuk basin and on the Seward Peninsula; and another that occurs over the length of the Ruby terrane (Patton and Box, 1989; Miller, 1989; Arth et al., 1989a; Patton et al., 1987, 2009; Till et al., 2010). The plutons—suggested by Arth et al. (1989b) to have been derived from a mixture of melts derived from old continental mantle and old continental crust—are highly variable along strike and range from calc-alkaline to ultrapotassic alkalic bodies. Those of the Ruby belt are all calc-alkaline, and also variable along strike in that those in the southwest were probably derived from old continental crust, whereas those to the northeast more likely originated as melts of oceanic rocks and young crust contaminated by old crust (Arth et al., 1989a). The plutons clearly cut the thrusts of the Ruby-Angayucham package (Patton et al., 2009), and are 112–99 Ma (Miller, 1989), so are clearly postcollisional with respect to the Brookian orogeny.

Located to the south of the Ruby terrane is the remote Farewell terrane (Fig. 5), which is a composite terrane comprising the Nixon Fork, Dillinger, Minchumina, and Mystic subterranes in some schemes (Silberling et al., 1992) or two distinctive sedimentary sequences, the White Mountain and Mystic, in others (Decker et al., 1994). The Nixon Fork subterrane (Fig. 5) consists of a SE-dipping Precambrian metamorphic basement of greenschist, greenstone, and minor siliceous plutonic rocks (Patton et al., 1980) that range in age from at least 1265 ± 50 Ma to ~850 Ma (Dillon et al., 1985; McClelland et al., 1999), all overlain by ~5000 m of Ordovician–Late Devonian carbonates (Patton et al., 1994). The Dillinger subterrane (Fig. 5) comprises up to nearly 3000 m of complexly faulted and folded Paleozoic turbiditic rocks, minor greenstone, black shale and chert, laminated limestone, sandstone, and breccia, which were interpreted to represent a basinal facies for the platformal and slope-facies rocks of the Nixon Fork subterrane (Churkin and Carter, 1996; Decker et al., 1994). Pennsylvanian–Permian to Lower Cretaceous terrigenous clastic rocks of the Mystic subterrane, or sequence, unconformably overlie the older rocks of both Dillinger and Nixon Fork subterranes (Decker et al., 1994; Bundtzen et al., 1997). The Minchumina subterrane (Fig. 5) consists of deep water carbonate, chert, argillite, and quartzite of Late Proterozoic and lower Paleozoic sedimentary rocks that may represent offshore deposits of the Nixon Fork–Dillinger succession (Patton et al., 1994). Paleozoic rocks of the Nixon Fork are now known to be platformal facies rocks that interfinger with more basinal facies rocks of the Dillinger succession; so Decker et al. (1994) included them together as the White Mountain sequence. Lower Permian $^{40}Ar/^{39}Ar$ plateau ages suggest a late Paleozoic deformational event, which Bradley et al. (2006) named the Browns Fork orogeny.

The obvious bend of Alaskan terranes around the Kobuk-Koyukuk basin (Fig. 5) has drawn the attention of geologists since Carey (1955, 1958) included it as one of his oroclines. Johnston (2001, 2008) used the oroclinal concept to fold the ribbon continent as it migrated northward, whereas Box (1985) invoked an irregular margin, or promontory model, to explain the sinuous trace of the terranes.

Wrangellia Composite Terrane

Wrangellia is a dismembered composite superterrane located in westernmost Canada and southern Alaska (Fig. 5) where it spans both the Canadian and Alaskan sectors. It is generally considered to contain three terranes: Alexander, Wrangell, and Peninsular (Jones et al., 1977; Nokleberg et al., 1994, 2000). Wrangell terrane is dominated by 230–225 Ma flood basalts, known collectively as the Karmutsen Formation on Vancouver and the Queen Charlotte islands, where ~6 km of high-Ti basalt and picrites, pillows, hyaloclastites with uppermost subaerial lavas are well-preserved, and the Nikolai Formation in Alaska and Yukon Territory, where 1–3.5 km of dominantly subaerial high-Ti basalts with basal high-Ti pillow basalts are preserved (Lassiter et al., 1995; Greene et al., 2008, 2009, 2010). On Vancouver Island the basalts are the basal part of the uppermost sheet of an imbricate thrust stack (Monger and Journeay, 1994) and are overlain by platformal carbonates, correlative with similar rocks in south-central Alaska (Carlisle and Susuki, 1974; Jones et al., 1977; Yorath et al., 1999). The platformal limestones are overlain by uppermost Triassic and Jurassic shale, limestone, and argillite intercalated with, and overlain by, the Jurassic Bonanza arc volcanics (Caruthers and Stanley, 2008; Nixon et al., 2006).

The 202–165 Ma Bonanza arc of Vancouver and the Queen Charlotte islands comprises up to 2500 m of interbedded lava, tuff, and breccia ranging in composition from basalt to rhyolite (DeBari et al., 1999). Two groups of intrusions, the West Coast Crystalline complex and the sheeted Island Intrusions are generally considered to represent the plutonic roots of the arc and range in age from 190 to 169 Ma (Isachsen, 1987; DeBari et al., 1999; Canil et al., 2010).

Rocks interpreted to be part of the Bonanza arc occur on the mainland, and nearby islands, of southern British Columbia where they are known as the Bowen Island Group (Friedman et al., 1990). Strata of the group were tightly folded and foliated prior to the emplacement of a pluton dated to be 154 Ma (Friedman and Armstrong, 1995). This deformation might be more extensive (McClelland and Gehrels, 1990; Monger, 1991).

The Alexander terrane (Berg et al., 1972) underlies most of southeastern Alaska and parts of Yukon Territory and British Columbia (Fig. 5). It is dominantly a series of amalgamated and variably deformed and overlapping lower Paleozoic arc terranes, themselves overlain by Upper Paleozoic and Triassic carbonate and clastics intercalated with variable amounts of basaltic lavas and breccias (Jones et al., 1972; Gehrels and Saleeby, 1987; Gehrels, 1990; Gehrels et al., 1996). Based on varied and distinctive megafauna, as well as detrital zircons, rocks of the terrane appear to be of Baltic or possibly Siberian origin (Bazard et al., 1995; Gehrels et al., 1996; Soja and Antoshkina, 1997; Blodgett et al., 2002, 2010; Pedder, 2006). A pluton dated to be Pennsylvanian at 302 Ma intrudes both Wrangell and Alexander terrane and so provides a minimum age for amalgamation of those two terranes (Gardner et al., 1988).

The Peninsular terrane is located mostly in Alaska, although the Bonanza arc represents the terrane in southern Canada (Nokleberg et al., 2000). The terrane, which stretches for over 1000 km within south-central Alaska (Fig. 5), is dominated by the Talkeetna-Bonanza arc and what is known as the Border Ranges ultramafic and mafic complex, which lies between rocks of the arc and the Border Ranges fault (Smart et al., 1996; Pavlis and Roeske, 2007) separating Peninsular terrane from the Chugach accretionary complex, and is interpreted to represent basement to the Talkeetna arc (Burns, 1985; DeBari and Coleman, 1989; Kusky et al., 2007; Farris, 2009).

The Talkeetna arc is well exposed in oblique cross section and exposes a dismembered section of the arc from mantle tectonite to overlying sedimentary rocks (Greene et al., 2006; Hacker et al., 2008). In the Chugach and Talkeetna mountains, ~7 km of subaerial and submarine lavas, breccias, tuff, ash-flow tuff, and volcaniclastic deposits are preserved (Clift et al., 2005) and cut by elongate quartz diorite and trondhjemitic plutons (Rioux et al., 2007). The main phase of Talkeetna arc volcanism occurred from 202 Ma to 175 Ma and several plutons intruded the northern part of the area between 190 Ma and 153 Ma (Rioux et al., 2007; Hacker et al., 2011). Farther west on Kodiak Island and the Alaskan Peninsula is an extensive Jurassic batholith dominated by quartz diorite and tonalite with the oldest age of 213 Ma on Kodiak Island and a younger group to the north on the peninsula with ages ranging from 184 to 164 Ma (Rioux et al., 2010). Thus, in both the east and west magmatism is younger northward. Isotopes, as well as detrital and inherited zircon populations, vary systematically within the arc along strike from more juvenile in the east to more evolved in the southwest, suggesting differences in basement geology (Clift et al., 2005; Greene et al., 2006; Rioux et al., 2007; Amato et al., 2007).

Within the eastern Alaska Range there are several linear bands of plutons: the Late Jurassic–Early Cretaceous, calc-alkaline Chitina plutons, which intrude rocks of the southern part of the superterrane; the Chisana arc, which is located a bit farther north in Wrangellia and comprises calc-alkaline plutons and andesitic lavas; and the Kluane suite of latest Cretaceous–early Tertiary plutons that intrude both Wrangellia and Yukon-Tanana terranes (Trop and Ridgway, 2007). Hildebrand (2009) interpreted the Late Cretaceous–early Tertiary plutons to be part of an extensive band of slab-failure magmatism.

During the Upper Jurassic, rocks along the southern margin of Wrangellia were progressively thrust northward along south-dipping thrust faults contemporaneous with development of the clastic Oxfordian to Tithonian Nutzotzin–Wrangell Mountains basin, which sits unconformably upon platformal carbonates and coarsens upward from marine mudstones and sandstones to conglomerate (Trop et al., 2002; Manuszak et al., 2007). This basin is generally considered to represent a back-arc basin to the Talkeetna arc (Trop and Ridgway, 2007; Manuszak et al., 2007). Another Oxfordian to Tithonian basinal fill succession, comprising huge fan-delta complexes with coarse bouldery debris shed from reverse fault scarps sits on the south side of the Talkeetna arc, where it is known as the Naknek Formation (Trop et al., 2005). Paleomagnetic and faunal data indicate that the Wrangellian superterrane was translated at least 30° northward relative to cratonic North America since the Triassic and that it was at its present latitude by ~52 Ma (Jones et al., 1977; Hillhouse, 1977; Hillhouse et al., 1985; Irving et al., 1996; Pedder, 2006).

Orogenic Basins in Central Alaska

Along the northern and eastern boundary of Wrangellia lie two flysch basins, the Late Jurassic–Cretaceous Kahiltna basin (Figs. 5 and 39) and the similar age Gravina-Nutzotin+Dezeadesh(?) basin (Fig. 5) described earlier with the Coast plutonic complex of western Canada. In the southern Alaska Range, nearly 6 km of marine clastic rocks occur within the Valenginian–Cenomanian Kahiltna basin (Kalbas et al., 2007). Rocks of the basin occur for nearly 800 km along strike between Wrangellia on the south and Yukon-Tanana and Farewell terranes to the north, but the area is

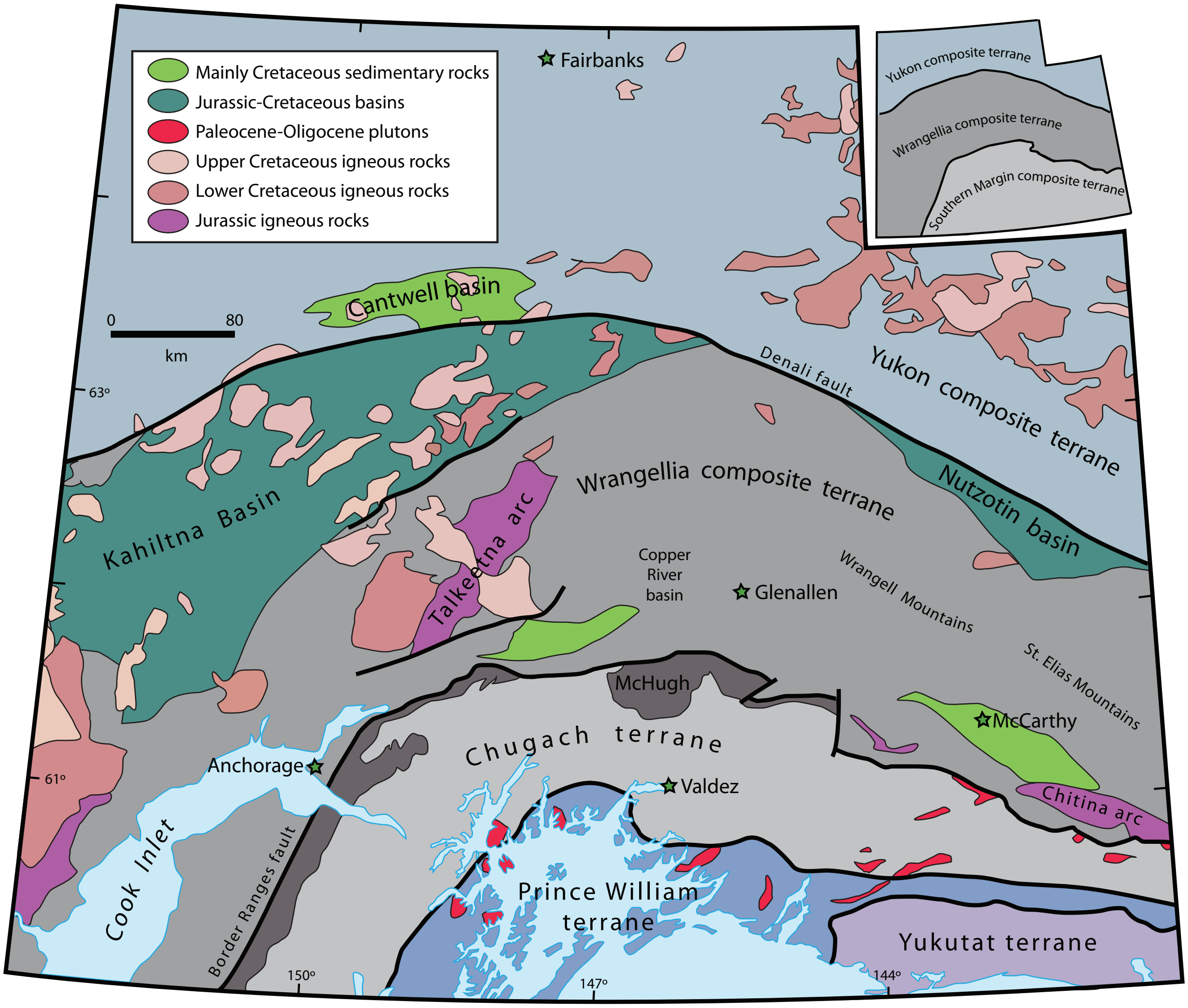


Figure 39. Geological sketch map of south central Alaska showing distribution of some geological features discussed in text. Modified from Trop and Ridgway (2007).

bisected by the Denali fault (Fig. 5). Detailed sedimentological studies, such as facies analysis, detrital zircon dating, and clast counts, suggest that at least 4 km of the sedimentary fill within the basin were deposited during the Aptian–Albian on westerly-dipping axial submarine fan complexes and were derived from both sides of the basin (Kalbas et al., 2007). Sedimentation within the basin ceased and the rocks were deformed and thrust southward between 110 and 90 Ma as the basin closed simultaneously with significant retrogression of rocks in the upper Yukon-Tanana plate (Ridgway et al., 2002). Apparently the collision zone was reactivated during the Campanian–Maastrichtian when the rocks were locally metamorphosed to kyanite-bearing assemblages and intruded by 74 Ma synkinematic sheets of tonalite (Davidson et al., 1992; Ridgway et al., 2002). This deformation was coincident with development of the Campanian–Maastrichtian Cantwell basin (Figs. 5 and 39), a thrust-top basin located just to the north and formed during northward 84–65 Ma folding and thrusting of the Kahiltna rocks, which contain a 60–54 Ma basalt-rhyolite sequence in its upper part (Cole et al., 1999; Ridgway et al., 2002; Trop and Ridgway, 2007).

The Chugach Accretionary Complex

Another group of rocks normally considered to represent a Mesozoic–Tertiary accretionary complex outcrops along the southern part of Alaska, where it is known loosely as the Chugach accretionary complex, or the Southern Margin composite terrane (Fig. 5). The complex is separated from Wrangellian arc terranes by the Border Ranges fault (Smart et al., 1996; Pavlis and Roeske, 2007) and, like the Franciscan, is progressively younger and lower grade outboard and seaward, away from the fault (Plafker et al., 1994; Roeske et al., 2003; Pavlis and Roeske, 2007). Discontinuous, fault-bounded slabs of 200 ± 10 Ma blueschist occur adjacent to the Border Ranges fault in the north (Roeske et al., 1989). Younger mélanges, known either as the Uyak or McHugh complexes depending on location, outcrop farther south and comprise internally complex belts of argillite and graywacke holding blocks and slabs of ophiolitic rocks, such as chert, metabasalt, gabbro, and less common ultramafic blocks (Kusky and Bradley, 1999). Recent detrital zircon studies reveal that the McHugh complex was deposited at two different times: one no older than Oxfordian (157–146 Ma) and the other with maximum ages that range between 91 and 84 Ma (Amato and Pavlis, 2010). Farther outboard lies the Chugach flysch, the most voluminous part of the Southern Margin composite terrane. The flysch was deposited after 85 Ma, but dominantly at around 68–67 Ma during the Maastrichtian (Sample and Reid, 2003; Kochelek and Amato, 2010; Kochelek et al., 2011), and so there might be a depositional gap as large as 15 Myr in sedimentation within the accretionary prism.

Based on paleomagnetic, isotopic, and provenance data, the flysch was probably derived from the Coast plutonic complex (Farmer et al., 1993; Sample and Reid, 2003; Kochelek et al., 2011), which is presently located mostly in British Columbia, but at the time of deposition was located much farther south. This finding was supported by Roeske et al. (2003), who matched a distinctive 170 Ma intrusive suite in the Wrangell Mountains with the West Coast intrusive suite located on the western part of Vancouver Island, and documented that the northward transport may have begun as early as 85 Ma but was more or less continuous from 70 to 51 Ma.

Even farther seaward are Paleocene–Eocene flysch belts of the Prince William terrane (Fig. 5), which are interbedded with basaltic lavas—the Ghost Rocks Formation—on Kodiak Island located today at ~55° N (Moore et al., 1983). The lavas yielded paleolatitudes from ~48 to 41° N both in original work (Plumley et al., 1983) and in a recently completed study (Housen et al., 2008; Roeske et al., 2009). A suite of intrusions, termed the Sanak-Baranof plutonic belt (Hudson et al., 1979) intruded the accretionary rocks at 61 Ma in the west, migrated eastward until 51 Ma, and is broadly interpreted to be the product of ridge subduction (Bradley et al., 2003b).

DISCUSSION

Ever since Monger and Ross (1971) recognized two different fusulinid populations within the Canadian Cordillera, and suggested that some areas might be far traveled, there have been many attempts to unravel the tectonic collage of the North American Cordillera. Perhaps the most influential overall was the contribution by Coney et al. (1980), who pointed out the great diversity of Cordilleran terranes and suggested an accretionary history for them. Particularly important within the western United States were the early papers by Burchfiel and Davis (1972, 1975; Davis et al., 1978). Another early attempt—quite sophisticated for its era—to tie the various terranes within the Canadian Cordillera together into larger superterranes was the effort by Monger et al. (1982). They suggested that two previously amalgamated superterranes docked with the North American margin at different times. In their scheme they gathered the East (Slide Mountain), Quesnel, Cache Creek, and Stikine terranes into the Intermontane superterrane, which docked with the North American margin during the Jurassic, and the Alexander, Wrangell, and Gravina-Nuztotin terranes into the Insular terrane, which arrived during the Cretaceous. Other smaller terranes, such as the Bridge River terrane, caught between the Intermontane and Insular superterranes, or the outboard Chugach terrane, may have arrived separately. They also recognized that the Omineca belt and Coast plutonic complex are two separate plutonic belts in which many plutons intruded previously amalgamated terranes. The summaries by Oldow et al. (1989), Saleeby (1983), Saleeby and Busby-Spera (1992), and Coney and Evenchick (1994) were exquisite and timely overviews, but accepted the basic Cordilleran model as a starting point, so are quite different from the analysis presented here. The fundamental difference between this and earlier syntheses is that in a general sense the older contributions presented accretionary models in which exotic terranes were progressively added to North America, whereas I argue that the majority of terranes were assembled offshore into an enormous ribbon continent, which collided with North America, initially during the Sevier orogeny at ~124 Ma, but ultimately during the ~80–75 Ma Laramide orogeny. In this regard the original Jurassic–Cretaceous model of Moores (1970) in which he suggested the Cordillera was generated by continent-arc collision above a westward-dipping subduction zone, was especially prescient, as were more local models by Mattauer et al. (1983), Templeman-Kluitt (1979), and Chamberlain (Chamberlain and Lambert, 1985; Lambert and Chamberlain, 1988). Readers of this paper will find the ideas of Stephen Johnston (2001, 2008), which are similar to my own, but developed independently, to be insightful and worth careful study.

In what follows, I present a summary of the major events in the assembly of the Cordillera based on data presented in earlier sections. Many uncertainties exist and by no means should this synthesis be considered a final document, even if gleaned from information written in stone! Hopefully, it will serve as a guide to focus future research on key topics and areas.

Paleozoic Events within the Great Basin and Canadian Sectors

During the Early Mississippian, lower Paleozoic chert-shale sequences and scattered alkali basalts of the Roberts Mountain allochthon (Fig. 40) were complexly faulted and folded as they were emplaced upon the western margin of the Antler platform during what is known as the Antler orogeny (Roberts et al., 1958; Nilsen and Stewart, 1980; E.L. Miller et al., 1992; Poole et al., 1992). A foreland basin filled with coarse terrigenous clastics developed to the east of the allochthon (Gehrels and Dickinson,

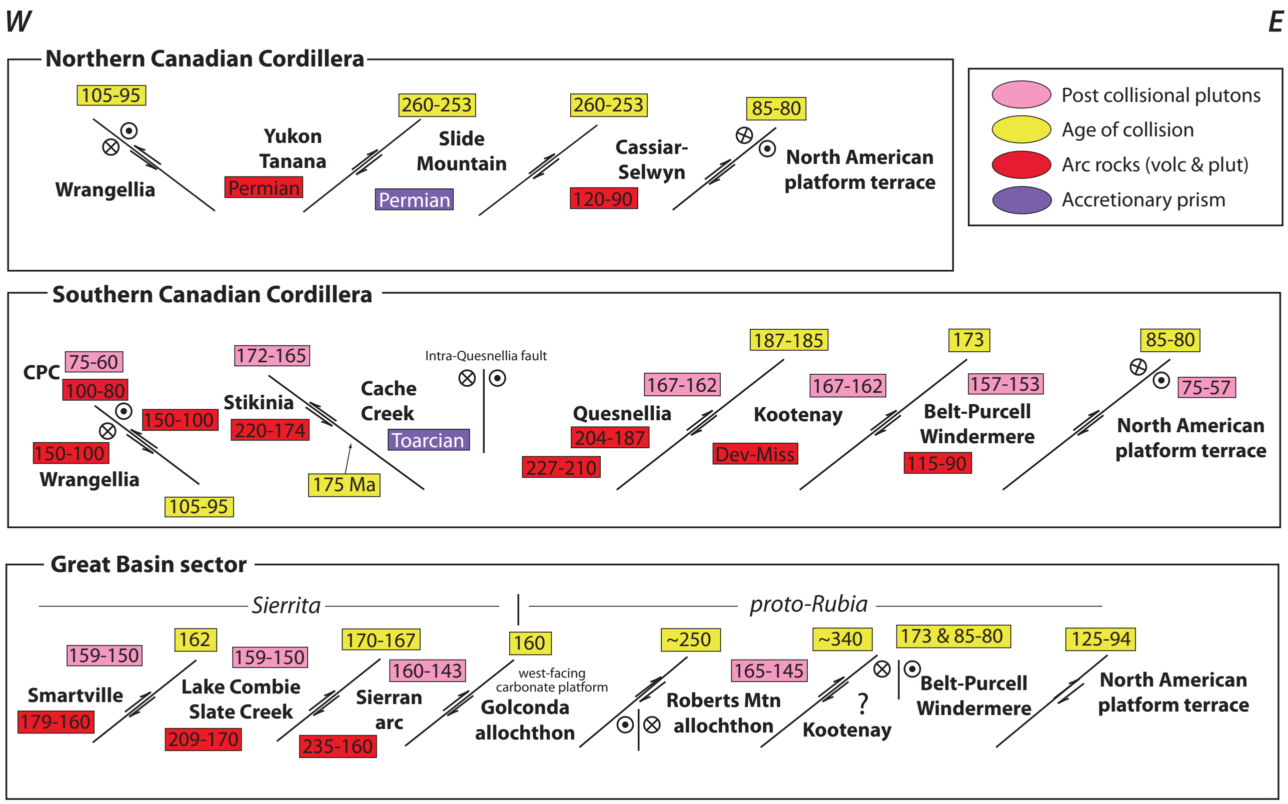


Figure 40. Schematic cross sections through three sectors of the Cordillera illustrating the ages of pre-collisional magmatism, the collision, and postcollisional, typically slab-failure, magmatism. Based on the data summarized here, there is no evidence to support a progressive accretion of terranes to western North America.

2000) but never reached the North American platform, less than 100 km away. While some workers (Finney and Perry, 1991; E.L. Miller et al., 1992; Poole et al., 1992) argued that rocks of the Roberts Mountain allochthon are simply fine-grained off-shelf deposits of the North American margin, others, most notably Ketner (1968, 1977, 1991), argued that the coarse, immature nature of many of the Paleozoic sediments demanded a western source as North America was rimmed by carbonate banks and covered by epeiric seas for much of this time. Wright and Wyld (2006) nicely summarized the various sedimentological arguments, as well as more modern detrital zircon studies, and concluded that the rocks of the Roberts Mountain are part of a larger family of terranes, including the Shoo Fly complex, Yreka-Trinity complex, and Alexander terrane, of Gondwanan provenance, and were all transported from the Ligerian ocean into the Panthalassic ocean during the Devonian (see also: Grove et al., 2008b). Whatever the original provenance of the Roberts Mountain allochthon, the absence of arc volcanism and plutonism suggests that its westernmost parts, including its inferred upper Paleozoic arc, were removed after its accretion to the Antler platform yet prior to arrival of the Golconda allochthon just to the west (Fig. 5). One possibility is that the arc rocks occur today within southern Canada's Kootenay terrane, which has many similarities with the Roberts Mountain allochthon and contains a Devonian–Mississippian arc within the Eagle Bay assemblage (Schiarizza and Preto, 1987; Smith and Gehrels, 1992b; Paradis et al., 2006). Another possibility is the Yukon-Tanana terrane, which also contains Devonian–Mississippian arc rocks (Colpron et al., 2006). Burchfiel and Royden (1991) argued that the absence of an arc was not a problem if the Antler orogeny represented an Apennine-type collision, where the arc was severely dismembered and actively subsiding, but given the mobility of terranes along the Cordilleran margin, it is worth examining some of the translated arcs elsewhere in the Cordillera.

Rocks of the Golconda allochthon are complexly deformed marine sedimentary and volcanic rocks ranging in age from latest Devonian to Permian (Silberling and Roberts, 1962; E.L. Miller et al., 1992). In most models they were emplaced upon rocks of the Roberts Mountain allochthon (Fig. 40) along west-dipping thrust faults during the Late Permian–Early Triassic Sonoman orogeny (Burchfiel et al., 1992; E.L. Miller et al., 1992). Two problems stand out: (1) there are no Permian arc rocks in the area; and (2) as pointed out by Wright and Wyld (2006), there is no evidence of any foredeep to the east. Mississippian and Permian arc rocks do occur within the Yukon-Tanana terrane (Colpron et al., 2006) located today in the Canadian and Alaskan sectors, and it is possible that this terrane represents the arc portion of the Golconda allochthon. The lack of any foredeep is difficult to overcome in any sort of collisional model, which led Wright and Wyld (2006) to develop a basinal model for the Golconda. In any case, rocks of the combined Golconda–Roberts Mountain allochthons were united and overlain by a west-facing carbonate platform by the Triassic (Wyld et al., 2001).

As stated in the descriptive section of the paper, Wrangellia and Alexander terrane were stitched together by a pluton dated to be 309 ± 5 Ma (Gardner et al., 1988). Therefore, these terranes developed in close proximity to one another during the late Paleozoic and Mesozoic.

The amalgamation of the Yukon-Tanana and Slide Mountain terranes with the Cassiar platform–Selwyn basin block is one of the major events of the Canadian Cordillera and was recognized by Templeman-Kluitt (1979), who argued that the western edge of North America was subducted westward beneath the arc-bearing Yukon-Tanana terrane with oceanic and accretionary prism rocks of Slide Mountain caught between the two. Yukon-Tanana terrane comprises Devonian metamorphic and polydeformed basement cut by Late Devonian–Early Mississippian plutons and overlain by Mississippian–Permian arc rocks (Mortensen and Jilson, 1985; Colpron et al., 2006). Between 260 and 253 Ma (Lopingian) the western margin of the Cassiar platform was pulled down beneath Yukon-Tanana terrane (Fig. 40) on the Inconnu thrust in an abortive subduction attempt (Murphy et al., 2006; Beranek and Mortensen, 2011). Rocks of the oceanic Slide Mountain terrane were telescoped and sit structurally between the two terranes (Nelson, 1993).

Rocks of the Slide Mountain terrane contain giant fusulinids that are known only from a few locales: Kettle Falls, Washington; the eastern Klamaths; Sonora, Mexico; and an autochthonous locale from the miogeocline in West Texas, all of which indicate that rocks of Slide Mountain terrane are now far north of their warmer-water zones of origin (Carter et al., 1992). The fossil data are supported by paleomagnetics, which indicate at least 2000 km of northward movement (Richards et al., 1993).

Neoproterozoic to Paleozoic rocks of the Selwyn basin are generally considered to represent fine-grained off-shelf deposits of the North American passive margin with an elevated rim upon which Late Silurian–Middle Devonian Cassier platform developed (Gabrielse et al., 1973; Cecile, 1982; Gordey and Anderson, 1993), but they have much in common with exotic rocks of the Roberts Mountain allochthon in that they are cherty, fine-grained sedimentary rocks with sporadic alkali basalt throughout the section, extensive barite beds, and localized sedimentary-exhalative deposits (Goodfellow et al., 1995). Rocks of the basin are allochthonous and were transported over the North American margin after the Late Jurassic but prior to 104 Ma as a huge composite allochthon along the Dawson and Broken Skull faults and have no rocks in common with their footwall (Mair et al., 2006; Gordey and Anderson, 1993). To the south, rocks of the basin were folded into south-verging overturned nappes and intruded by carbonatite during the Devonian (McLeish et al., 2010; McLeish and Johnston, 2011), an event unknown from the North American platform to the east.

Upper Triassic to Middle Jurassic Arcs and Collisions

Upper Triassic to Middle Jurassic arc terranes are common throughout the Cordillera with some built on oceanic crust, others built on older sialic basement, and some spanning both. These include the Talkeetna, Kootenay, Stikinian, Quesnellian, Black Rock, several Klamath and Sierra Nevadan arcs, and poorly known arcs of the Sonoran sector. Each segment is unique, but in some cases there are enough similarities to suggest that they were formerly connected. In poorly exposed or little studied areas, the youngest magmatic date probably approximates the time of accretion.

Talkeetna Arc

In south-central and southwestern Alaska, the Talkeetna arc is exposed over a strike length of over 1000 km within the Peninsular terrane of the Wrangellian superterrane (Fig. 5). The main phase of Talkeetna arc magmatism occurred from 202 Ma to 175 Ma and plutons intruded in the northern part of the area between 190 Ma and 153 Ma (Rioux et al., 2007). In the west on Kodiak Island and the Alaskan Peninsula the Jurassic batholith yielded ages of 213 Ma on Kodiak Island and 184–164 Ma on the Alaskan Peninsula (Rioux et al., 2010). Basement to the arc varies along and across strike as documented by isotopic data and xenocrystic zircons (Clift et al., 2005; Greene et al., 2006; Rioux et al., 2007; Amato et al., 2007). To the north of the arc on Wrangellia are south-dipping thrust faults that developed in consort with the Oxfordian to Tithonian Nutzotzin–Wrangell Mountains basin, which sits unconformably upon platformal carbonates and coarsens upward from marine mudstones and sandstones to conglomerate (Trop et al., 2002; Manuszak et al., 2007). This basin is generally considered to represent a back-arc basin to the Talkeetna arc (Clift et al., 2005; Trop and Ridgway, 2007; Manuszak et al., 2007), but given that the arc appears to have been thrust upon the carbonate platform, I suggest that it might instead have developed as a typical foredeep and subsequent foreland fold-thrust belt related to ~170 Ma collision between the Talkeetna arc and Wrangellia. The emplacement of the arc upon Wrangellian crust could explain the lack of pre–170 Ma plutons cutting Wrangellia; the extreme tectonic thinning of the arc, from 25 to 28 km to 7 km within 10 Myr of the collision, as deduced from thermobarometry and Ar dating (Hacker et al., 2008; Rioux et al., 2007); the northward younging of arc magmatism; and the Oxfordian to Tithonian Naknek basin, which is a coarse-clastic, debris-filled basin that formed along northward-dipping reverse fault scarps located along the south side of the arc (Trop et al., 2005). Postcollisional plutons (<170 Ma) as young as 153 Ma (Rioux et al., 2007) of the area might then represent slab-failure magmas.

Bonanza Arc

In contrast to the Talkeetna arc of Alaska with which it is commonly correlated, the Bonanza arc of southern British Columbia appears to have developed upon Wrangellian crust between ~202 and 165 Ma and comprises up to 2500 m of interbedded lava, tuff, and breccia ranging in composition from basalt to rhyolite cut by plutons ranging in age from 190 to 169 Ma (Isachsen, 1987; DeBari et al., 1999; Canil et al., 2010). The plutons range in composition from gabbro-diorite mainly to tonalite and granodiorite, and they appear to intrude rocks of Wrangellia beneath the arc rocks (Nixon et al., 2011a, 2011b, 2011c, 2011d). Given those relationships, it is unclear why arc magmatism ceased at ~165 Ma.

Quesnellia-Kootenay-Belt-Purcell-Windermere

These three terranes were joined by collision during the Jurassic and display a remarkably clear temporal progression of eastwardly migrating magmatism and accretion (Fig. 40). Quesnellia, the westernmost of the three, contains abundant Late Triassic–Lower Jurassic volcanic rocks that are as young as Toarcian (Tipper, 1984). It also contains abundant plutons in the age range 212 to 204 Ma (Armstrong, 1988). In the north, rocks of the Cassiar platform were pulled beneath the eastern edge of Quesnellia at 186 Ma on west-dipping thrust faults (Nixon et al., 1993); whereas to the south, the western edge of Kootenay terrane was pulled beneath Quesnellia between 187 and 185 Ma to form an eastward-vergent fold-thrust belt (Murphy et al., 1995; Colpron et al., 1996, 1998). Early southwest verging structures such as the Scrip nappe were overprinted by northeast verging folds and thrusts between ~173 and 168 Ma and were intruded by the syntectonic Kuskanax batholith (Fig. 32) at 173 Ma (Parrish and Wheeler, 1983). The second phase of deformation corresponds to the attempted westward subduction of the Belt-Purcell-Windermere (BPW) block beneath Kootenay terrane. One can easily see the age progression in plutonic rocks on Figure 32, where Quesnellia contains abundant Early Jurassic plutons, absent from Kootenay and BPW; whereas both Quesnellia and Kootenay contain abundant Upper Jurassic plutons, but BPW contains only a few tiny stocks of that age; and a younger period of 115–90 Ma plutons known as the Bayonne Suite (Logan, 2002). An even younger suite of latest Cretaceous–early Tertiary plutons cut all three terranes and was attributed by Hildebrand (2009) as slab-failure magmas related to the failure of the North American plate during the Laramide orogeny. Muscovite- and garnet-bearing plutons in the age range 157 to 153 are clearly postcollisional, as well as the 159 Ma biotite granite of the Nelson batholith (Fig. 32) (Armstrong, 1988; Sevigny and Parrish, 1993). Both groups probably represent slab-failure magmas related to the Kootenay-BPW collision; whereas other variably deformed plutons in the age range 167 to 162 Ma (Ghosh, 1995) might be slab-failure bodies related to the failure of the Kootenay plate. Sorting out which bodies are subduction related and which are slab-failure related will be an arduous task given the short duration between the two collisions. The doubly-vergent fan structures characteristic of the Kootenay terrane (Wheeler, 1963; Brown and Tippett, 1978; Price, 1986; Brown and Lane, 1988; Colpron et al., 1996, 1998) might be related to wedging by Quesnellia as the narrow terrane was caught between Quesnellia and the BPW block during the second collision.

An interesting point to consider is the similarity in ages for the Kootenay-BPW collision and the deformation in the hinterland belt of the Great Basin, where deformation, metamorphism, and the formation of large recumbent nappes took place at 165–160 Ma (Snoke and Miller, 1988; Miller and Hoisch, 1995; Zamudio and Atkinson, 1995; Camilleri et al., 1997; McGrew et al., 2000) and was termed the Elko orogeny by Thorman (2011). As the rocks immediately to the east of the hinterland belt are thick Neoproterozoic clastic successions similar to the Windermere, it seems possible that slivers of Kootenay terrane might be caught in the collision zone of the Great Basin hinterland.

Black Rock Arc

The latest Triassic–Early Jurassic Black Rock arc terrane of northwestern Nevada (Quinn et al., 1997; Wyld, 2000) was the upper plate in a 163–160 Ma collision between it and the west-facing carbonate-dominated passive margin of western Rubia. The collision zone is marked by the east-vergent Luning-Fencemaker fold-thrust belt (Wyld et al., 2003; Wyld and Wright, 2009). A collapsed clastic basin (Burke and Silberling, 1973; Speed, 1978) occurs between the arc and the Rubian margin. Based upon interfingering relationships and the occurrence of ultramafic to granitic intrusions, basaltic pillow lavas and intermediate composition lavas, breccias and tuff (Dilek and Moores, 1995), the basinal facies rocks appear to have been proximal to the arc. The rocks of the basin could thus represent a deformed fore-arc basin. As far as I'm aware there are no direct data that constrain the width of the ocean between the arc and the western passive margin of the Rubian superterrane. Subduction within this ocean was clearly westward beneath the Black Rock arc (Fig. 41), and if that arc is correlative with the Sierra Nevada as generally accepted, it strongly suggests that subduction was also westward beneath that Jurassic arc as well.

Klamath Arcs

Several Jurassic arc terranes occur within the Klamath Mountains. The easternmost Triassic to Middle Jurassic arc, which was built upon older basement of Redding subterrane is here considered to be part of a much larger arc terrane extending from Nevada through the Sierra Nevada to Mexico and will be discussed with that arc in a subsequent section.

The 177–168 Ma Hayfork arc in the Klamath Mountains block is part of the composite Hayfork terrane (Fig. 19), which comprises an eastern belt of Permian–Triassic mélange and a western arc complex including volcanic rocks and 170 Ma plutons, which were accreted to the western margin of the North Fork terrane between 169 and 164 Ma (Wright and Fahan, 1988). Although the current fault between the Hayfork and North Fork terranes dips eastward, which led most workers to think that the Hayfork arc was subducted to the east beneath the North Fork (Harper and Wright, 1984; Irwin, 1981, 2003; Irwin and Wooden, 2001), I suggest that the fault is either back-rotated or a younger

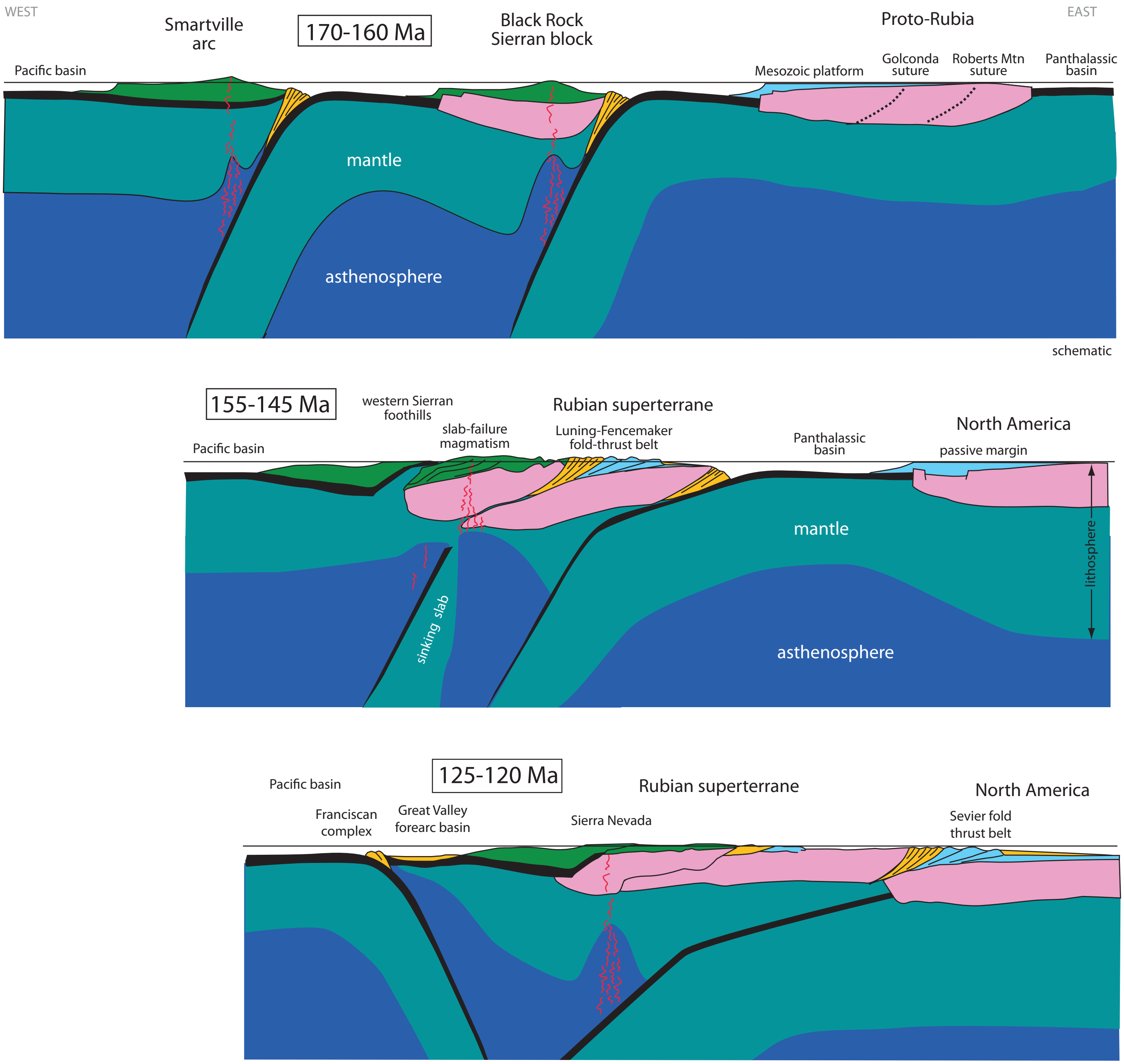


Figure 41. Plate model cartoon illustrating the tectonic model proposed in the text. Note that in the western part of the 170–160 Ma panel that the Slate Creek–Lake Combie arc is not shown as it would basically be a duplicate of the Smartville arc.

fault as there is no arc magmatism of the correct age east of the fault, which suggests that subduction was to the west beneath the Hayfork arc. In this hypothesis, the Hayfork terrane represents an east-facing arc-accretionary prism complex beneath which the more easterly terranes of the Klamath Mountains were thrust. As documented by Wright and Wyld (1994), the arc was built upon a substrate consisting of a Triassic accretionary complex, suggesting that the arc had migrated eastward over older parts of its own accretionary complex due to eastward slab rollback.

The 164–162 Ma Josephine ophiolite (Harper et al., 1994), which lies west of the Hayfork arc (Fig. 19), would represent oceanic crust of a back-arc basin that was apparently opening during, or just prior to, the more easterly Hayfork-Northfork collision (Wyld and Wright, 1988), much in the same way the 168–161 Ma Coast Range ophiolite seems to have formed behind the Smartville arc just prior to its 159 Ma collision in the western Sierra foothills. The possible remnant arc to the west, known as the Rogue Valley terrane, formed during the Late Jurassic and comprises thick successions of andesitic pyroclastic rocks, breccias, and lava flows that interfinger with the Galice flysch (Garcia, 1979, 1982), both of which are older than 150 Ma, the age of dacitic dikes that cut both units (Saleeby, 1984). This arc fragment is probably part of a more extensive arc found in the westernmost Sierra Nevada and represented by volcanic rocks that interfinger with the Mariposa Formation, which is interpreted to be a Galice equivalent (Bogen, 1985; Snow and Ernst, 2008; Ernst et al., 2009b).

The Ironside batholith (Charleton, 1979) and associated 171–168 Ma mafic-intermediate plutons (Fig. 19) intrude the Hayfork terrane; lack crustal input; postdate thrusting of the West Hayfork over the East Hayfork mélange; predate the accretion of the Hayfork with the Central terranes; and represent a poorly explained pulse of plutonism (Wright and Fahan, 1988; Barnes et al., 2006a). By interpreting the Hayfork arc as the upper plate in the collision with rocks to the east, the plutons are readily interpreted to have formed due to an increased influx of rise-slope sediment on the descending plate and its subsequent dewatering just prior to collision, which would cause larger amounts of melting within the mantle wedge and an increased magmatic flux within the arc. This is the mechanism Hildebrand (2009) envisioned to create Cordilleran-type batholiths in general, but in this instance the margin on the lower plate was young and so the width of the borderlands zone was likely much narrower than that of the rifted and long-lived western North American margin; thus, the pulse of magmatism was much shorter.

Several suites of postcollisional Jurassic plutons intrude rocks of the Klamaths. The oldest suite ranges in age from 162 Ma to ~156 Ma (Fig. 19), intruded the Hayfork and more easterly terranes, and ended prior to accretion of the western Klamath arc terrane (Harper and Wright, 1984; Harper et al., 1994). A younger suite ranges in age from 151 to 144 Ma (Fig. 19), was generated by an influx of primitive, mantle-derived H_2O rich basalt into the crust, and also intrudes several previously sutured terranes (Allen and Barnes, 2006; Barnes et al., 2006b). Both suites are poorly explained in existing models, but are easily understood as typical slab-failure magmas because both groups clearly postdate their respective collisions and are not confined to the upper-plate arcs but instead intrude across several terranes. The cause and some more general features of slab-failure magmatism are discussed in a subsequent section. An earliest Cretaceous (142–136 Ma) suite that intrudes rocks in the central part of the Klamaths (Fig. 19) also occurs in the western Sierran foothills (Figs. 17 and 18) (Irwin and Wooden, 2001; Irwin, 2003), and although they represent a short-lived pulse of magmatism and are more likely related to slab failure, it is possible that they were the initial products of eastwardly-dipping subduction that generated the Franciscan complex.

Lake Combie–Slate Creek Arc

Within the western Sierra Nevada foothills, rocks of the Slate Creek–Lake Combie arc range in age from 210 to 172 Ma and were thrust over the partly coeval Fiddle Creek complex (Fig. 18), which contains serpentinite-matrix, ophiolitic mélange overlain by Triassic–Early Jurassic pillowed basalt, volcaniclastics, argillite, and radiolarian chert (Edelman et al., 1989a; Fagan et al., 2001; Moores and Day, 1984; Day and Bickford, 2004). The contact of the two complexes was intruded by the 167 Ma Scales pluton, which provides a minimum age for deformation (Day and Bickford, 2004). Amphibole cooling ages fall in the range 156 to 152 Ma (Fagan et al., 2001), and the region was cut by a group of postkinematic 159–150 Ma plutons (Day and Bickford, 2004). Arc-related volcanic and plutonic rocks the same age as the Slate Creek–Lake Combie do not occur to the east, and much like the Hayfork, with which these rocks are commonly correlated (Fagan et al., 2001; Irwin, 2003), subduction was probably westwardly directed beneath the arc. In this scenario the Fiddle Creek complex represents a collapsed fore-arc basin-accretionary prism complex associated with the Smartville-Combie arc. The swarm of 159–150 Ma plutons that intrude the region (Figs. 17 and 18) probably represent slab-failure magmatism.

Smartville Arc

The farthest outboard terrane in the Sierran Foothills belt contains thick sequences of Jurassic arc rocks that sit upon bits of oceanic crust (Menzies et al., 1980; Bogen, 1985; Day et al., 1985). The youngest volcanic rocks in the complex were dated as 159 Ma (Saleeby, 1981), just older than the Yuba Rivers pluton (Saleeby et al., 1989), which is cut by the thrust between the Smartville and Slate Creek–Combie belts but metamorphosed rocks to the west (Figs. 17 and 18); so the collision of the Smartville block with rocks located to the east occurred at 159 Ma, and magmatism within the arc continued up to the time of collision. Exhumation of the Slate Creek complex was already well under way by 156 Ma (Fagan et al., 2001).

Most workers since Moores (1970) have recognized that the east-facing oceanic Smartville arc, which rose above a westerly-dipping subduction zone, collided with the western margin of the Foothills terrane (Dickinson et al., 1996a, 1996b; Moores and Day, 1984; Godfrey and Dilek, 2000). Thus, for some period before the collision at 159 Ma, any magmatism to the east could not have been generated by eastward subduction unless there was another subduction zone. Consequently, several workers (Schweickert and Cowan, 1975; Schweickert, 1978; Schweickert et al., 1984; Ingersoll and Schweickert, 1986; Dickinson et al., 1996b; Ingersoll, 2008) proposed Moluccan sea–type models, in which bipolar east-west subduction above a continental arc to the east and an oceanic arc to the west, led to collision between the two. A similar scenario was recently developed for the 159–154 Ma collision of the Olds Ferry and Wallowa arc located to the north in the Blue Mountains of Oregon (Schwartz et al., 2011). Yet, another possibility is that subduction was westward along the eastern margin of the Black Rock–Sierran block so that there were two westward-dipping subduction zones (Fig. 41).

In either of the models, the Coast Range ophiolite is readily interpreted as an ophiolite formed to the west of the Smartville complex in a back-arc, suprasubduction, setting prior to docking of the upper-plate Smartville arc at 159 Ma (Shervais, 2001; Shervais et al., 2004, 2005). At 168–161 Ma, it could have formed in equatorial latitudes some distance from any land and in a back-arc setting, yet still satisfy the C.A. Hopson et al. (2008) open sea model, which is based largely on the existence of an oceanic volcanopelagic sedimentary veneer and the products of a Late Jurassic (152–144 Ma) disruption and/or brecciation capstone sequence. They suggested (C.A. Hopson et al., 2008) that the ophiolite must have passed near enough to the Americas to receive distal ash transported from the NE by trade winds, but Hildebrand (1988) showed that ash is generally transported from west to east by winds in the troposphere, not the low level winds such as trades. Thus, there may have been another arc located to the west of the ophiolite as there was in the case of the correlative Josephine ophiolite.

The peculiar breccias that sit atop the ophiolite and appear to be younger than 154 ± 5 Ma (Blake et al., 1987) might have resulted from deformation during the ongoing collision when the sea floor to the west of the Smartville arc buckled and broke, much in the way the sea floor of the Indian Ocean deformed during the Indian-Eurasian collision in the Himalayas (Beekman et al., 1996). C.A. Hopson et al., (2008) argued that the breccias formed due to grinding in a transform fault zone, and this might be partially correct and the breccias still generated during collision, for the deformation within the Indian sea floor apparently not only led to pervasive reverse faults throughout the area of buckled sea floor (Weissel et al., 1980; Zuber, 1987), but reactivated older transform faults as well (Bull, 1990; Bull and Scrutton, 1990, 1992).

The Smartville and associated terranes within the Sierran foothills area are apparently relatively thin and emplaced atop continental crust because the area was cut at 140 Ma by a suite of large granitoid plutons (Saleeby et al., 1989; Irwin and Wooden, 2001; Day and Bickford, 2004; Figs. 17 and 18) and for those areas that were accreted earlier, a different suite of plutons, dated at ~160 Ma, intruded those terranes (Edelman et al., 1989a). That the exotic terranes are thin and have continental crust beneath them is supported by the presence of the plutons themselves, by geophysical models derived from seismic-refraction data, as well as gravity-magnetic data and tomographic inversion of earthquake data that require a less dense lower crust beneath mid to upper crustal mafic-ultramafic slab regarded to be part of a large ophiolitic slab—possibly the link between the ophiolite of the Coast Ranges and the Smartville block (Stanley et al., 1998; Godfrey and Dilek, 2000). Precambrian inherited zircons within both deformed and post-tectonic intrusions, such as the Yuba Rivers, Deer Creek, and Smartville plutons (Fig. 17) provide additional evidence for the presence of old continental crust beneath the tectonic flakes (Day and Bickford, 2004). It is important to note that the provenance of the continental crust beneath the Foothills belt is entirely unknown and lies west of the 0.706 isopleth, generally interpreted as the western edge of old cratonic crust (Moore, 1959; Kistler and Peterman, 1973; Chen and Tilton, 1991).

The observation that the western oceanic arc terranes sit atop continental crust is important because it likely defeats the Moluccan sea–type model in which opposed eastward and westwardly-dipping subduction zones occurred between the Smartville and Sierran blocks. In a doubly-vergent Moluccan type situation, neither arc is attached to a subducting slab; so, other than plate motions, there is no driving force to subduct one arc beneath the other; instead, these arc-arc collisions are soft (Pubellier et al., 1991). Thus, a model in which there are two westward-dipping subduction zones, one east of the Sierran block, and the other west, seems the most reasonable. Rocks in western Nevada support such a model because there the Luning-Fencemaker thrust belt represents westward subduction of a westerly facing marginal platform beneath a Triassic–Middle Jurassic arc terrane just after 163 Ma, the age of the pre-deformational Humboldt complex (Wyld et al., 2001; Dilek and Moores, 1995).

Thus, an implication of the Smartville and Slate Creek–Combie accretionary events is that any magmatism prior to the collision cannot be related to the younger Cretaceous magmatism in the Sierra because they were derived from entirely different subduction zones. The style and composition of magmatism between ~158 Ma and 145 Ma was very different from that before or afterwards. Magmatism during this period, which is a transitional magmatic event, includes the 5–600-km-long, 148 Ma Independence dike swarm and various small plutons, such as those in the Goddard pendant and 152–148 synkinematic dikes of Owens Mountain (Wolf and Saleeby, 1995), within the Sierra and possibly 155–150 Ma dioritic plutons in the Mojave region (Miller and Glazner, 1995). It is poorly explained in current models.

Sierran Jurassic Arc

Although it is much broken and distorted by younger strike-slip and normal faults, and masked by Cretaceous–Tertiary plutons, the Jurassic Sierran arc may be part of one continuous arc from the Redding subterrane of the eastern Klamaths and northern Sierra Nevada south-southeastward through the White-Inyos, Mojave and northern Sonoran deserts to the Gulf of Mexico (Busby-Spera, 1988; Barton et al., 1988; Busby et al., 2002; Tosdal et al., 1989; Mauel et al., 2011). Based on fragmentary exposure and dating, the arc was apparently active from the Triassic to 160 Ma, when the arc was deformed and arc magmatism ceased (Fig. 42). Subsequent magmatism, generally bimodal and alkaline, is perhaps best viewed as slab-failure magmatism, and will be discussed in the following sections. Despite the large number of studies on the Jurassic arc in the Sierra and Mojave regions, the Sonoran region of southwestern Arizona and extreme southeastern California might be the best exposed and understood part of the arc. There, plutonic rocks of the Lower to Middle Jurassic Kitt Peak–Trigo Peaks–Cargo Muchacho super units and their volcano-sedimentary host rocks (Fresnal Canyon sequence) were metamorphosed and deformed just before 158 Ma, as demonstrated by U-Pb dating of deformed and non-deformed igneous units (Tosdal et al., 1989). These authors also

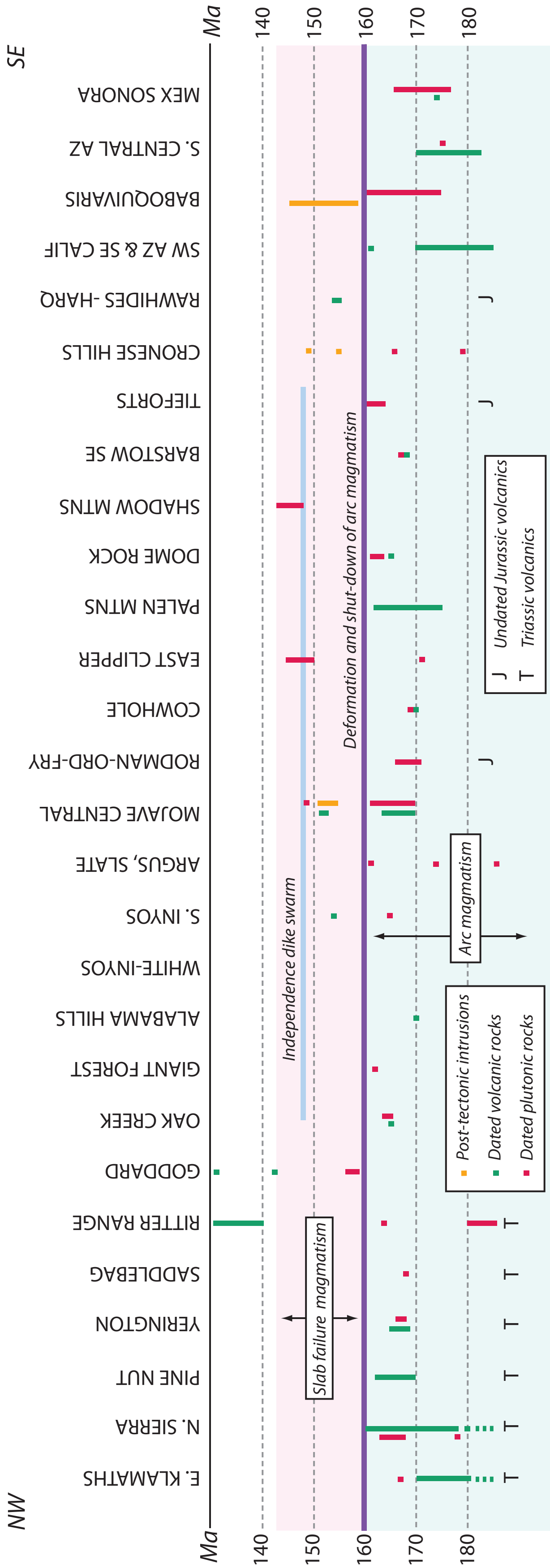


Figure 42. Schematic diagram summarizing the ages of well-dated magmatic rocks of some areas in the Klamath–Black Rock–Sierran arc. As discussed in the text, everywhere along strike the arc was deformed during the Sierrita-Rubia collision at 160 Ma coincident with shutdown of arc magmatism. New postcollisional magmatism started soon after the collision and is interpreted here as slab-failure magmatism. Data from sources listed in text.

showed that Upper Jurassic rocks of the lower-grade and dominantly alkaline Artesia sequence sit atop the deformed and metamorphosed Fresnal–Kitt Peak rocks, and were cut by the bimodal and largely alkaline Ko Vaya plutonic suite (Fig. 31). Similarly, in the Dome Rock Mountains (Fig. 21), 164 Ma granodiorite and deformation are cut by 161–158 Ma nondeformed leucogranite (Boettcher et al., 2002). Vestiges of the two periods of magmatism, arc and slab failure, can be found all along the arc from the northern Sierra to Sonoran Mexico (Fig. 5) and thus a coherent story of Triassic to Jurassic arc magmatism, collision, and shutdown of arc magmatism at ~160 Ma, followed by a period of slab-failure magmatism, emerges.

The most vexing issue with this arc is its polarity, which is generally assumed to be west-facing (Burchfiel and Davis, 1972, 1975; Schweickert and Cowan, 1975; Busby-Spera, 1988; Ingersoll, 2008). The arc and its deformation are the same age as the Black Rock arc of northwestern Nevada, which was the upper plate in a 160 Ma collision with a more easterly carbonate passive margin (Dilek et al., 1988). Based on age of magmatism and terminal collision, it could represent a more northern continuation of the same arc (Barton et al., 1988). If so, then it implies that the Jurassic Sierran arc was east facing and not west facing as generally assumed. While the collisional belt between the arc and the eastern lower plate was recognized in Nevada, where it is known as the Luning-Fencemaker fold-thrust belt (Wyld et al., 2003; Wyld and Wright, 2009), it is poorly known farther south, where it is known as East Sierran fold-thrust belt (Stevens et al., 1998). However, fragments of the fold-thrust belt are known from the Saddlebag Lake area southeastward through the Inyo Mountains–Mojave Desert region and it is characterized by southwestward-dipping, northeastward-vergent thrusts that postdate the main phase of Jurassic arc magmatism, but predate the 148 Ma Independence dike swarm (Dunne et al., 1978; Dunne, 1986; Walker et al., 1990b;Gerber et al., 1995; Howard et al., 1995; Stevens et al., 1998; Miller and Walker, 2002; Martin et al., 2002; Dunne and Walker, 2004; Stone et al., 2009). It is also preserved in the Maria fold-thrust belt of Arizona (Fig. 8) where 175–160 Ma volcanic rocks were deformed and thrust northward before deposition and eruption of 154 Ma sedimentary and volcanic rocks (Reynolds et al., 1986; Richard et al., 1987; Spencer et al., 2011).

The possibility of eastward subduction is difficult to evaluate because the only place to put an eastward-dipping subduction zone at that time was in a basin in the middle of the Sierra as there were two westerly-dipping collisions, the Slate Creek–

Combie and Smartville arc collisional events, which occurred to the west at 167 Ma and 159 Ma, respectively. The width of the basin, if there was one, is unknown because it closed during the subsequent 105–100 Ma transpressional event (discussed in a subsequent section) between the eastern and western Sierra. Also, it is not even known whether there was ever oceanic lithosphere in the basin. A less convincing piece of evidence, but perhaps no less important, is the observation that contacts within the Western Sierran foothills trend north-south, whereas the Jurassic arc trends southeast-northwest, which creates significant divergence between the Mojave-Sonora and southern Sierran regions. Overall, existing data support the hypothesis that the Jurassic Black Rock–Sierran formed a continuous eastward-facing arc terrane; that it shut down at 160 Ma as it collided with, and was thrust eastward onto, the western margin of the Rubian superterrane; and that subsequent slab failure created a bimodal alkaline swarm of magmatism.

Boulders of ash-flow tuff occur within debris-flow deposits and volcanic granules, and sand grains occur in finer-grained deposits of the Jurassic Carmel Formation located on the Colorado Plateau in southern Utah (Chapman, 1989, 1993; Blakey and Parnell, 1995). These materials were likely shed northward from Jurassic volcanic arc terranes south of the Phoenix fault. Given the likely 1500 km of sinistral slip on the fault (Hildebrand, 2009), the original source area is probably located well to the southeast today, probably within Oaxaquia (Fig. 5). Oaxaquia contains Jurassic arc rocks built on a largely Grenvillian basement covered with Permo-Triassic sedimentary rocks (Ruiz et al., 1988; Keppie et al., 2001, 2003; Solari et al., 2003; Ortega-Obregón et al., 2003; Centeno-García and Silva-Romo, 1997; Jones et al., 1995; Barboza-Gudiño et al., 2004), which fits well with detrital zircon profiles from Jurassic rocks of the southern Colorado Plateau (Dickinson and Gehrels, 2009; Mauel et al., 2011).

Slab Break-Off

Because many geologists aren't familiar with the concept of slab failure or break-off, a short review is perhaps warranted. Slab failure, or as it is sometimes called, slab break-off or delamination, is a natural consequence of subduction, yet Cordilleran geologists have been slow to embrace its effects. As early as 1981, seismologists recognized that when the edges of large continental masses are partially subducted, their buoyancy leads to failure of the subducting slab quite close, if not right at, the leading edge of the continental shelf (McKenzie, 1969; Osada and Abe, 1981; McCaffrey et al., 1985; Welc and Lay, 1987). This is because the buoyancy forces resisting the subduction of continental lithosphere are as large as those pulling oceanic lithosphere downward (Cloos, 1993). Eventually, the greater density of the oceanic lithosphere causes the lower plate to tear at its weakest point and sink into the mantle. When the subducting slab fails and the lower plate is freed of its oceanic anchor, rocks of the partially subducted continental margin rise due to buoyancy forces. Hildebrand (2009) argued that during break-off, the plates still converge, albeit slowly, because the tearing is typically somewhat diachronous. The combination of convergence and uplift leads to strong frictional drag along the widening contact between the upper and lower plates, which compresses the crust of the lower plate to create thick-skinned structures, such as folds and trans-crustal thrust faults. In the very young arc-continent collision in New Guinea, the Mapenduma anticline, a thick-skinned anticlinal thrust structure, developed in the collisional foreland during the waning stages of the collision (Cloos et al., 2005). The similarity, both in tectonic setting and deformational timing, of thick-skinned Laramide structures of the western United States suggested to Hildebrand (2009) that they formed in the same way as those in New Guinea. Similar structures are known from the southern Andes in the Sierras Pampeanas ranges and in the Magellanic foredeep of Argentina (Fosdick et al., 2011; Cristillini et al., 2004), Wopmay orogeny (Hoffman et al., 1988), and within other Canadian orogens (Hoffman, 1989).

Another major feature of slab break-off is related magmatism. When a subducting slab tears and breaks, asthenosphere can rise up as much as 100 km through the tear and adiabatically melt, which produces magmas that can intrude either the lower plate at the site of break-off (Hildebrand and Bowring, 1999), or the upper plate above the tear (McDowell et al., 1996; Housh and McMahon, 2000; McMahon, 2000a, 2000b; Chung et al., 2003), or both. Perhaps the most critical factors in determining the pattern of break-off magmatism are the stretching rate and the actual break-off speed (Cloos et al., 2005). The resultant magmatism may be highly variable, both in time and space, due to both mantle and crustal heterogeneities (Housh and McMahon, 2000), and range from pure asthenospheric melts to complex crustal melts and, of course, mixtures of the two.

Typically, slab-failure magmatism occurs during collision and consequent arc shutdown. If the magmas intrude the upper plate, they may form a linear belt atop or alongside the old arc, which is temporally continuous with older magmatism, and so be readily confused with it. Magmas might also intrude rocks of the foredeep and/or the shortened passive margin of the lower plate, or both (Hoffman, 1987; Hildebrand and Bowring, 1999; Hildebrand et al., 2010a).

Jurassic Slab-Failure Magmatism

Following the widespread shutdown of Jurassic arc magmatism at 160 Ma, the Independence dike swarm—first recognized by Moore and Hopson (1961), and initially dated by Chen and Moore (1979)—was emplaced in the east-central Sierra Nevada as well as in the Mojave Desert region to the south (James, 1989; Carl and Glazner, 2002; R.F. Hopson et al., 2008). The dikes are poorly explained in current models (Dickinson and Lawton, 2001), but since they postdate the Luning-Fencemaker collision by just a few million years, they might represent slab-failure magmatism related to the failure of the westward-dipping slab due to the difficulty of subducting the Rubian continental crust (Fig. 41). Slab failure readily explains the long, linear, and fairly narrow breadth of the swarm, its locally bimodal character

(McManus and Clemens-Knott, 1997), and its generally alkalic nature (Karish et al., 1987). There are some 90 Ma Cretaceous dikes within the Sierra Nevada in the same area as, and more or less colinear with, the Independence dike swarm, but the 60 Myr gap between the two periods of magmatism suggests that they are unrelated and that the younger dikes might be related to individual Sierran plutons (Coleman et al., 2000).

Magmatism younger than 160 Ma clearly postdates the deformation along the length of the arc (Fig. 42) and is represented in the Goddard pendant, the central Mojave Desert, and southern Arizona. The linear band of bimodal alkaline plutons, which were discussed earlier, and are unnamed in the Mojave, but called Ko Vaya in southern Arizona, are about the same age as the dike swarm and also possibly related to slab failure as they aren't typical subduction-related magmas. Along with the dike swarm, they occur over the length of much of the defunct arc. The general tendency has been to consider both the dike swarm and the bimodal plutonic suite as part of continuous Jurassic arc magmatism, but separated from the main period by a period of intra-arc compression. While it is difficult to define slab-failure magmatism by composition, the long linear nature of the belt and its peculiar bimodal composition, suggest that it is not typical arc magmatism. Its occurrence after 160 Ma deformation and arc shutdown, yet well before known Cordilleran-type Sierran magmatism, makes it an excellent candidate for slab-failure magmatism.

Blue Mountains Assembly

Within the Blue Mountains amalgamation the two arc terranes, the Olds Ferry and Wallowa terranes, and their intervening accretionary prisms of Baker terrane were joined by what appears to have been a soft collision as accretionary rocks were thrust over each arc (Avé Lallemant, 1995; Dorsey and LaMaskin, 2008). The collisional event is bracketed to be between 159, the age of the youngest deformed sedimentary rock, and 154 Ma, the age of the Goldbug pluton which cuts the faulted contact of the Greenhorne-Bourne subterranes within Baker terrane (Schwartz et al., 2011). It is unknown where this amalgamated terrane was located during the Late Jurassic, but by the Late Cretaceous it was part of the Rubian collage as the Ochoco basin, which sits unconformably upon the older rocks, was part of the same basin as the Hornbrook farther south (Surpless et al., 2009). One clue as to its location during the Late Jurassic–Early Cretaceous comes from the McCloud fauna, which occur in the accretionary prism rocks of the Baker terrane and are present in Permian–Jurassic accretionary complexes that occur within the western block of Jurassic Sierran arcs and possible continuations farther north. The two Triassic–Jurassic arc terranes are similar to the Stikine and Quesnellian arcs farther north and Klamath-Sierran belt farther south, so overall there is good reason to include them with those broad packages of rocks, all of which joined Rubia by 150 Ma. The contact of the Blue Mountains superterrane with the Belt-Purcell-Windermere block to the east in Idaho is a major strike slip fault and might be one part of the Intra-Quesnellia fault hypothesized to exist on the west side of Quesnellia (Irving et al., 1995).

Jurassic Amalgamation of Rubia

It should be clear from the analysis presented above that the Middle to Upper Jurassic was a time of major amalgamation of terranes within the Rubian superterrane. A western block, which consists of Triassic–Jurassic arc terranes of varying types formed a western superterrane, which I name Sierrita. It joined the eastern block, known as Proto-Rubia, to form the Rubian ribbon continent, or megaterrane, at ~160 Ma. By that time, virtually all of the Rubian ribbon continent, except for a few local stragglers to the west, were amalgamated. Subduction was generally westward so that more easterly terranes were pulled beneath more westerly terranes. Following the accretion of the Smartville and Rogue Valley arcs and their slabs of oceanic marginal basin lithosphere, new eastward-directed subduction commenced along the western side of Rubia.

Subduction on the Western Side of Rubia

New easterly-dipping subduction started up west of the accreted Smartville block some time after 159 Ma, the age of Smartville accretion. New magmatism related to this subduction was located much farther to the west in the Sierran foothills (Saleeby, 1981) than the older Jurassic subduction. The westward jump in magmatism apparently represents the jump in trench location from eastern side of the accreted arcs to the western side of the accreted Smartville block. A suite of 142–140 Ma plutons that cut all terranes of the Sierran foothills and Klamaths (Hietanen, 1973; Irwin, 2003), but are, as far as I'm aware, generally sparse in the Sierran batholith, might be related to this subduction; but if so, the magmatism was scarce to absent for the next 15 Myr.

The oldest clear-cut evidence for easterly subduction occurs within the Franciscan subduction complex where the easternmost zone, the South Fork Mountain Schist, contains detrital zircons as young as 131 Ma (Dumitru et al., 2010). This differs from the generally accepted ideas, which attempt to date the inception of subduction by dating the oldest high-grade blocks in mud matrix mélange of the Central belt, even though serpentinite rinds around the blocks (Cloos, 1986) indicate that they were reworked prior to incorporation into the mélange. The high-grade blocks might have been derived from older belts located to the east and distributed by mass wasting processes (Jayko, 2009). Whatever the case, if the Franciscan complex originally formed west of the Sierran foothills, it should be no older than 159 Ma; was clearly active just after 131 Ma; and indicates active eastward subduction along the western margin of the Rubian superterrane.

As mentioned above, the oldest plutons within the Cretaceous Sierra Nevada are 125 Ma mafic to intermediate bodies located along the western part of the batholith (Saleeby and Sharp, 1980; Stern et al., 1981; Bateman, 1992; Clemens-Knott and Saleeby, 1999), and because there is a 15 Myr gap in magmatism, there is no simple correlation with eastward subduction. This presents a problem for models that derive Sierran magmatism by eastward subduction of oceanic crust beneath the Sierra. We will see in a subsequent section that the Sierran magmatism might not be related to eastward subduction at all and instead better generated by westerly-dipping subduction (Hildebrand, 2009). An alternative possibility is that perhaps the rocks of the Franciscan complex, Coast Range ophiolite, and Great Valley group were not adjacent to the Sierra at that time (Wright and Wyld, 2007).

Arctic Alaska–Angayucham Collision in Alaska

The attempted subduction of Arctic Alaska beneath the Koyukuk arc started during the Late Jurassic–Lower Cretaceous and as the rocks within the North Slope subterrane of Arctic Alaska are unlike those of North America, this collisional event is considered to have occurred within the Rubian superterrane while it was offshore from North America. Intense deformation and metamorphism during the Devonian Ellesmerian orogeny of Arctic Alaska and the overlying Lower Mississippian foredeep sequence are much more typical of various terranes within the Rubian superterrane, such as the Roberts Mountain allochthon, than western North America, which simply do not have Devonian deformation or orogenic debris shed from the west.

In the general collisional scheme, the Brookian orogeny of Alaska occurred when the Angayucham ocean closed and the Koyukuk terrane, which is interpreted to represent an arc terrane on the upper plate, collided with Arctic Alaska during the latest Jurassic–Neocomian (Roeder and Mull, 1978; Box, 1985; Box and Patton, 1987; Mull, 1985; Moore et al., 1994). The exact timing of the collision is unresolved at present. The difficulty lies with the interpretation of the strongly deformed, Late Jurassic–Neocomian Okpikruak Formation, which sits atop the Endicott and DeLong Mountain allochthons in the central and western Brooks Range, and is interpreted to represent debris shed from the moving allochthons to the south, thus dating the collision (Moore et al., 1994). The main problem is that rocks of the Okpikruak Formation don't contain a detrital zircon profile compatible with allochthonous rocks of the Brooks Range (T. Moore, 2011, personal commun.), so the Okpikruak might be pre-collisional. A younger age for the collision would better fit the evidence for dramatic downward flexure of the Colville Basin during the Barremian–Aptian (Cole et al., 1997) and appears to coincide with thrusting to the west in the Lisburne Hills (Moore et al., 2002), the Berriasian–Aptian magmatism of the uppermost Koyukuk arc (Box and Patton, 1989), and the 3400 m of Aptian–Cenomanian sandstone, conglomerate, mudstone, and coal beds of the foredeep (Mull, 1985; Siok, 1989; Crowder, 1989; Bird and Molenaar, 1992). Episodes of cooling, inferred to be related to uplift occurred at 100 ± 5 Ma, 60 ± 4 Ma, and at 24 ± 3 Ma (O'Sullivan et al., 1997).

Possibly overlapping at least partly with the collision, was the opening of the Canada Basin, as indicated by the formation of Jurassic rift basins along the northern margin (current coordinates) of Arctic Alaska, and ultimately new oceanic crust by the Hauterivian (Grantz and May, 1982). Some workers suggested, based largely on paleomagnetism, that Arctic Alaska rifted and rotated from the northern Arctic Islands of Canada during the opening and are thus conjugate margins (Grantz and May, 1982; Ziegler, 1988; Embry, 1990; Plafker and Berg, 1994; Grantz et al., 2011); but Lane (1997) presented several lines of evidence—such as different ages of Devonian deformation, differing ages of rift-drift transitions, and 600 km of overlap of the Russian shelf and Canadian Arctic islands if 66° of rotation is restored—that challenged the viability of the conjugate model. Macdonald et al. (2009) compared Neoproterozoic stratigraphy to reach the same conclusion. Paleomagnetic data suggest that Arctic Alaska rotated 105° +49°/–43° counterclockwise and was located 12° ± 5° south of its present position at 130 Ma (Halgedahl and Jarrard, 1987), which seem hard to reconcile with initiation of the Amerasian Basin at that time unless there were a lengthy southward-trending rift arm. More recently, Helwig et al. (2011) interpreted seismic profiles in the Beaufort Sea, from the Mackenzie River delta to Banks Island, to image an extinct spreading center that could represent the northern part of such a rift.

The two regional bands of highly variable, calc-alkaline to alkaline bodies that intrude the Yukon-Koyukuk basin–Seward Peninsula and the Ruby terrane (Patton and Box, 1989; Miller, 1989; Arth et al., 1989a; Patton et al., 1987, 2009; Till et al., 2010) clearly cut the thrusts of the Ruby-Angayucham package (Patton et al., 2009), and are 112–99 Ma (Miller, 1989), so are clearly postcollisional with respect to the Brookian orogeny. These relations suggest to me that they are slab-failure plutons formed when the Arctic Alaska slab failed during attempted subduction of the North Slope terrane as originally proposed by Wartes (2006).

Terranes of Alaska have similar attributes to a group of terranes found throughout the Cordillera as noted by many previous workers. Overall, the Farewell terrane contains similar successions to those of the Selwyn basin and Roberts Mountain allochthon: Cambrian to Devonian deeper-water sedimentary rocks and Devonian to Pennsylvanian carbonates, Devonian and Triassic phosphatic black shale, barite, and sandstone, with a variety of gabbroic sills and pillowed basalts (Bundtzen and Gilbert, 1983; Bradley et al., 2006). These rocks are also similar in lithology, age, and metallogeny to rocks fairly widespread within the Kootenay terrane (Smith and Gehrels, 1992a, 1992b; Colpron and Price, 1995) and in many respects, such as the presence of voluminous barite and Zn deposits (Kelley and Jennings, 2004), to rocks of Arctic Alaska. The broad temporal and lithological similarities are suggestive that rocks in these areas could have originated in the same basin (Turner et al., 1989) and were dispersed along Rubia prior to, and/or during, collision with North America. Bradley et al. (2007) used U-Pb ages and fossils to suggest that Farewell, Kilbuck, and Arctic Alaska terranes were part of one microcontinent that lay between Siberia and Laurentia from Late Neoproterozoic to the Devonian, but it is possible that only the oldest parts of the terrane reflect Siberian origins, while the younger development reflects a different history (Malkowski et al., 2010). Yukon-Tanana Uplands terrane of the Alaska-Yukon border region (Dusel-Bacon et al., 2006) is also similar, but as pointed out by Bradley et al. (2006) the Juro-Cretaceous histories are quite different, so that it was likely juxtaposed against the Farewell terrane later.

Upper Paleozoic deformation(s) are present in all those terranes, whereas rocks of the North American passive margin show no evidence for these deformations. These collisional events, of which there were probably several, commonly have contemporaneous sedimentation in adjoining basins, which likely represent foredeep fill such as the Oquirrh–Wood River basin (Geslin, 1998), and in some cases carbonatite complexes that likely represent sutures (McLeish and Johnston, 2011).

One possible link, not often considered, is the similar age of Angayucham, and Coast Range ophiolite (T. Moore, 2010, personal commun.). As the Coast Range ophiolite is the same age as the Josephine of the Klamaths and the Ingalls of the North Cascades (MacDonald et al., 2008), it seems that it might be more than pure chance that ophiolites of the same age occur throughout the Cordillera. All were formed between 170 and 160 Ma, and are some of the most outboard units in each area, which suggest that they could have formed in the same paleo-ocean. Scattered outcrops of Permian to Middle Jurassic volcanic and plutonic rocks within the Koyukuk arc just south of the Angayucham rocks (Box and Patton, 1989) occupy a similar outboard position as the Rogue arc to the Josephine ophiolite.

Sevier Fold-Thrust Belt and Early Collision on Western North America

The formation of the Western Interior basin, generally interpreted as a foredeep (Price, 1973; Kauffman, 1977; Jordan 1981; Beaumont, 1981) generated as a flexural response to loading by thrusts of the Sevier fold-thrust belt located to the west, provides the best estimate for the age of thrusting within the Sevier fold-thrust belt, which is located within the Great Basin sector. Because initial subsidence of the basin by elastic flexure of the lithosphere is coeval with loading (Turcotte and Schubert, 1982), the basal foredeep strata date the onset of thrusting within the flexural half-wavelength of the basin. As the oldest sediments of the basin sit atop the 124 Ma gravels (Fig. 13) formed when the North American platform rode over the outer swell, they should provide a maximum estimate for the inception of thrusting, that is, Aptian. This age fits with other estimates for the time of initial thrusting in the belt (Heller et al., 1986; Heller and Paola, 1989).

Such an age presents problems as there is no metamorphism and no crustal thickening of this age within the immediately adjacent hinterland belt, for as discussed, the two periods of deformation and thickening known from that region are Jurassic and latest Cretaceous. Neither period corresponds to the age of main-phase Sevier thrusting, which is no younger than 94 Ma and possibly no younger than mid-Albian. The answer to this problem may lie to the north within the Canadian sector. There, the farthest west, and presumably the oldest, thrust that cuts rocks of the North American platform, is the Bourgeau thrust (Larson et al., 2006; Price, in press). Within the footwall syncline to the east and structurally beneath the thrust are sedimentary rocks of the Cenomanian–Santonian Alberta Group (Leckie and Smith, 1992), which suggests that the thrusting that affected rocks of the North American platform in that sector was Santonian or younger. Immediately to the west in much finer-grained carbonates and clastics rather different from those of the platform, backfolded thrust faults are cut by intrusions dated by $^{40}Ar/^{39}Ar$ to be 108 Ma, so the deformation there is older by some 25 Ma (Larson et al., 2006) than thrusts to the east. A swarm of lower Paleozoic diatremes occurs in the intervening area (Fig. 32) just west of the Bourgeau thrust (Pell, 1994) suggesting that a fundamental suture may occur there (Burke et al., 2003; Johnston et al., 2003). The bulk of the Belt-Purcell and Windermere lie west of this belt as well. Similarly, the huge thrust sheets of the Selwyn basin were emplaced prior to 104 Ma (Gordey and Anderson, 1993; Mair et al., 2006).

As discussed earlier, except for the 105–100 Ma transpressional collisions within the Cordilleran batholiths, most of the terranes within Rubia were assembled by the Jurassic, so paleomagnetic data from other parts can provide constraints on the migration of the Rubian superterrane after its collision with North America during the Sevier event. Kent and Irving (2010) recently constructed a new composite apparent polar path for North America and demonstrated that from the Triassic through Early Cretaceous that Rubia moved northward more slowly than North America, so shear between the two was sinistral; but from the latest Cretaceous to Eocene, shear between Rubia and North America was dextral. Thus, following collision of Rubia with the Great Basin block of North America, it and the rest of the Rubian superterrane continued to move southward relative to North America (Kent and Irving, 2010). This sinistral motion probably continued to ~80 Ma when the relative motion between Rubia and North America became dextral as North America started to move southward. This more modern work supports the original ideas of Avé Lallemant and Oldow (1988), who much earlier suggested Triassic–mid-Cretaceous sinistral migration of exotic terranes along the North American margin followed by dextral migration.

Given that there are large temporal mismatches in deformation in both the Canadian and Great Basin sectors, I suggest that the "missing" collider in the Great Basin area is now present within the Canadian rocks west of the Main Ranges and that it was the Canadian sector that originally collided with western North America during the Sevier orogeny. Subsequently, they moved southward relative to North America only to return after 80 Ma. This fits well with metamorphism of the autochthon within the Monashee complex (Fig. 32), which was initially thickened at around 125 Ma and exhumed rapidly at

~60 Ma (Parrish, 1995) coeval with exhumation in the Belt-Purcell allochthons (Sears, 2001) and foreland basin (Price and Mountjoy, 1970; Ross et al., 2005). It also fits well with the end of Cordilleran magmatism at 105–100 Ma in the Omineca belt, the arc within the Canadian sector (Hart et al., 2004).

Effects of Sevier Collision on Franciscan Complex and Great Valley Group

There is plenty of evidence for sedimentation along the eastern margin of the collision zone during the attempted subduction of North America beneath the Rubia superterrane, because it is well preserved in the Western Interior basin. However, scant attention has been paid to the effects of sedimentation along the western margin of Rubia, yet uplifted areas within the collision zone probably drained to the west as well as to the east. In fact, a major change from nonaccretionary to accretionary style at 123 Ma in the Franciscan complex was noted by Dumitru et al. (2010) and as they pointed out, based on the work of others (Clift and Vannucchi, 2004; Scholl and von Huene, 2007, 2010; von Huene et al., 2009), the change to an accretionary regime is best explained by sediment flooding of the trench. Dumitru et al. (2010) also pointed out that the change in Franciscan sedimentation coincided with a major petrofacies change (Ingersoll, 1983) and major discontinuity (Constenius et al., 2000), marked by faulting, warping, and erosion, in rocks of the Great Valley Group at ~125 Ma (Fig. 26). Because the timing of these two events so closely mirrors the initiation of collision between North America and the Rubian superterrane on the eastern side of Rubia, I suggest that the attempted subduction of the western margin of North America generated deformation, uplift, and erosion throughout Rubia and substantial quantities of sediment were eroded and transported westward by rivers draining into both the fore arc and trench. As the western margin of North America was oriented more or less N-S during this period and at mid-latitudes (Kent and Irving, 2010), easterly flow of polar front storms coming off the Pacific basin would have amplified the erosional effects of any uplift within Rubia (Hoffman and Grotzinger, 1993) to produce copious quantities of sediment. This situation was analogous to that of the collision of India and Eurasia, where voluminous quantities of sediment were shed southward into the Indian Ocean and flooded the northward-dipping trenches both east and west of India (White and Louden, 1982; Kopp et al., 2000, 2001).

Plate Reorganization Due to Collision

At 125–120 Ma several major events occurred that affected large areas of the orogen: (1) thin-skinned thrusting started up within the Sevier fold-thrust belt; (2) Cordilleran magmatism commenced; (3) sedimentary rocks within the Great Valley Group were deformed; (4) the Franciscan accretionary complex was flooded with sediment; and (5) the Pacific plate started to drift northward. I relate all of these events to the attempted subduction of the North American craton and its passive margin cover beneath the Rubian superterrane (Fig. 41).

When North America collided with Rubia at ~125 Ma, the convergence rate between the two plates probably started to decrease dramatically and plate boundaries of adjacent plates—along with their motion vectors—were likely reorganized, just as they were during the collision of India with Asia (Copley et al., 2010). Thus, the closure of the Panthalassic ocean and accretion of the Rubian superterrane probably had a major impact on plates along its western side within the Pacific basin. In fact, as pointed out by Dumitru et al. (2010), this time corresponds within error to a major cusp—from southerly to northerly migrating—in the apparent polar wander path (APWP) for the Pacific plate (Beaman et al., 2007; Sager, 2007).

Great Basin Slab-Failure Magmatism and STEP Faults

Following collision of the Rubian ribbon continent with North America in the Great Basin sector, arc magmatism shut down. This occurred at ~105–100 Ma within the Omineca belt (Hart et al., 2004; Johnston, 2008), the arc terrane of the eastern Canadian sector, which, based on age of deformation, is considered to have been the colliding sector of Rubia. Following emplacement of the huge allochthons characteristic of the Sevier thrusting, which in southern Canada, appears to have ended by 108 Ma based on termination of thrusting there (Larson et al., 2006), and in the Great Basin sector by ~105 Ma, based on data in Utah (Lawton et al., 2007), several linear suites of 96–90 Ma plutons were emplaced into the overriding Rubian superterrane. These include hundreds of small volume mineralized plutons of the Tombstone-Tungsten-Mayo suites in northern Canada (Fig. 22) and the Livengood and Fairbanks-Salcha suites of eastern Alaska (Hart et al., 2004, 2005; Reifenstuhl et al., 1997a, 1997b; Newberry et al., 1990, 1996). The bands of plutons extend for over 1000 km along strike after restoration of ~430 km separation on the Tintina fault (Gabrielse et al., 2006). Compositionally, the plutons are highly variable, but dominantly alkaline, biotite granites, monzogranites, and quartz monzonites, with associated scheelite skarns, Cu, Sb, and Au mineralization (Hart et al., 2004, 2005).

The failure of the slab in the Great Basin sector means that there must be cratonic continuations of STEP faults (Subduction-Transform Edge Propagator faults of Govers and Wortel, 2005) in the lower North American plate on both sides of the sector as the slab in the Alaskan and Canadian south-Sonoran sectors didn't fail at that time. Evidence supporting the existence of a tear between the North American and Canadian sectors occurs in the foreland where sedimentation on both sides of the Lewis and Clark lineament differs greatly and is much thicker along the south side (Wallace et al., 1990). According to them the zone is made up of braided, anastomosing faults that were active from ~100–78 Ma: basically from Sevier to Laramide deformation. To the south lies the Phoenix fault, and there the foredeep sedimentation apparently did not continue south of the fault.

Cretaceous Cordilleran Batholiths

The Cretaceous Cordilleran batholiths—typified by magmatism of the Sierra Nevada, but also represented widely throughout the Cordillera (Figs. 5 and 22)—have long been considered to be the products of eastward subduction beneath the western edge of North America (Hamilton, 1969a, 1969b). However, there are enough problems and complexities with this interpretation to warrant a closer look. Based on the idea of Ducea (2001) that the batholiths represent exceptionally high magmatic flux, Hildebrand (2009) suggested that the batholiths owed their origin to dewatering of sediments during westward subduction of the outer extended passive margin of North America, whereas DeCelles et al. (2009) attempted to explain them with underthrusting and melting of nearly 400 km of North American middle and lower crust, followed by delamination of a dense restite. Both models require westward subduction beneath the Sierra Nevada to pull the buoyant continental margin to depths of magma generation in the mantle although DeCelles et al. (2009) did not invoke an oceanic slab as the driving force. Instead they invoked strong coupling between the craton and easterly subducting oceanic lithosphere as the driving force.

In a general sense it seems reasonable to assume that the flux of basalt should be about the same whether it be beneath oceanic or continental crust as the process is the same. First, let's examine the best estimates for magmatic flux in oceanic arcs then evaluate the approximate flux in the post-collisional post–100 Ma Cordilleran batholiths, which are the best-known phases. Although they aren't well constrained, estimates for volumes of basaltic magma arriving at the base of the crust in oceanic arcs vary widely and lie between 1 and 100 km^3/Myr per kilometer of arc length (Marsh, 1979; Reymer and Schubert, 1984; Crisp, 1984; Taira et al., 1998; Holbrook et al., 1999; Larter et al., 2001; Dimalanta et al., 2002; Scholl and von Huene, 2009; Stern and Scholl, 2010; Schmidt and Jagoutz, 2012). Within the continental crust of the Peninsular Ranges batholith, Silver and Chappell (1988) estimated the flux for the eastern post-collisional La Posta plutons to be about 75 km^3/Myr per kilometer of arc length, based on an average plutonic thickness of 20 km. For the post-collisional 100–80 Ma plutons of the Coast plutonic complex, Gehrels et al. (2009) estimated magmatic flux to be between 40 and 50 km^3/Myr per kilometer of arc length. In the eastern Sierra Nevada, rocks of the Sierran Crest magmatic event occur in a 50-km-wide belt and were emplaced over a period of about 15 Myr (Coleman and Glazner, 1988). If one assumes that the magmatic rocks are 10 km thick, the calculated flux rate is 33 km^3/Myr per kilometer of arc length and if one assumes them to be 20 km thick the rate would be double, or 66 km^3/Myr per kilometer of arc length. Even though estimates for both oceanic and continental flux are not particularly robust, the possible flux rates in the Cretaceous Cordilleran batholiths are consistent with flux rates estimated for oceanic arcs.

Rather than the collisional model presented by Hildebrand (2009), DeCelles et al. (2009) presented a model for the origin of Cordilleran batholiths that involves cyclic magmatism created by underthrusting of lower continental lithosphere in a back-arc setting and melting of that lithosphere by asthenospheric magmas. Even if the setting were back arc, their model would be untenable based on several lines of reasoning.

First, their hypothesis supposes that all of the Triassic–Jurassic magmatism is related to the same subduction, which we have already seen is unlikely as there are major deformational events unaccounted for in their model. Second, because their back-arc model calls for ~3–400 km of thin-skinned shortening east of the Sierra Nevada (DeCelles, 2004; DeCelles et al., 2009), some 3–400 km of cratonic basement from that area must have been disposed of to balance the crustal section. Some workers (Ducea, 2001; DeCelles, 2004; Ducea and Barton, 2007; DeCelles et al., 2009) suggest that this 3–400 km of cratonic crust disappeared beneath the Sierra Nevada, where it was melted to create Sierran magmas and a dense restite, which then sank into the mantle (Fig. 43). The difficulty of subducting 300–400 km of cratonic crust without attached oceanic lithosphere to counteract and overcome the buoyancy forces of cratonic crust is extreme and left unexplained in their model; but even were it possible, the model suffers from severe mass balance and room problems. Consider that the Sierra Nevada batholith is ~100 km wide so that a strip of crust some 3–400 km wide by 30–40 km thick means that sufficient North American crust was underplated to thicken the Sierran crust to 120–150 km, which seems excessive given the amount of exhumation. Melting the crust and delaminating the residue doesn't help resolve the problem because average continental crust contains ~61% silica (Rudnick and Gao, 2003), and the upper 30 km of Sierran crust, according to Ducea (2002), contains ~65% silica, only a 6% difference, which means that since nearly an equal volume of cratonic crust must be melted to create the same mass with the bulk composition of the Sierra Nevada, there would be little restite. And that melting doesn't remove the crust but only serves to distribute it upwards. Even if you just want to melt the middle and lower crust, compositionally it would only be ~11% different from the bulk Sierra Nevada, so there would still be little residue to drip or delaminate into the mantle. And all of this doesn't even address the source of energy necessary to melt all that crust. Where could that come from?

Based on the observation that the Cordilleran batholiths coincided temporally with the period of thin-skinned thrusting within the Sevier fold-thrust belt, Hildebrand (2009) suggested that they owed their origin to dewatering of the continental rise-prism sediments on the outer parts of the North American passive margin during abortive westerly-directed subduction of that margin. However, the recognition that the batholiths are composite bodies suggests alternative possibilities, all of which are more complex and not particularly well constrained, largely because the locations of the Sierra Nevada and Franciscan complex with respect to each other and North America, at different times, are poorly known. Additionally, the initiation of easterly subduction beneath the western part of the Sierra Nevada may be much

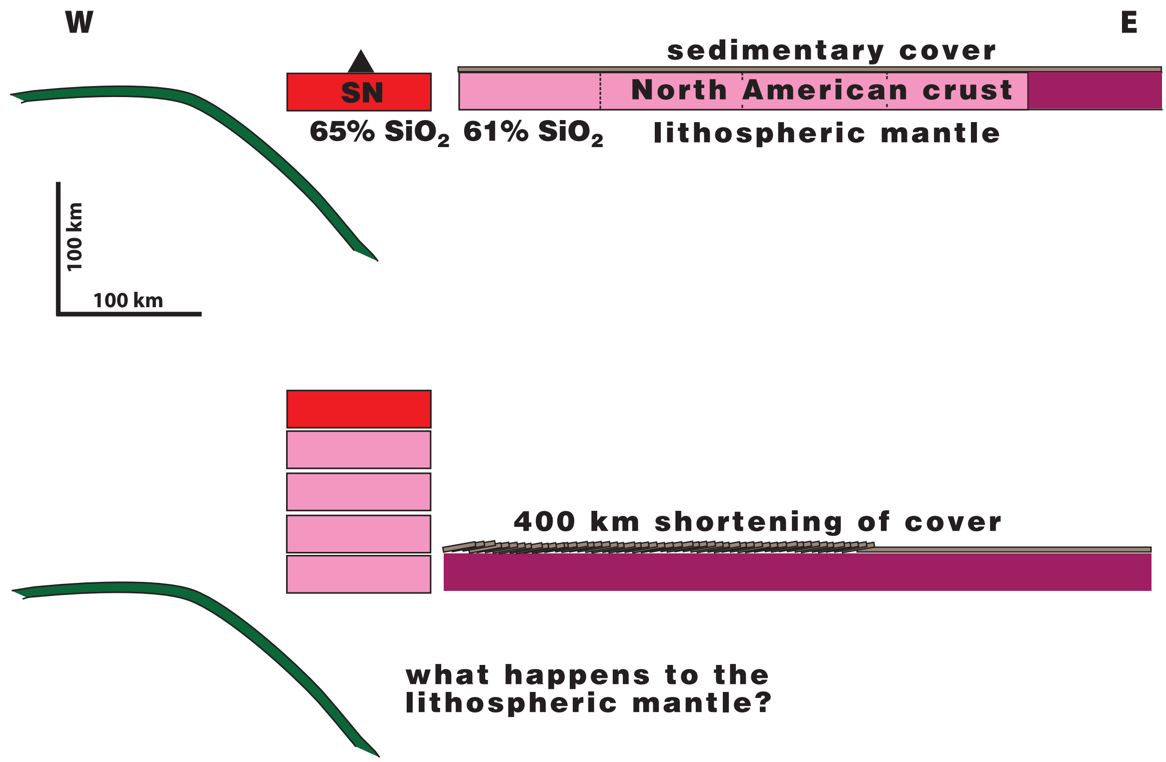


Figure 43. Other workers (DeCelles et al., 2009) tried to explain the origin of Cordilleran-type batholiths and the overall development of the Cretaceous–Tertiary Cordilleran orogen by westwardly-directed underthrusting of hundreds of kilometers of cratonic North America in a back-arc setting above an easterly-dipping subduction zone as shown in this figure. In that model 3–400 km of cratonic crust disappeared beneath the Sierra Nevada, where it was melted to create Sierran magmas and a dense restite, which then sank into the mantle. Such a model suffers from mass balance and room problems, because the Sierra Nevada is ~100 km wide so that a strip of crust some 3–400 km wide by 30–40 km thick means that sufficient North American crust migrated westward to thicken the Sierran crust to 120–150 km. Melting the crust and delaminating the residue doesn't help resolve the problem because the average continental crust contains ~61% silica, and the upper 30 km of Sierran crust contains ~65% silica, only a 6% difference, which means that nearly an equal volume of cratonic crust must be melted to create the same mass with the bulk composition of the Sierra Nevada, which leaves little restite. And that melting doesn't remove the crust but only serves to distribute it upwards. This model also fails to explain why magmatism shut down at ~80 Ma along the entire Cordillera.

younger than previously considered. First, there is a paucity of evidence for arc magmatism prior to about 125 Ma. Older latest Jurassic–Early Cretaceous magmatism is limited to short pulses of magmatism that appear to have been more likely related to collisional events in the western Sierran foothills. These include the pulses of 159–150 Ma, likely related to slab failure following the Smartville collision, and the short-lived burst at 142–140 Ma, which is poorly explained in all models. The earliest plutons of the western arc terrane appear to be 125–120 Ma ring complexes found in the westernmost Sierra (Clemens-Knott and Saleeby, 1999). Other bodies of similar age, such as the 123 Ma Ward Mountain trondhjemite and 121–105 Ma Bass Lake tonalite, occur just west of Yosemite National Park farther north (Lackey et al., 2012a, 2012b).

A younger date for the start of Sierran magmatism might better fit with the initiation of the Franciscan complex. Although the start-up of Franciscan subduction is generally placed at about 160 ±10 Ma, based on the ages of high-grade metamorphic blocks in mélange, it may be that the Franciscan wasn't formed until shortly before 130 Ma, the maximum age of the oldest clastic sedimentary rocks within the complex: the easternmost and structurally highest coherent blueschists, which contain detrital zircons of 131 Ma (Dumitru et al., 2010). The high-grade blocks are clearly polycyclic as they are encased in actinolite-chlorite, probably once serpentinite, yet float in terrigenous clastic mélange (Coleman and Lanphere, 1971; Cloos, 1986). It could therefore be argued that the blocks are entirely exotic with respect to the Franciscan complex and that Franciscan subduction started much later than generally thought. Thus, if rocks of the Franciscan complex could be demonstrated to have been adjacent to the Sierra at about 125 Ma—and it's not clear that they were (Jayko and Blake, 1993)—easterly subduction may have started just prior to 130 Ma.

The main problem with such a hypothesis is that the Franciscan may not have been adjacent to the Sierra prior to the 100 Ma collision (Jayko and Blake, 1993) that joined the eastern and western halves. Given that the 125–100 Ma plutons in the western Sierra young from west to east (Lackey et al., 2012a, 2012b), it is possible that prior to collision there was no easterly directed subduction, instead subduction may have been westerly directed, beneath the western area. In that case subduction and rollback would have occurred in the basin between the western and eastern blocks and generated the west to east age progression. This hypothesis is supported by the mismatch in deformation between pre-100 Ma rocks of the Sierra, which are strongly deformed (Bateman et al., 1983a; Wood, 1997) and those of the 125–100 Ma parts of the Great Valley Group, which are much less deformed (Constenius et al., 2000).

Mid-Cretaceous Transpressional Deformation in Batholithic Terranes

As described in earlier sections, rocks of the Coast plutonic complex, the Sierra Nevada, Idaho batholith, and the Peninsular Ranges batholith were deformed at about 100 Ma during poorly understood events. Large faults, in places strike-slip, and elsewhere thrusts, appear to dominate the structure and divide the batholiths into two parts, which have long been recognized to contain different basements and display age, geochemical, and isotopic changes across the contacts (Fig. 44).

In the Peninsular Ranges batholith of southern and Baja California, a long-known boundary (Gastil et al., 1975, 1990) between the western Alisitos arc, possibly erupted on and through young crust, and a more easterly, continental arc terrane, formed between ~114 Ma, the age of deformed plutons in the Alisitos terrane, and ~98 Ma, the age of the oldest postkinematic La Posta pluton (Johnson et al., 1999a; Kimbrough et al., 2001). Models for the origin of the deformation include Cretaceous collision above an eastwardly dipping subduction zone (Gastil et al., 1981; Wetmore et al., 2003); collapse of a marginal basin above an easterly-dipping subduction zone (Busby et al., 1998), and collision above a westerly-dipping subduction zone (Dickinson and Lawton, 2001). Based on the occurrence of Cretaceous metavolcanic rocks and orthogneiss just to the east of the Main Martir thrust Johnson et al. (1999a) argued for a collision between two arcs, but the rocks are sufficiently high grade and deformed such that they could all be part of the Alisitos block.

Given that the Alisitos arc occurs on the western side of the collision zone and it is of much lower metamorphic grade than rocks to the east, it is reasonable to assume that the polarity of the collision was west-dipping and that the leading edge of the eastern zone (Caborca-Cortes terranes) was partially subducted beneath the arc. The contrasts in metamorphic grade across the suture support this basic concept. The east-dipping Main Martir thrust may be a back thrust formed as the Alisitos arc rode up and over the eastern block with the more easterly west-dipping thrust the main suture between blocks.

As discussed earlier, the post-collisional La Posta plutons represent a short-lived magmatic pulse ranging in age from 99-92 Ma (Silver and Chappell, 1988; Walawender et al., 1990; Kimbrough et al., 2001) and were emplaced at depths of 5–20 km into upper greenschist to amphibolite grade wall rocks that are in many places migmatitic (Gastil et al., 1975; Todd et al., 1988, 2003; Grove, 1993; Rothstein, 1997, 2003). The plutons were emplaced during a period of exhumation when rocks at depths of 10 km were brought to the surface by detachment faulting and collapse coincident with a pulse of early Cenomanian to Turonian coarse clastic sedimentation in basins located to the west and containing 100–90 Ma detrital zircons (George and Dokka, 1994; Lovera et al., 1999; Kimbrough et al., 2001). While the generally accepted view of the La Posta intrusions is that they are subduction-related, the magmatism coincident with exhumation immediately following collision suggests to me the possibility that La Posta plutons resulted from slab failure magmatism related to break-off of the west-dipping slab beneath the Alisitos arc. Once freed of its oceanic lithosphere, the eastern plate would have risen rapidly due to buoyancy forces. The asthenospheric

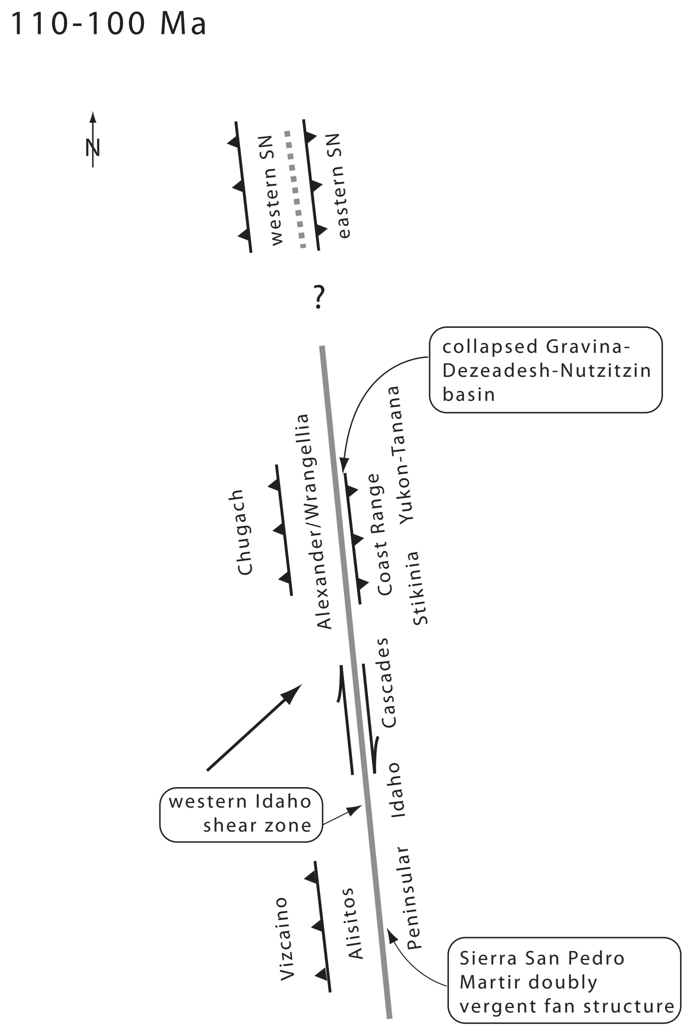


Figure 44. Sketch map illustrating the proposed relationships between various Cordilleran batholithic terranes and the 110–100 Ma transpressional event. SN—Sierra Nevada.

melts then rose into the overlying crust to generate the La Posta intrusions.

Within the Coast plutonic complex, after a period of probable sinistral transpression (Monger et al., 1994; Chardon et al., 1999; Chardon, 2003; Hampton et al., 2007; Gehrels et al., 2009), an extensive deformational event affected rocks of the complex and those of the Gravina-Dezadeash-Nutzotin-Gambier belt. The two blocks—190–110 Ma to the east and 160–100 Ma to the west—are readily delineated by U-Pb ages (Gehrels et al., 2009) and are shown on Figure 24. A west-verging thrust belt, that developed between ~100–90 Ma (Haeussler, 1992; Rubin et al., 1990), places high-grade rocks of the eastern belt over lower-grade rocks of the Gravina-Dezadeash-Nutzotin belt, and possible equivalents farther south such as the Gambier Group (Fig. 34), to form a thrust stack that is of higher metamorphic grade upwards (Lynch, 1992, 1995; Journeay and Friedman, 1993; Crawford et al., 2000; McClelland and Mattinson, 2000). Likewise, rocks of the turbiditic Kahiltna basin of southern Alaska, which lay between Wrangellia and more inboard parts of Rubia, were initially deformed on southward-vergent structures (present-day coordinates), between 115 and 100 Ma (Pavlis, 1982; Wallace et al., 1989; Ridgway et al., 2002; Trop and Ridgway, 2007; Hampton et al., 2007), in what appears to represent a diachronous south-to-north basin closure (Kalbas et al., 2007). The similar paleopoles for both Stikinia and Wrangellia during the Triassic suggest that the two were not far apart, at least in terms of latitude, at that time (Kent and Irving, 2010), and possibly that the intervening basin may not have been very wide, although constraints are poor. The similarity of the deformation in both the Peninsular Ranges batholith and the Coast plutonic complex at about the same time opens the possibility that the two might have been formerly conjoined; but only if the detrital zircon profiles (Kapp and Gehrels, 1998; Wetmore et al., 2005; Manuszak et al., 2007; Hampton et al., 2010; Alsleben et al., 2011) within the basinal facies rocks reflect varied (and proximal?) sources, as they are somewhat different.

Rocks in and adjacent to the western Idaho shear zone, which lies along the western margin of the Atlanta lobe of the Idaho batholith (Figs. 5 and 8), were strongly deformed by dextral shear after 105 Ma, the magmatic age of orthogneisses of the Little Goose Creek complex, and before 90 Ma, the age of leucopegmatites that cross cut the older fabrics (Manduca et al., 1993; Giorgis et al., 2008). This zone of deformation is most likely a continuation of that within the Coast plutonic complex, but the Idaho rocks were offset in sinistral fashion along the Orofino shear zone after 90 Ma and before 70 Ma (McClelland and Oldow, 2007).

As discussed earlier, the Sierra can also be divided into older western and younger eastern halves based on geochemistry, magnetic susceptibility, age, radiometric and stable isotopes, wall rock provenance, and basement types (Nokleberg, 1983; Chen and Tilton, 1991; Bateman et al., 1991; Kistler, 1990, 1993; Saleeby et al., 2008; Lackey et al., 2008, 2012a, 2012b; Chapman et al., 2012). Rocks of the Sierra Nevada contain evidence of an ~100 Ma deformational event that postdates all known sedimentary and volcanic wall rocks within the Sierran batholith (Bateman et al., 1983a; Bateman, 1992; Memeti et al., 2010a; Wood, 1997; Saleeby et al., 1990) yet predates, or is partly synchronous with, plutons of the 98–85 Ma Sierran Crest magmatic event (Greene and Schweickert, 1995; Coleman and Glazner, 1998; Davis et al., 2012). Intrusion of the plutons was synchronous with development of mylonitic shear zones, a rapid increase in cooling rates at 90–87 Ma, and increased sedimentation within the Turonian of the Great Valley group (Mansfield, 1979; Renne et al., 1993; Tobisch et al., 1995).

In the southern sector of the batholith, 105 Ma to 102 Ma metavolcanic, metasedimentary, and plutonic rocks are isoclinally folded (Fig. 45), penetratively deformed, and thrust to the west (Wood, 1997; Saleeby et al., 2008). Orthogneisses and plutons, ranging in age from ~115 to 100 Ma, were exhumed from 9 to 10 kb at 98 Ma to 4 kb by 95 Ma (Pickett and Saleeby, 1993; Wood and Saleeby, 1998; Saleeby et al., 2007). In the northern Sierra, recent work by Christe (2011) demonstrated that 127 Ma rhyolitic tuffs of the Trail Formation, located within the Mount Jura block, and overturned beneath the Taylorsville thrust, are 129–127 Ma. The minimum age for this deformation is unconstrained. Similarly, along the western side of the Sierra just north of Fresno (Fig. 17), folds of the 121–105 Ma Bass Lake tonalite (Bateman et al., 1983a; Lackey et al., 2012a, 2012b) are isoclinal and locally overturned. Two additional periods of folding refold the older isoclines (Fig. 46). In the central Sierra, rocks of the Goddard pendant (Fig. 17) were also deformed at about this time (Tobisch et al., 2000). The intense deformation with westerly-vergent recumbent folds and thrusts in the Sierra Nevada at 105–100 Ma is incongruous with the lack of deformation within rocks of the Great Valley group and supports the Wright and Wyld (2007) model for a far-traveled Great Valley group, or possibly that the bulk of the group was not deposited until after the deformation as suggested by detrital zircon ages dominantly younger than 100 Ma within most sedimentary rocks of the Great Valley Group (Surpless et al., 2006).

Memeti et al. (2010a) dated detrital zircons from deformed metasedimentary rocks sitting on Jurassic arc rocks in several pendants that yielded U-Pb ages in the range 100±4 Ma. As the metasedimentary rocks are cut by 101–95 Ma plutons (Memeti, 2010a), the units are dated quite precisely. These are part of a more extensive suite of deformed Albian volcanic and sedimentary rocks known from pendants throughout the central Sierra Nevada as were described earlier. These rocks are slightly younger than rocks of the Gravina-Dezadeash-Nutzotin basin, also sit on mid-Jurassic arc rocks, and were deformed shortly after deposition; so it is possible that the Sierran examples represent basinal facies rocks between the two Sierran blocks. As Memeti et al. (2010a) suggested that the Snow Lake fault was active sometime during the interval 145–102 or 87 Ma, depending on its precise location, it may be that the rocks with Death Valley provenance now located within the Snow Lake pendant were transported northward during this period of intra-arc

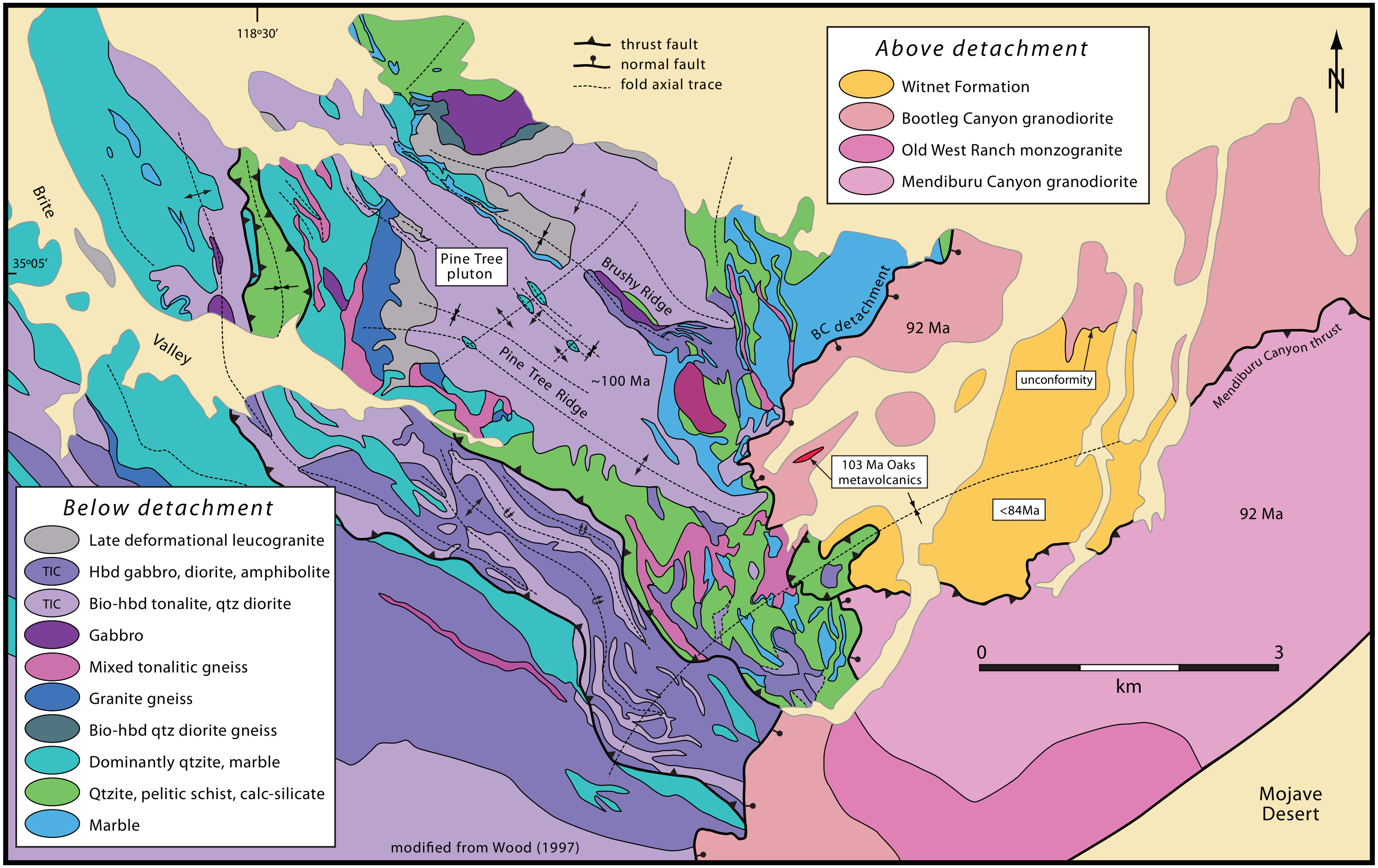


Figure 45. Geological sketch map of the eastern Tehachapi Mountains showing the isoclinal recumbent folded sheet-like plutons of the ~100 Ma Tehachapi Intrusive complex (TIC) structurally beneath the Blackburn Canyon detachment and the thrusting of 92 Ma plutonic rocks over the Late Cretaceous (?) Witnet Formation.

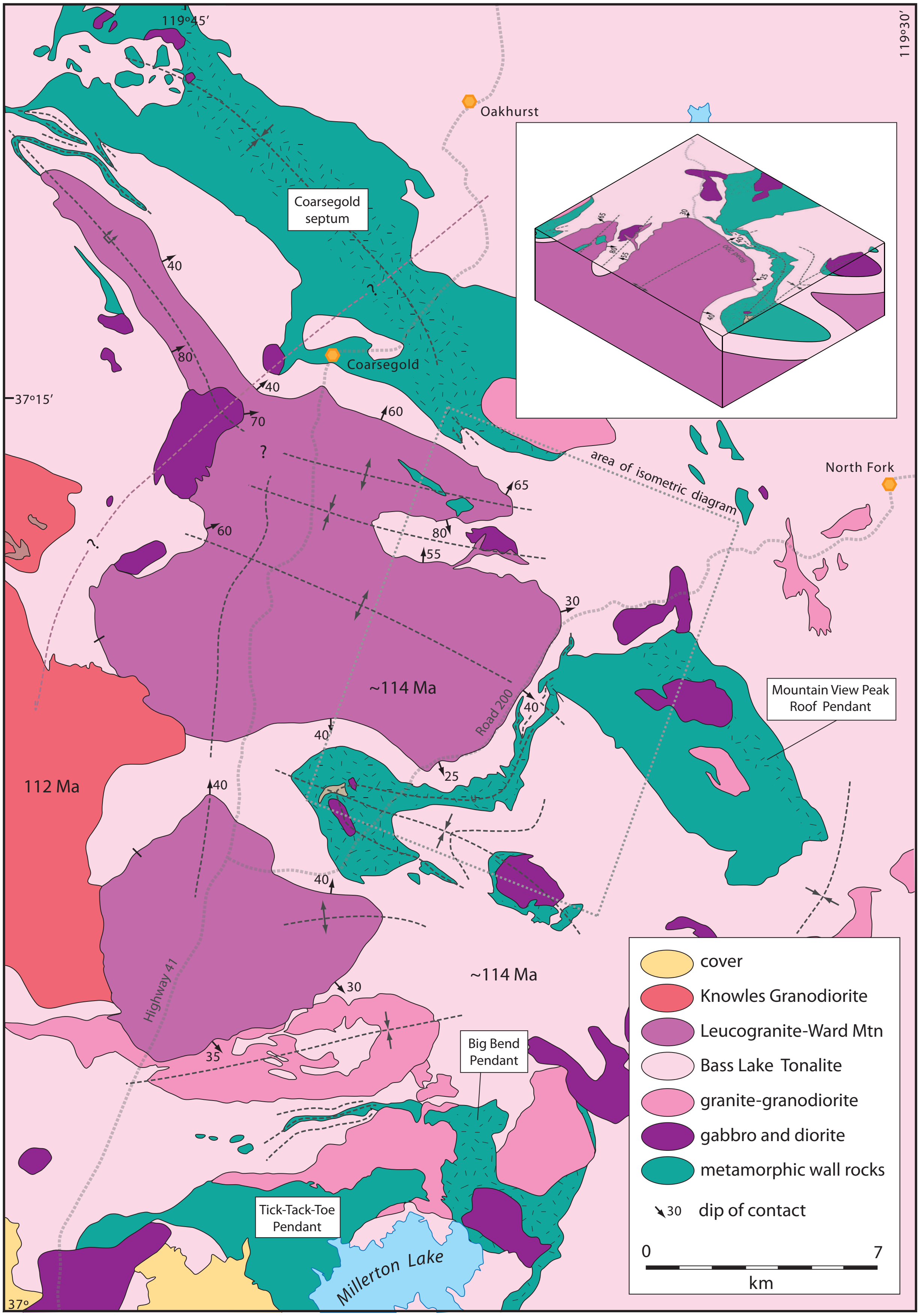


Figure 46. Sketch map showing the geology along the western side of the Sierran batholith illustrating the several generations of folds affecting pre–100 Ma plutons and their wall rocks. Modified from Bateman et al. (1983a) with U-Pb zircon ages from Lackey et al. (2012a, 2012b). Note the early F_1 isoclines.

transpression to be caught between the two sides following basin closure.

Between about 100 Ma and 95 Ma rocks of the southern Sierras were exhumed from 9 kbs to 4 kbs (Saleeby et al., 2007), which was just prior to the emplacement of plutons of the 95–84 Ma Domelands intrusive suite (Saleeby et al., 2008), the southern equivalent of the Sierran Crest magmatic suite (Coleman and Glazner, 1998). Also in the far-southern part of the batholith, there appears to be limited strike-slip motion and ~25 km of shortening that took place on the proto–Kern Canyon shear zone between ~100 Ma and 90 Ma with an additional 10 km of dextral slip along the fault between 90 Ma and 80 Ma (Nadin and Saleeby, 2008).

The similar relations in the Coast plutonic complex, the Sierra Nevada, and the Peninsular Ranges batholith suggest that the eastern post-kinematic (100–85 Ma) plutons might all be the result of slab failure rather than subduction as generally believed. The eastern zones of the batholiths are about 50 km wide and contain short-lived magmatic pulses immediately following basin closure and collision. In the Sierra the eastern zoned (?) plutonic complexes, such as the Tuolumne and Mount Whitney intrusive series, were emplaced during a short-lived, but intense, burst of magmatism known as the Sierran Crest magmatic event, which occurred from 98 to 86 Ma (Coleman and Glazner, 1998; Davis et al., 2012). In Baja California the compositionally zoned plutons of the La Posta suite were intruded between 98 and 92 Ma (Kimbrough et al., 2001). In both instances the plutons appear to have been emplaced during or just after rapid exhumation and are contemporaneous with deposition of thick Cenomanian-Turonian clastic successions to the west (Mansfield, 1979; Surpless et al., 2006; Kimbrough et al., 2001).

Because all of the dated supracrustal rocks predate the collisional event and are deformed, one of the problems with a subduction origin for the post-collisional eastern Sierran plutons is that there is little, if any, volcanic debris, such as ash flows, debris flows, and richly tuffaceous epiclastic rocks, that typically occur adjacent to magmatic arcs, for example in the Cascades (Fiske et al., 1963; Smith, 1985, 1991). Although several possibilities for this exist—the batholith was tectonically displaced from its adjacent areas at a later date, or the volcanic rocks were eroded away—the slab failure model might present a simple explanation for the absence of volcanic debris in that the magma chambers never vented to the surface and so there were no volcanoes.

All these areas appear to contain ongoing magmatism with a component of strike-slip motion along major faults (Fig. 43). As has long been recognized (Fitch, 1972; Jarrard, 1986; Oldow et al., 1989; McCaffrey, 1992, 2009), such faults are the locus of the strike-slip component of an obliquely convergent plate regime as exemplified by the Semangko (Barisan Mountain) fault of Sumatra, where they follow the arc front (van Bemmelen, 1949; Westerveld, 1953) and partition the strain among different lithospheric blocks. In the western North American cases, the different basements on either side of the sutures, coupled with the deformation and shut-down of the western arcs suggest collisions, but in each case there is strong evidence that at least the final relative motions were strongly oblique. Overall, this event, which occurred at about 100 Ma along the western margin of the Rubian superterrane, apparently after part of it docked with North America during the Sevier event, is more widespread than generally recognized.

Folding of Sierra Nevada

Over fifty years ago, Bateman and Wahrhaftig (1966) postulated that the wall rocks (their "framework" rocks) of the Sierra Nevada batholith form a large syncline, and in those pre-plate tectonics days, speculated that a downwarping of the crust led to melting of the axial zone and upwelling of the resultant magmas to form the batholith. With the advent of plate tectonics, most, if not all, workers have related the batholith to subduction processes and have considered the regional deformation to be Late Jurassic and caused by more westerly collisional events (Nokleberg and Kistler, 1980; Schweickert, 1981; Tobisch et al., 2000) or that deformation was generated by rising diapirs and consequent downflow of wall rocks (Moore, 1963; Hamilton and Myers, 1967; Sylvester et al., 1978a; Bateman, 1992; Pitcher, 1993; Saleeby and Busby-Spera, 1993; Saleeby et al., 1990; Paterson et al., 1996; Saleeby, 1999; Paterson and Farris, 2008) even though mid-Cretaceous metavolcanic and metasedimentary rocks throughout the batholith are folded (Saleeby and Busby-Spera, 1986; Saleeby et al., 1990; Fiske and Tobisch, 1994). In fact, throughout the batholith, the framework rocks, which range in age down to 96 Ma (Memeti et al., 2010a), are regionally folded about NNW axes, and by another set of folds that trend more or less easterly (Peck, 1980; Nokleberg and Kistler, 1980; Nokleberg, 1981, 1983; Wood, 1997; Saleeby et al., 1990, 2008), so all pre-96 Ma plutons must also be folded.

The obvious questions for the Sierra Nevada are: how much younger than 96 Ma was the age of folding, and how many periods of folding were there? These are difficult questions to answer definitively as there are no supracrustal rocks known to be younger than 96 Ma; so that in their absence, paleohorizontal is generally unknown. However, as the two-dimensional shape of the Cretaceous plutons appears to be about the same irrespective of age, it seems plausible that all of them might be folded.

Throughout the Sierran batholith, single stratigraphic units, in the broad sense, sit along intrusive contacts for many kilometers suggesting that roofs and floors were originally horizontal or gently dipping. It is this general concordance of plutonic contacts with bedding in framework rocks of even the younger plutons, such as the granite of North Dome (Fig. 47) that would be difficult to reproduce if the bodies were intruded after folding. This is supported by the large number of concordant screens of wall rock between individual plutons (Bartley et al., 2002; Grasse et al., 2001b; Bartley et al., 2012). It is important to remember that, even though roofs and floors of the bodies might be concordant with bedding in wall rocks, large and significant departures from the general sheet-like shape, perhaps along lateral margins,

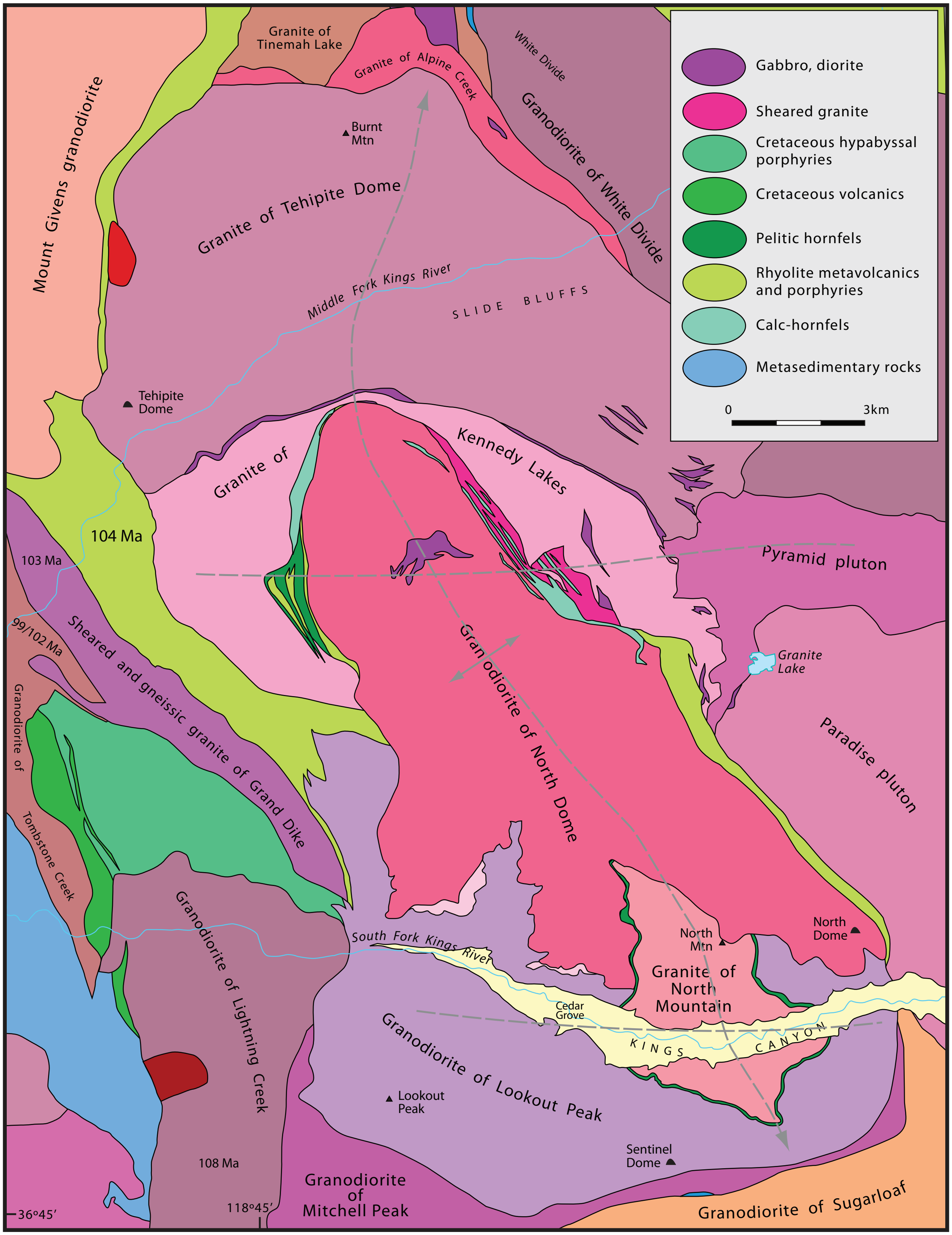


Figure 47. Geological sketch map showing the relationships of plutons and wall rocks in the Sierra Nevada of the Kings Canyon area, California. Note the apparent stacking of plutons northward from the granodiorite of North Dome and the way plutons appear to wrap around one another. I interpret these relations, along with the concordant nature between the plutons and their wall rocks, to indicate that all of these bodies are folded with flat floors and roofs. Geology from Moore (1978), Moore and Nokleberg (1992), and Grasse (2001).

or where sections of a body climb upward to a different stratigraphic level, might provide widely divergent and discordant map patterns. In this way formerly steep sections might even be overturned by younger folding, and in the absence of wall rocks containing features that allow top determination along the contact, the original three-dimensional shape might be obscure and unresolvable. Even more interesting and difficult to resolve would be the case where such contacts are rotated on overturned limbs of recumbent folds—a real possibility given that overturned limbs of folds are known in deformed Cretaceous wall rocks of the Sierra (Peck, 1980; Wood, 1997).

Despite the difficulties of delineating folds without extensive wall rocks, it is still possible to identify many folded plutons throughout the Sierra Nevada. For example, the Jurassic Granite of Bear Dome (Fig. 48) is obviously folded as foliations and contacts dip outward and wall rocks wrap around it. Within the adjacent 94–91 Ma Lamarck granodiorite foliations delineate folds as shown on Figure 48. Just to the southeast of this figure, and on strike with the syncline in the southeasterly portion of the figure, a detailed study (Davis et al., 2012) revealed that intrusive units as well as contacts and foliations clearly define a syncline as shown in Figure 49. In this area it is possible to verify that it is a foliation, defined by biotite clots and plate-like enclaves, that is folded, and that originally the foliation was roughly horizontal, parallel to external contacts of the formerly sill-like intrusion. Presumably these units continue along strike to the northwest.

Cross folds, that is, more or less east-trending anticlines, might be marked by linear bands with few wall rocks. On the regional maps (Bateman, 1992), such concentrations appear to alternate with linear belts in which wall rocks are more prevalent. The belts with greater concentrations of wall rock probably represent synclines in which roof rocks to the plutons are extensively exposed, whereas the belts with few likely represent anticlines exposing the core regions of plutons.

Additional support for the folding model comes from detailed studies of foliations within Sierran plutons. Many of the foliations within Sierran plutons have long been considered to be magmatic (Bateman et al., 1963; Bateman and Wahrhaftig, 1966; Bateman, 1992), yet the recent recognition that many of the plutons are markedly composite with foliations that cross sharp internal and external contacts (Coleman et al., 2005; Žák and Paterson, 2005, 2009; Žák et al., 2007; McNulty et al., 1996; de Saint Blanquat and Tikoff, 1997; Tikoff and de Saint Blanquat, 1997) makes it difficult to entertain the hypothesis that the foliations are predominantly magmatic features generated by viscous flow during or shortly after emplacement.

In one detailed study, Žák et al. (2007) documented four different foliations in rocks of the Tuolumne intrusive series, and argued that the two younger foliations (N-S and NW-SE), which clearly transect internal contacts, were related to postemplacement stress (Fig. 50). Recently, detailed studies by other workers have confirmed the easterly-trending foliation in plutons of the Tuolumne intrusive series (Economos et al., 2005; Johnson and Miller, 2009). It might be axial planar to the easterly-trending folds. The foliations might also provide a sound temporal constraint because the northwest to westerly foliation (Type 4 of Žák et al., 2007) in the Tuolumne intrusive series affects all units, including the youngest and most central unit, the Johnson porphyry, whereas the northerly to north-northwesterly foliation (Zák's Type 3) affects all members of the suite except the Johnson granite porphyry (Žák et al., 2007). This observation must be tempered with the caveat that where they could observe the relative timing of the two foliations on the outcrop, the northwesterly fabric (their Type 4) postdated the more northerly fabric (their Type 3). The simplest explanations for the conflict might be that the northerly fabric was simply not recognized in the field, or possibly that the Johnson porphyry was not solid enough during the deformational event to record it, but was so by the second event. If this is correct, then the N-S fabric formed during the cooling of the Johnson porphyry, dated by U-Pb on zircons to be 85–84 Ma (Coleman et al., 2005); whereas the transverse foliation must be younger.

In detailed studies of the ~98 Ma Jackass Lakes pluton (Stern et al., 1981; McNulty et al., 1996), located just south of the Tuolumne intrusive series, Krueger (2005) and Pignotta et al. (2010) found folded dikes of the pluton with axial planes parallel to metamorphic foliation in wall rocks and enclaves. Pignotta et al. (2010) also documented that the dominant NW foliations, and steep lineations in the pluton are basically the same as those in the metavolcanic wall rocks, suggesting formation after pluton emplacement.

To the east of the Sierran syncline within the White-Inyo Mountains (Fig. 5) is a series of major folds including the White Mountain–Inyo anticlinorium (Ross, 1967; Morgan and Law, 1998), and, although there are many structural complications, such as faults and folds of different ages, two major sets of folds (Fig. 51) form a basin and dome interference pattern (Ramsay, 1967; Thiessen and Means, 1980). Within the area, careful and detailed field-based studies have documented several ages of folds, including upright and overturned NW-SE and NE-SW trending folds at scales from mountain-range size to cm (Nelson, 1966a, 1966b, 1971; Bateman, 1965a; Ross, 1965; Morgan and Law, 1998). Sedimentary rocks of the area were cut by a number of Middle Jurassic and Late Cretaceous plutons (Krauskopf, 1968; Crowder et al., 1973), but the relationship of folding to plutonism has remained controversial, with some workers arguing that the Jurassic plutons postdate folding (Morgan and Law, 1998); others that the Late Cretaceous bodies postdate folding, inflated during emplacement, and shouldered aside or lifted their wall rocks (Nelson and Sylvester, 1971; Sylvester et al., 1978b; Morgan et al., 1998, 2000; de Saint-Blanquat et al., 2001); and some suggesting that Cretaceous plutons were deformed, but not necessarily folded, after emplacement (Paterson et al., 1991). Additionally, Glazner and Miller (1997) proffered the idea that some of the Jurassic plutons formed giant load casts with marginal rim anticlines; Vines and Law (2000) suggested that space for the 164 Ma Santa Rita Flat pluton (Fig. 51), was initially created as a cavity opening along a hinge zone of a syncline

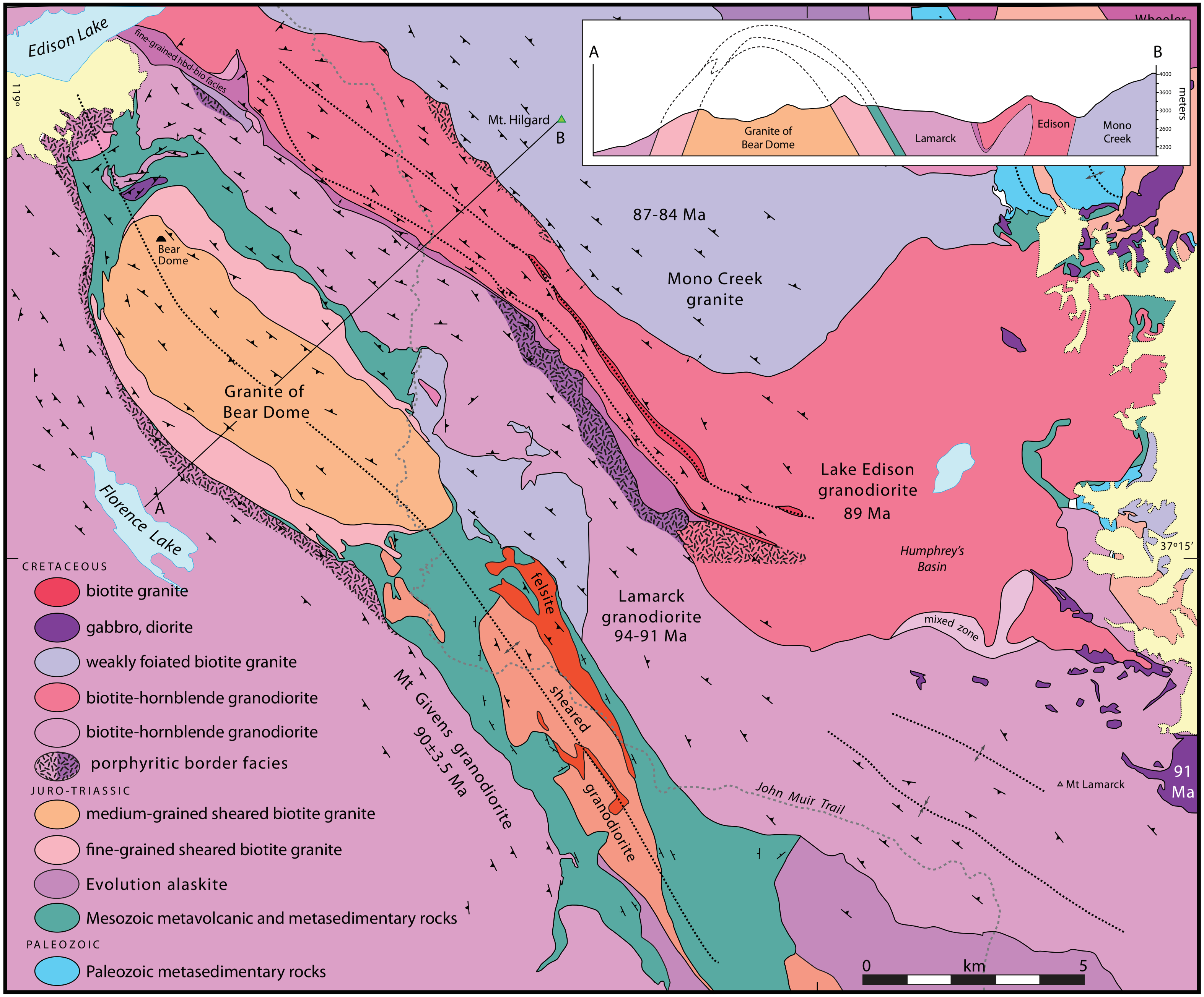


Figure 48. Sketch map showing the geology at the north end of the Goddard pendant with the folded Jurassic Granite of Bear Dome and other folds in Cretaceous plutons. Geology after Bateman (1965a, 1965b), Bateman and Moore (1965), and Lockwood and Lydon (1975).

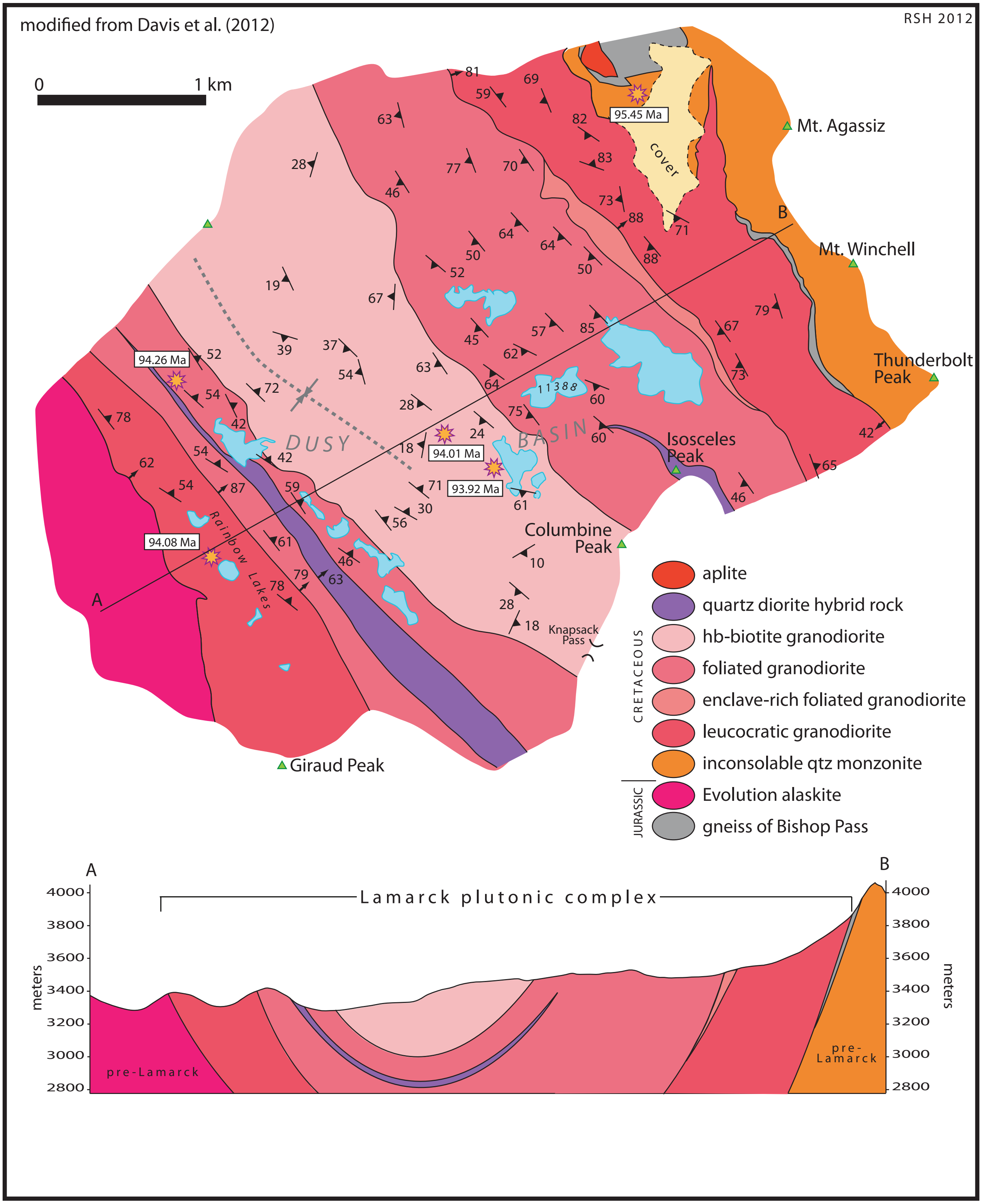


Figure 49. Geological sketch map from Davis et al. (2012) showing a detailed section across the Lamarck intrusive complex illustrating the folded nature of the body in Dusy Basin. Note the symmetry of units across the interpreted fold axis. This is a prime area to see that the cleavage is not axial planar, but a concordant, originally horizontal, fabric. This area is located just off the southeast corner of a previous figure (Fig. 47) and is a continuation of the prominent syncline in the Lamarck.

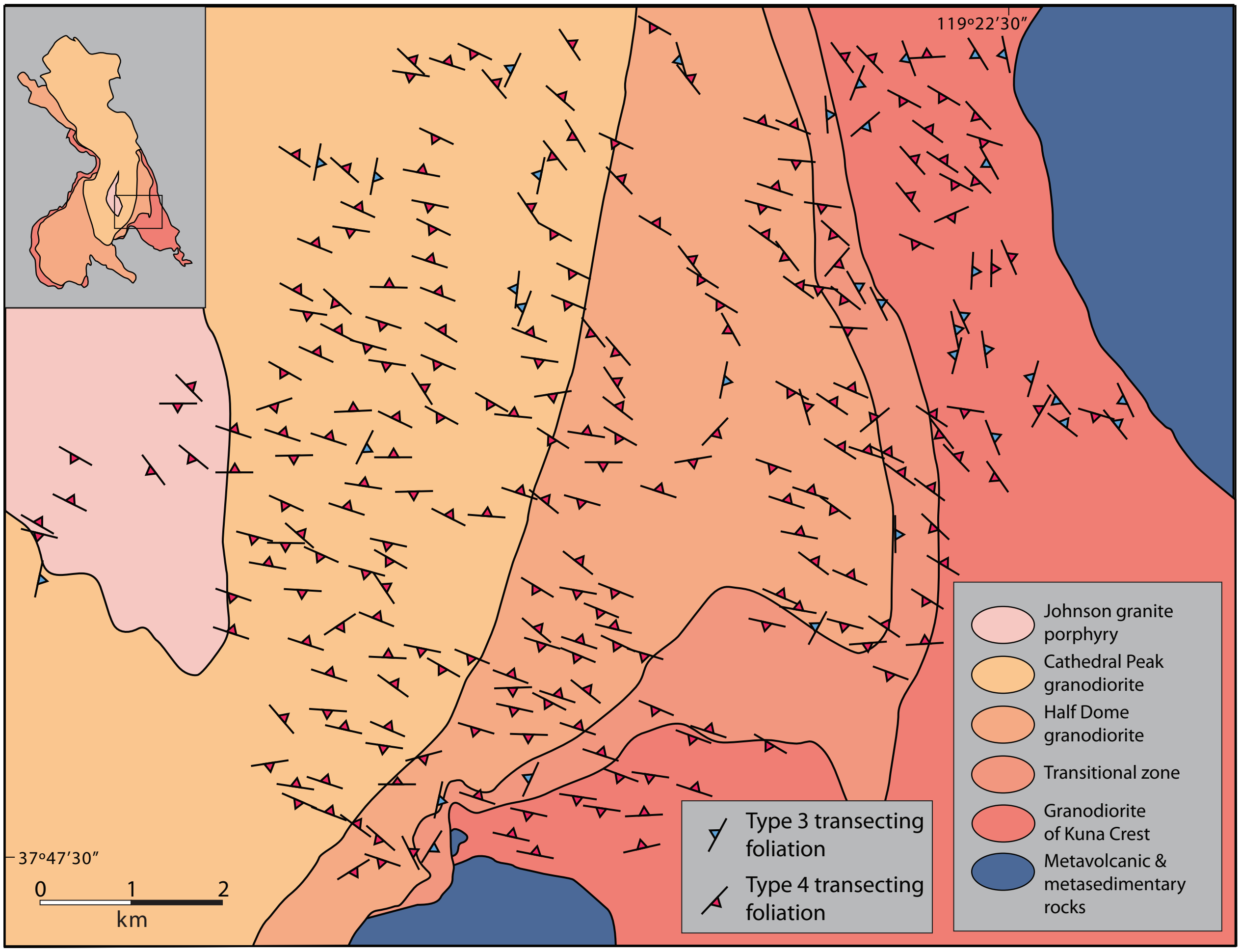


Figure 50. Sketch map illustrating two different transecting cleavages within the Tuolumne Intrusive series. These cleavages are clearly postmagmatic and are here interpreted to have been produced during two periods of folding, both of which are younger than the emplacement of the plutons. Figure modified from Žák et al. (2007).

augmented by upward doming of the plutonic roof; and Stein and Paterson (1996) theorized that a Jurassic compositionally zoned pluton in the range was emplaced by downward displacement and lateral ductile flow of wall rocks.

Figure 51 shows part of the polydeformed White-Inyo Range with its two major fold sets that together form a complex Type 1 interference pattern. Several folds are recumbently overturned (Ross, 1967). Note that most, but not all, of the Jurassic plutons have dominantly concordant contacts. That is, only one or two stratigraphic horizons sit along the plutonic contacts for many kilometers. Several plutons, such as the Beer Creek–Joshua Flat and the Marble Canyon bodies, occupy basins similar to those seen elsewhere in sedimentary units formed by refolding older folds. Given the degree of folding in the area, it is hard to imagine that these plutons could have exploited such a terrane after folding to create the lengthy concordant contacts. Additionally a close inspection of the northeast end of the Joshua Flat body (Fig. 51) clearly shows it wrapping around fold noses. The Marble Canyon body consists of many inwardly dipping sheets (Fig. 51), so is reasonably convincing as a refolded fold. The 83 Ma Papoose Flat pluton (Fig. 51) occupies the core of an anticline that Sylvester et al. (1978b) suggested represented an inflationary blister. However, close examination of the eastern end shows two E-SE trending extensions, that core anticlines separated by a syncline, which serve to suggest that this body is also folded. This is consistent with the detailed studies of Paterson et al. (1991), who suggested that there was significant regional deformation during and after emplacement of Cretaceous plutons in the area, and of Morgan et al. (1998), who used preexisting porphyroblast inclusion trails in the roof rocks to demonstrate that the pluton was initially a concordant sheet-like body. Additionally, quartz diorite sills (Fig. 51), interpreted to be of Cretaceous age (Nelson, 1971), are clearly folded and faulted along with their wall rocks.

All of these observations open the possibility that even the Cretaceous plutons of the region are folded and that their roofs and floors now have the same trend as the batholith and its septa. In fact, based on observations elsewhere (Hildebrand et al., 2010b), the contacts between plutons and wall rocks might,

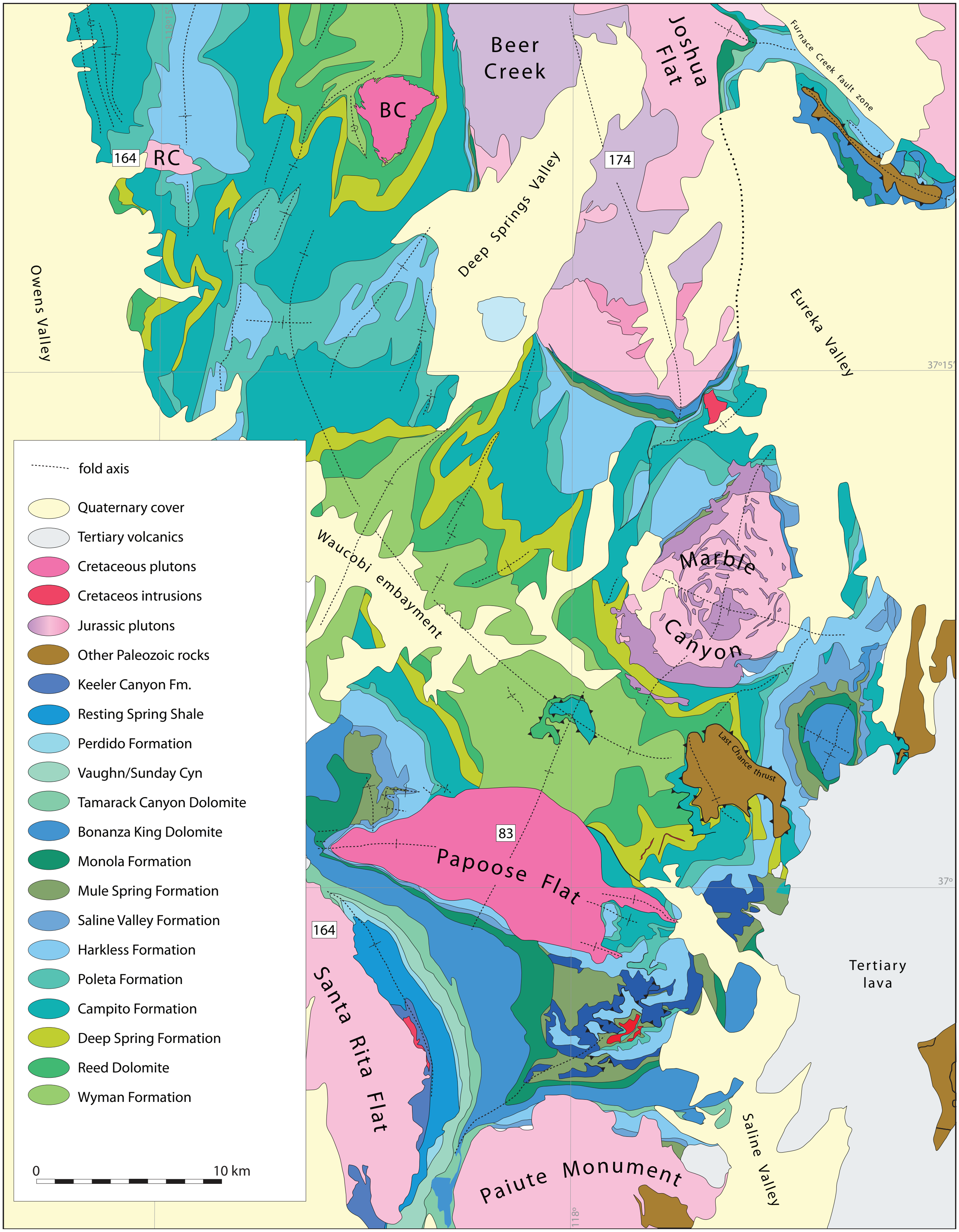

Beer Creek
Joshua Flat
Furnace Creek fault zone
BC
RC
164
174
Deep Springs Valley
Owens Valley
Eureka Valley
37°15'
Marble Canyon
Waucobi embayment
Last Chance thrust
83
Papoose Flat
37°
164
Santa Rita Flat
Tertiary lava
Saline Valley
Paiute Monument
118°
fold axis
Quaternary cover
Tertiary volcanics
Cretaceous plutons
Cretaceos intrusions
Jurassic plutons
Other Paleozoic rocks
Keeler Canyon Fm.
Resting Spring Shale
Perdido Formation
Vaughn/Sunday Cyn
Tamarack Canyon Dolomite
Bonanza King Dolomite
Monola Formation
Mule Spring Formation
Saline Valley Formation
Harkless Formation
Poleta Formation
Campito Formation
Deep Spring Formation
Reed Dolomite
Wyman Formation
0
10 km

Figure 51. Simplified geological map of the central White-Inyo ranges, California, showing the interference pattern created by the two periods of folding. Note that both Jurassic and Cretaceous plutons appear to have been folded as the plutonic contacts are nearly everywhere concordant with the wall rocks, which would be exceedingly difficult, if not impossible, to produce if the plutons intruded already folded rocks. The Marble Canyon pluton clearly occupies a structural basin indicating that it predates both periods of folding. The eastern margin of the Papoose Flat pluton is readily interpreted to be folded as its contact closely follows the folds of bedding. Map based on original mapping by Nelson (1966a, 1966b, 1971). BC—Birch Creek pluton; RC—Redding Canyon pluton (Coleman et al., 2003). Ages from Miller (1996), Chen (1977), and Coleman et al. (2003).

depending on the depth of emplacement, commonly have been horizontal, so that now many of the contacts between plutons and wall rocks represent tilted roofs and floors of tabular plutons (see also Hamilton, 1988a). Based on map patterns and foliations described earlier, as well as the relations just to the east in the White-Inyo Mountains, it is likely that the post–100 Ma Sierran Crest intrusions are folded as well.

Lending credence to the folding model for Sierran and White-Inyo plutons is the well-documented and doubly-folded composite Mount Stuart batholith, which is located within the High Cascades of Washington (Fig. 37) and contains plutons as young as 91 Ma (Matzel et al., 2006; Paterson and Miller, 1998). Similarly, within the Coast plutonic complex of British Columbia, plutons as young as 84 Ma (Figs. 34 and 35) appear to be folded along with slightly older folded bodies in the age range 104 to 94 Ma (Brown et al., 2000; Brown and McClelland, 2000).

An implication of the folded hypothesis for the Sierra Nevada is that many plutons exposed along the eastern margin, and originally suggested to have been composite dike-filled bodies, such as the McDoogle (Mahan et al., 2003) and Jackass Lakes (McNulty et al., 1996), were possibly emplaced as sill complexes rather than dikes. Also implied in the hypothesis, and supported by mapped relations in the Triple Divide Peak area (Moore and Sisson, 1987; Moore, 1981), where Independence dikes appear to be folded along with plutons and framework rocks (Fig. 52), is that the main swarm of dikes may have been emplaced as sills, rather than dikes as previously suggested (Moore and Hopson, 1961; Moore, 1963).

In general, the preservation of only small amounts of Cretaceous cover sitting unconformably on older Mesozoic and Paleozoic rocks throughout the Sierran batholith, suggests to me that many of the plutons had their roofs close to the contact of the volcanic cover with its Jurassic basement. During subsequent uplift, the less resistant volcanics were eroded leaving the more resistant plutonics and their metamorphic wall rocks. The Minarets caldera of the Ritter Range pendant (Fiske and Tobisch, 1994) is probably preserved because the core was down dropped along its ring fracture faults.

If the plutons are folded, then the age of the main phase of deformation must be younger than the youngest folded rock in the region, which is ~83 Ma in the White-Inyos. As discussed earlier, the two youngest foliations in complexes such as the 95–84 Ma Tuolumne intrusive series cut across contacts suggesting deformation after emplacement (Bateman et al., 1983b; Žák and Patterson, 2005; Žák et al., 2007), as do deformed dikes emanating from younger plutons, such as those related to the 92 Ma Lamarck granodiorite (Coleman et al., 2005). If plutons of the Sierran Crest magmatic event, the youngest magmatism of the Sierran batholith, are folded, then the deformation would appear to be younger than ~85–83 Ma, making it most likely Laramide in age.

Data from the southern Sierran batholith in the eastern Tehachapi Mountains support this conclusion. There, probable Campanian–Maastrichtian sedimentary rocks of the Witnet Formation, known to sit unconformably on 92 Ma Sierran granodiorite (Fig. 45), were overthrust and folded, even locally overturned, along northward-vergent thrusts carrying Sierran granitoid rocks prior to the Miocene (Lechler and Niemi, 2011; Wood, 1997). This suggests that those folds and thrusts were formed during the Late Cretaceous Laramide orogeny and that the cross folds of the Sierra were formed at that time.

For the most part, paleomagnetic poles of plutonic rocks ranging in age from 100 to 83 Ma within the Sierra Nevada form a tight array and if taken at face value would seemingly preclude postemplacement folding (Frei et al., 1984; Frei, 1986; Hillhouse and Groomé, 2011), but as 100 Ma volcanic rocks are clearly folded, I suspect that the paleomagnetic results reflect younger resetting or re-equilibration due to elevated heat flow from continued intrusion until just before 80 Ma (Dumitru, 1990). If the poles set in at near 80 Ma, it would preclude large-scale latitudinal motion of the Sierran block with respect to the Great Basin region afterwards but not before.

Doubling of Cordilleran Batholiths in the Canadian Sector

It has long been recognized (Monger et al., 1972, 1982) that there are two parallel belts of Cordilleran-type batholiths within the Canadian sector: the Omineca belt, located in the east and the Coast plutonic complex, located along the Pacific Ocean (Figs. 5 and 22). Monger et al. (1982) attributed the two belts to docking, or collision, of their two superterranes, Insular and Intermontane, with North America. Here I adhere to the idea that such magmatism results from subduction and that the two belts are parts of continental arcs. In the generally accepted Cordilleran paradigm, plutons of the Omineca belt intrude North American crust and resulted from eastwardly directed thrusting, but here I argue that they resulted from westerly-directed subduction of North America beneath the Rubian ribbon continent. Whatever model one chooses, there must be a major fault that lies somewhere within the Canadian Cordillera along which the two batholithic belts were doubled. However, as we have seen, nearly every terrane within the Canadian sector was joined to adjacent blocks before the 125–80 Ma magmatism of the Cordilleran batholiths. The only plausible location for such a fault, which

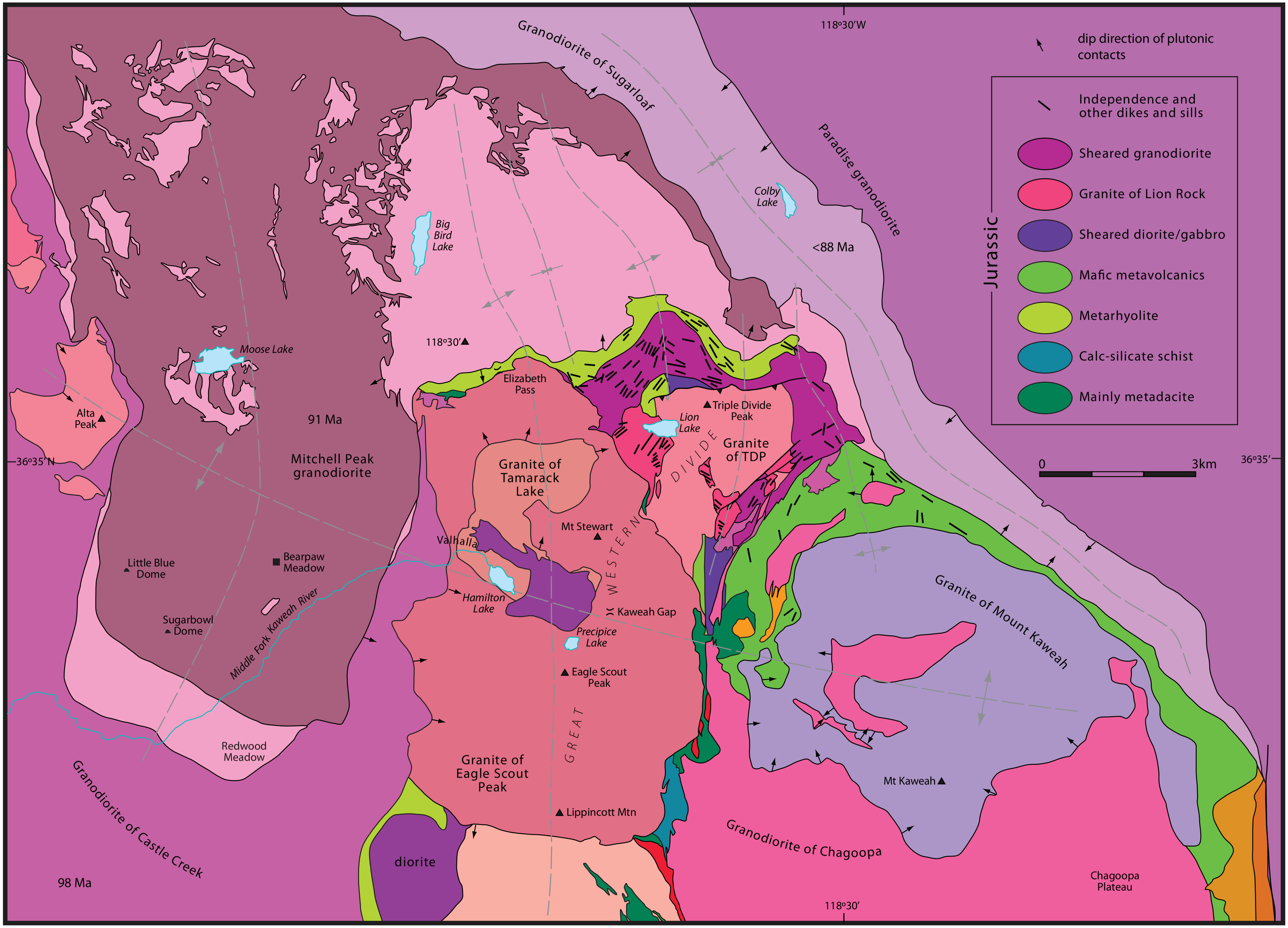


Figure 52. Geological sketch map of part of the Sierra Nevada batholith in Sequoia National Park illustrating the stacked nature of folded sill-like plutons. Note the change in strike of sills on opposing limbs of folds in the Triple Divide Peak area. Based on mapping by Moore and Sisson (1987) and Sisson and Moore (1994).

must lie between the two batholithic terranes, is along the eastern boundary of the Cache Creek terrane, bounded on that side by Cretaceous and Tertiary dextral strike-slip and oblique-slip faults (Wheeler and McFeely, 1991; Gabrielse, 1985; Struik et al., 2001). This fault would have at least 1500 km of separation and be one of the largest faults in the entire Cordillera. It could have been active from the Aptian (Sevier collision) to the Eocene, but its age is poorly constrained.

Just to the west of the proposed strike-slip fault, originally hypothesized to exist on the basis of paleomagnetic data and called the Intra-Quesnellia fault by Irving et al. (1995, 1996), lies the Sustut basin, which is a linear basin that sits on Stikinia and Cache Creek terranes and that parallels the eastern contact of the Cache Creek terrane, where the proposed fault would lie (Fig. 5). The basin, which contains more than 2000 m of coarse terrigenous clastics ranging in age from Aptian–Albian to Campanian (Evenchick and Thorkelson, 2005; Evenchick et al., 2007), seems out of place in the middle of the Cordilleran collage, but it logically can be explained to have formed along a bend or splay of the fault. Similarly, the 89 ± 2 Ma Table Mountain volcanic suite and the 85 ± 5 Surprise Lake batholith, both of which occur in the northern outcrop belt of Cache Creek terrane (Mihalynuk et al., 1992), seem to be out of place.

Within the southern Cordillera and spanning the British Columbia–Washington border are several small terranes of different ages and origin, such as the Chilliwack, Bridge River, Easton, Cadwallader-Tyaughton-Methow, and many more preserved only as structural slices (Monger and Struik, 2006) or onlap successions (MacLaurin et al., 2011). They haven't been described here because they are only peripheral to our story, but as they are caught between southern Wrangellia and Stikinia, they do bear on the assembly of the westernmost terranes. The slices are variously interpreted to represent bits of oceanic crust, open-ocean cherts, fore-arc basins, and blueschists of varying ages ranging from Paleozoic to Mesozoic (Monger et al., 1982, 1994; Umhoefer et al., 2002). Farther north the Wrangellia-Stikinia/Yukon Tanana contact is variously considered to be Jurassic or Cretaceous, but most workers agree that the Gravina-Dezadeash-Nutzotin belt, which represents a narrow basin developed between the two blocks, was closed by ~100 Ma (Haeussler, 1992; Rubin et al., 1990; Journeay and Friedman, 1993; McClelland and Mattinson, 2000). Additionally, paleomagmetic data (Fig. 53) indicate the Wrangellia and Stikinia were never very far apart latitudinally (Kent and Irving, 2010). Thus, if the basin was narrow, it is difficult to understand how the various oceanic and fore-arc blocks might have been incorporated between Insular and Intermontane superterranes. One compelling solution to the conundrum was presented by Monger et al. (1994), who suggested that there was a period of sinistral displacement on an intra-arc fault that transported the northerly Coast plutonic complex southward to double the width of the Coast plutonic complex in the south, which would explain its exceptional width. This migration overlapped and trapped fore-arc rocks previously outboard of the more southerly parts of the arc. This model was subsequently utilized by Gehrels et al. (2009) to explain doubling of the Coast plutonic complex and closing of the Gravina-Nuzotzin basin by ~110–100 Ma. Irrespective of how this is resolved, it is clear that the Insular and Intermontane

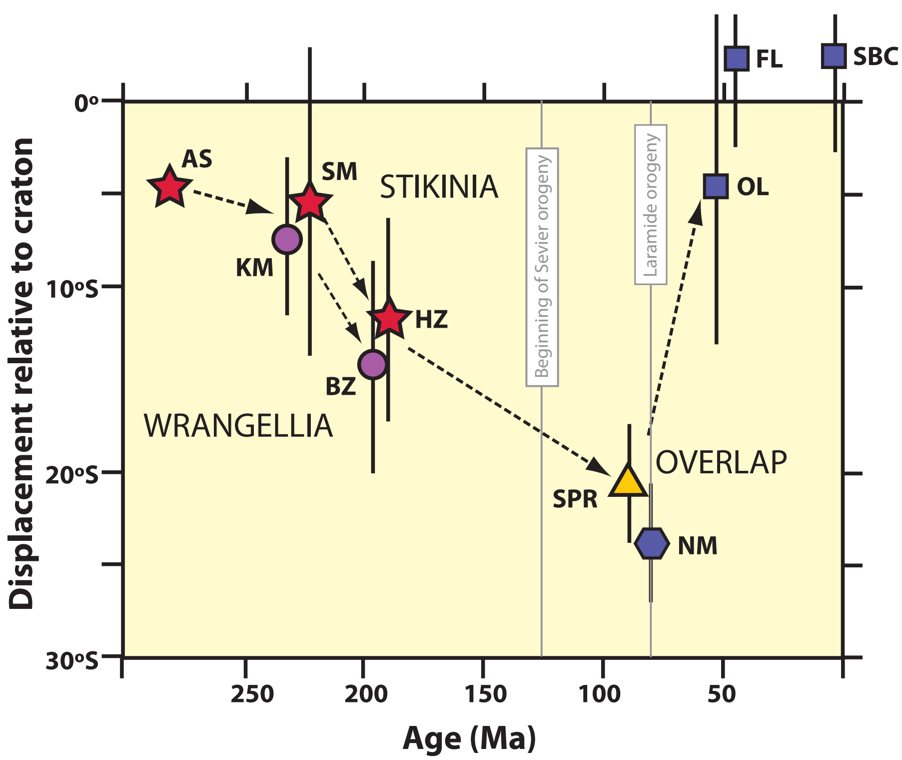


Figure 53. Latitudinal displacements with 95% error bars of selected formations of southern Wrangellia and Stikinia with respect to North America. Note that relative to North America the Rubian ribbon continent migrated south until 80–75 Ma when its relative displacement reversed to dextral. This means that after initial collision during the Sevier orogeny at ~125 Ma, Rubia migrated southward, then, after the Laramide event, migrated northward relative to cratonic North America. as—Permian Asitka Group; Triassic-Jurassic units: km—Karmutsen lavas (~225 Ma); sm—Savage Mountain (~210 Ma); hz—Hazleton Group (~195 Ma); bz—Bonanza arc (195 Ma); Cretaceous units: spr—Silverquick–Powell Creek; nm—Nanaimo Group of Vancouver Island; Cenozoic units: fl—Flores volcanics of Vancouver Island; ol—Ootsa volcanics of central British Columbia; sbc—late Neogene basalts of southern British Columbia. Modified from Kent and Irving (2010).

superterranes were joined by 100 Ma, completing the amalgamation of the Rubian ribbon continent.

The Laramide Event: ~80–75 Ma Deformation and Metamorphism

The ~80–75 Ma deformation and metamorphism that occurs throughout western North America is generally included within the Laramide event. Several scientists, myself included (Armstrong, 1968; Perry and Schmidt, 1988; D.M. Miller et al., 1992; Hildebrand, 2009), restricted the term Laramide structure, or orogeny, to the thick-skinned belt of the Colorado Plateau region, but deformation of that age occurs from South America to Alaska, and it appears reasonable to treat it all as the same event. Drewes (1978, 1991) used the term Cordilleran orogenic belt to refer to the entire belt of Jurassic to Eocene deformation in both North and South America, but his work centered on the Late Cretaceous–early Tertiary sector of the orogen from California to Texas. Hildebrand (2009) used the term Cordilleran orogeny for the Lower Cretaceous–early Tertiary deformation throughout the orogen from Alaska to southern Mexico and perhaps beyond. No name will be satisfactory to everyone, and there are obvious merits to the historical precedents, so that the terms, Laramide and Cordilleran should be retained. Here I use the term Laramide event to refer to all of the Late Cretaceous–early Tertiary deformation and metamorphism. Overall, it includes a variety of deformational features ranging in age from Late Cretaceous to Eocene and includes nappes in the Mojave-Sonora region, thick-skinned basement-involved thrusts and development of localized basins superposed upon the more extensive linear foredeep in the Great Basin sector, thin-skinned thrusting in the Canadian sector, eastern Mexico, Central America, and northern South America, and some folds in the Sierra Nevada region. Used in this manner it clearly separates this event from the earlier Sevier event in the Great Basin segment, and, as rocks unequivocally part of North America located within the Canadian and Sonoran sectors are largely unaffected by the older deformation, there should be little confusion. However, as it was locally ongoing, all of the Cretaceous–Tertiary deformation can be included together in the Cordilleran orogeny.

The classic area of Laramide deformation is located in the Great Basin sector of the orogen, bounded on the north by the Lewis and Clark lineament, and on the south by the Phoenix fault (Fig. 8). There, the deformation is dominated by thick-skinned thrusts that involve cratonic basement of North America (Grose, 1974; Smithson et al., 1979; Brewer et al., 1982; Rodgers, 1987; Hamilton, 1988b). The Sierran-White-Inyo block is generally considered to not have obvious Laramide deformation, but based on the analysis presented earlier, I suggest that it has post–85 Ma folds, and so was deformed during the Laramide event. North of the Lewis and Clark lineament, within the Canadian sector, Laramide deformation is dominated by thin-skinned thrusts of the Rocky Mountain fold-thrust belt; whereas to the south of the Phoenix fault as described earlier, deformation is highly variable, consisting of thick-skinned deformation involving basement in the Mojave–western Sonora region, basement-involved thrusts and metamorphism up to amphibolite grade in the Maria fold-thrust belt, and thin-skinned deformation in central-southern Mexico. In Central America and northern South America, deformation involved the obvious collision and accretion of the Great Arc of the Caribbean and its previously accreted oceanic plateau.

Ever since Coney and Reynolds (1977) summarized and interpreted radiometric data from California and southern Arizona—which suggested to them that arc magmatism swept inboard 1000 km at 80 Ma and then back again at ~45 Ma due to progressive slab shallowing followed by steepening—most models have suggested that the Laramide event was connected in some manner to shallowly dipping subduction beneath North America (Coney, 1976; Dickinson and Snyder, 1978; Bird, 1988; Hamilton, 1988a; Dumitru et al., 1991; Saleeby, 2003; Grove et al., 2003b; Jacobson et al., 2011). However, it appears that the trend of Laramide deformation and magmatism in southwestern North America has a trend much closer to E-W than N-S (Fig. 8), which means that Coney and Reynolds (1977) used data collected parallel to, or nearly along, strike of the belt rather than across it (Glazner and Supplee, 1982). Thus, it is not surprising that when plotted on distance-age plots, they form near-horizontal arrays.

Within convergent margins, the process of subduction itself—which some workers link to upper-plate compressional deformation (Monger and Nokleberg, 1996; Hutton, 1997; DeCelles, 2004; DeCelles and Coogan, 2006; DeCelles et al., 2009) despite considerable evidence to the contrary in the form of uncompressed fore-arc basins and regions (von Huene, 1984; Loveless et al., 2005); the low-standing, neutral to extensional nature of nearly every arc (Hamilton, 1981, 1988a, 2007; Hildebrand and Bowring, 1984; Hildebrand, 2009); and the dominant process of slab rollback—as dense oceanic plates sink vertically into the mantle (Elsasser, 1971; Karig, 1971; Molnar and Atwater, 1978; Garfunkel et al., 1986; Hamilton, 1985, 2007; Schellart et al., 2006; Clark et al., 2008)—cannot produce the kind of deformation observed or its extent. For example, how could sinking oceanic crust transmit enough force to move and lift the Belt-Purcell allochthons, some 450 × 200 × 25 km up and over the entire passive margin and out over Cretaceous rocks of the foredeep?

Additionally, as pointed out by others (Maxson and Tikoff, 1996; English et al., 2003), such upper-plate deformation requires the complete erosion of mantle lithosphere to transmit compressive stresses, yet isotopic studies of Cenozoic mantle xenoliths suggest its continued presence (Farmer et al., 1989; Livaccari and Perry, 1993; Lee et al., 2000, 2001). Saleeby (2003) recognized this problem so developed a segmented model in which only a narrow region of the subducting plate was shallow; however, the deformation is far more widespread, extending from South America to Alaska, than predicted by his model.

Alternatively, and shown to be critical to generating flat-slab subduction by Gutscher et al. (2000), Laramide deformation was hypothesized to have been caused by the collision of a high-standing oceanic plateau with the North American continent (Livaccari et al., 1981; Henderson et al., 1984; Saleeby, 2003). Based on ideas of Barth and Schneiderman (1996), the most recent model involves slightly different age collisions of hypothetical conjugates to the Hess and Shatsky rises (Liu et al., 2010), now located in the NW Pacific region. In the recent models the plateaus were mostly to entirely subducted, leaving little or no material behind, which makes the models difficult to test. However, there are several such collisions taking place elsewhere today so it is possible to study the effects of plateau collisions based on actual observations.

The most spectacular example of an impinging oceanic plateau occurs in the Solomon Islands of the SW Pacific. There, Pliocene collision of the Solomon arc with the extensive and 33-km-thick Ontong-Java plateau—a basaltic plateau formed during the Mesozoic, covered with at least 1000 m of pelagic sediments, and that sits on Pacific oceanic crust—apparently led to initiation of a new NE-dipping subduction zone along what is now the Woodlark basin (Hamilton, 1979, and references therein). The basin contains oceanic plateaux and a subducting spreading ridge. Geophysical studies (Miura et al., 2004; Phinney et al., 2004) demonstrated that during the collision only the upper frontal part of the plateau was detached, imbricated, and accreted to the leading edge of the Soloman arc within the Malaita accretionary complex (Hughes, 2004), while the lower and main bulk of the plateau continues to be subducted (Phinney et al., 2004; Mann and Taira, 2004; Taira et al., 2004). This model was substantiated by Hf-Nd-Pb isotopic data that revealed an old Pacific and Ontong Java component to arc magmatism in the Solomon arc (Schuth et al., 2009). Perhaps surprising for those who favor large-scale, extensive, upper-plate deformation related to plateau subduction, a detailed study of the Central Solomon intra-arc basin, active throughout the period of plateau accretion, shows only limited folding and faulting (Cowley et al., 2004). Only rollback can account for the subduction of such a thick plateau.

In a review of the effects of six different collisions along the western Americas, von Huene and Ranero (2009) showed that, although the arrival of thick blocks such as the Yukutat block (Perry et al., 2009; Worthington et al., 2012), an oceanic plateau, in southeastern Alaska can locally create high mountains and strong deformational features, the dominant processes related to collision of buoyant, high-standing basaltic areas within oceanic tracts are (1) complete subduction of the oceanic plate even where it has high relief, (2) localized and short-lived uplift of the upper plate, and (3) local subduction erosion of the accretionary prism. They concluded that impingement of high-standing oceanic features generally has low interplate friction, create upper-plate relief that is ephemeral, and that in order to create high coastal mountains, there must be long-lasting subduction of very extensive relief in one area, such as occurs where the Cocos Ridge is currently being subducted beneath Central America. This effectively rules out areas with strongly oblique convergence.

A recent study of subduction beneath the Trans-Mexican volcanic arc demonstrated that subduction becomes progressively flatter eastward until it appears to directly underlie the crust nearly 250 km inboard of the Guerrero terrane, thus explaining the acute divergence of arc magmatism from the orientation of the Middle American trench (Pardo and Suárez, 1995; Pérez-Campos et al., 2008). While there is apparently some interplay between the subducting plate and the North American crust as detected by the global positioning system (Payero et al., 2008), there doesn't seem to be deformation similar to that of the Laramide event or exceptionally thick crust. The flat subduction does lead to a wide zone of Quaternary magmatism (Blatter et al., 2007), and when combined with the observations that slab-derived, subducted water appears in both the asthenospheric wedge and resultant volcanic rocks (Jödicke et al., 2006; Johnson et al., 2009; Chen and Clayton, 2009), they suggest that flat slabs are able to transport volatiles well inboard of the normal arc-trench distance and still produce copious quantities of arc magmatism. Similarly, the flat-slab subduction of the Yukutat oceanic plateau, beneath North America in southeastern Alaska, has produced copious magmatism to form spectacular stratocones and shield volcanoes (Neal et al., 2001; Richter et al., 1995, 2006; Trop et al., 2012). Thus, even if a flat-slab model for the Great Basin was viable, there seems no obvious reason for the lack of subduction-related magmas there during the Laramide event.

Laramide Collision along the Orogen

The Laramide event occurred nearly synchronously over the length of the orogen as deduced from the age of thrusting and development of related orogenic foredeeps, although difficulties in accurately dating the thrusting from area to area and minor irregularities in the shape of the margin complicate the story. Based on the similarity in deformational ages, the collision between Rubia and North America must have been nearly orthogonal. This strongly suggests that the Rubian ribbon continent was one entity at that time.

In Alaska, rocks of the Valenginian–Cenomanian Kahiltna basin (Fig. 38) were thrust northward at ~74 Ma, coincident with development of the Campanian–Maastrichtian Cantwell basin, a thrust-top basin formed during the thrusting (Ridgway et al., 2002; Trop and Ridgway, 2007). Late Cretaceous–early Tertiary northward-vergent thrust and folds deformed Early Cretaceous features in northern Alaska (Moore et al., 1997).

In the Canadian segment, rocks of the North American continental terrace were separated from their basement along a detachment located within Cambrian shales, folded, and thrust eastward to form the Rocky Mountain fold-thrust belt (Figs. 5 and 32) during the Late Cretaceous–early Tertiary (Price and Mountjoy, 1970; Price, 1981; Price and Fermor, 1985; Fermor and Moffat, 1992). A thick clastic wedge of Campanian–Paleocene age developed to the east in the foreland basin during this deformation (Larson et al., 2006; Ross et al., 2005).

Thrusting was apparently ongoing from Sevier to Laramide time within the North American margin of the Great Basin sector, but it was much subdued (for example, DeCelles and Coogan, 2006; Yonkee and Weill, 2011). Intense deformation and metamorphism occurred within the hinterland region between 85 and 75 Ma (Camilleri et al., 1997; McGrew et al., 2000) and the classic thick-skinned deformation of the Colorado Plateau region started during the Maastrichtian (Dickinson et al., 1988; Lawton, 2008).

A continuous Late Cretaceous–Paleocene foreland fold-thrust belt and related foredeep (Fig. 5) occur throughout north-central and eastern Mexico just west of the Gulf of Mexico (Eguiliz de Antuñano et al., 2000). They formed during the accretion of the Guerrero superterrane (Tardy et al., 1994; Centeno-Garcia et al., 2008, 2011), the south-central part of Rubia. The western margin of Oaxaquia and Mixteca were deformed in the Late Cretaceous–early Tertiary in a dominantly east-vergent fold-thrust belt (Suter, 1984, 1987; Hennings, 1994; Fitz-Díaz et al., 2012) and the Tampico-Misantla foredeep developed in front of the advancing thrusts (Busch and Gavela, 1978). To the south is the Zongolica fold-thrust belt, which involved thrusting of deeper-water sedimentary rocks eastward over the reefal carbonate-dominated Cordoba platform during the Santonian–Campanian (Nieto-Samaniego et al., 2006).

In the Cuicateco terrane of southern Mexico, Maastrichtian schists, greenstones, gabbros, and serpentinites were thrust eastward over red beds of the Maya terrane during the latest Cretaceous–Paleocene (Pérez-Gutiérrez et al., 2009). At the southern end of the Maya block, a west-facing carbonate-dominated platform sitting on basement of the Maya block was drowned during the uppermost Campanian, buried by orogenic flysch during the Maastrichtian–Danian (Fourcade et al., 1994), and overthrust by ultramafic nappes. Rocks of the lower-plate crystalline basement were metamorphosed to eclogite at 76 Ma, which implies that part of the North America margin was subducted to greater than 60 km depth at about that time and exhumed to amphibolite grade a million years later (Martens et al., 2012), presumably by slab failure.

Even farther south in the rotated Chortis block, Rogers et al. (2007) documented a Late Cretaceous belt of southeast-dipping imbricate thrusts, which they interpret to represent the accretion of the Caribbean arc system to the Chortis block (see also, Pindell et al., 2005; Pindell and Kennan, 2009; Ratschbacher et al., 2009). The arc-bearing block (Fig. 54) continues through its diachronous collision zone with Bahamian Bank of North America represented on Cuba and Hispaniola, through the Virgin Islands (Schrecengost, 2010) to its still active Antillian segment before reaching northern South America, where it was diachronously deformed along the coastline from west to east (Ostos et al., 2005). That the Antillean arc is part of the Great Arc is supported by the presence of Jurassic oceanic basement and chert at La Desirade (Mattinson et al., 2008; Montgomery and Kerr, 2009). In Ecuador and Colombia, the arc collided with the western margin of South America above a westward-dipping subduction zone during the Campanian–Maastrichtian and was followed by eastward subduction of Pacific oceanic lithosphere (Jaillard et al., 2004; Luzieux et al., 2006; Vallejo et al., 2009; Altamira-Areyán, 2009). Thus, the Great Arc was also part of the Rubian ribbon continent (Fig. 55).

Overall, the Laramide orogeny was more or less synchronous from Alaska to northern South America, which effectively rules out flat-slab subduction or plateau subduction as its cause. Furthermore, in nearly all locations it can be demonstrated that North America was the lower plate in a collision with an arc-bearing block, interpreted here to be the Rubian ribbon continent. Therefore, models that treat the thick-skinned deformation in the Great Basin segment in isolation without accounting for the complete extent of the contemporaneous deformation (Saleeby, 2003; Liu et al., 2010; Jones et al., 2011) aren't likely to be successful.

Instead, it appears as though a collisional model, in which the Rubian ribbon continent collided with the western margin of the Americas, best fits the data. This ribbon continent included all the previously amalgamated terranes from Alaska to South America, including the Guerrero composite terrane and the Great Arc of the Caribbean. Plate motion reconstructions show that during this period there was a strong northerly component to the motions of the plates within the Pacific basin (Doubrovine and Tarduno, 2008).

During and following collision, subduction beneath the Franciscan and related accretionary complexes, such as the Chugach, stalled, the slab failed, a pulse of slab-failure magmatism rose into the collisional zone, Franciscan high-grade rocks were rapidly exhumed, and northward transpression drove a mass of previously amalgamated terranes within Rubia northward, where they impinged on cratonic North America to form the Rocky Mountain fold-thrust belt. This scenario is similar in many respects to those proposed by Maxson and Tikoff (1996) and Johnston (2001). We next explore some of the consequences of the Laramide collision.

Cretaceous–Early Tertiary Slab-Failure Magmatism

The distribution of Late Cretaceous–early Tertiary magmatism in the Southwest was treated by Coney and Reynolds (1977), but they assumed that the continental margin was oriented more or less N-S, and so developed a model where magmatism, derived from an eastwardly-dipping slab, swept inboard starting at 120 Ma then outboard again at 40 Ma. However, as we have seen, based on deformation and magmatism, this part of the margin is oriented today more or less E-W and faces south, which means that magmatism was more or less synchronous across the region during the period 80–60 Ma. This belt includes the so-called Laramide magmatism of southern Arizona and New Mexico, the post–80 Ma adakitic-alkaline plutons of the Mojave area, and the linear belt of 76–55 Ma plutons that continue southward through much of western Mexico (Damon et al., 1983; Zimmermann et al., 1988; Titley and Anthony, 1989; Barton

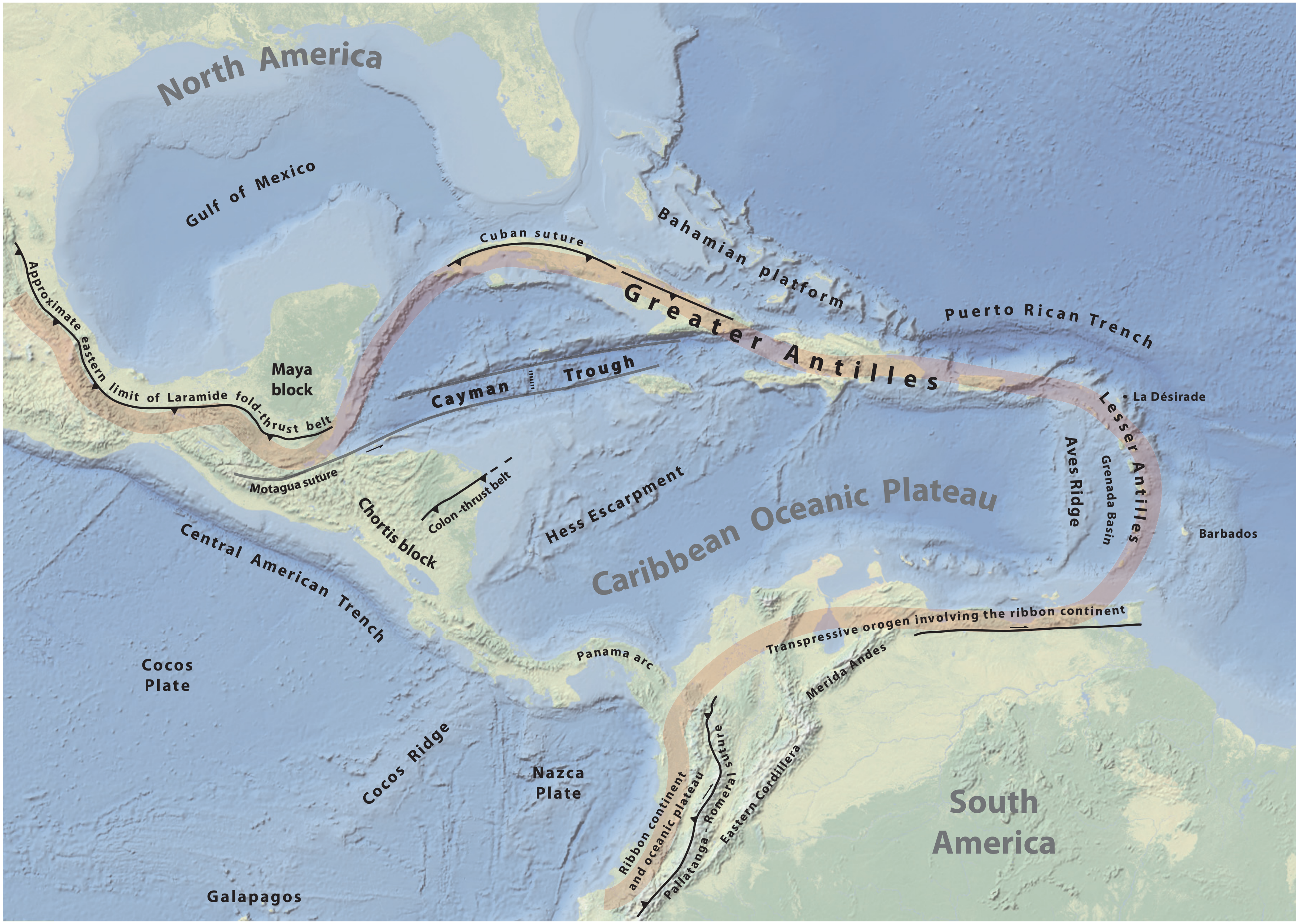


Figure 54. Shaded relief map of the Caribbean area showing approximate distribution of Great Arc rocks (gray band) and other features as mentioned in text.

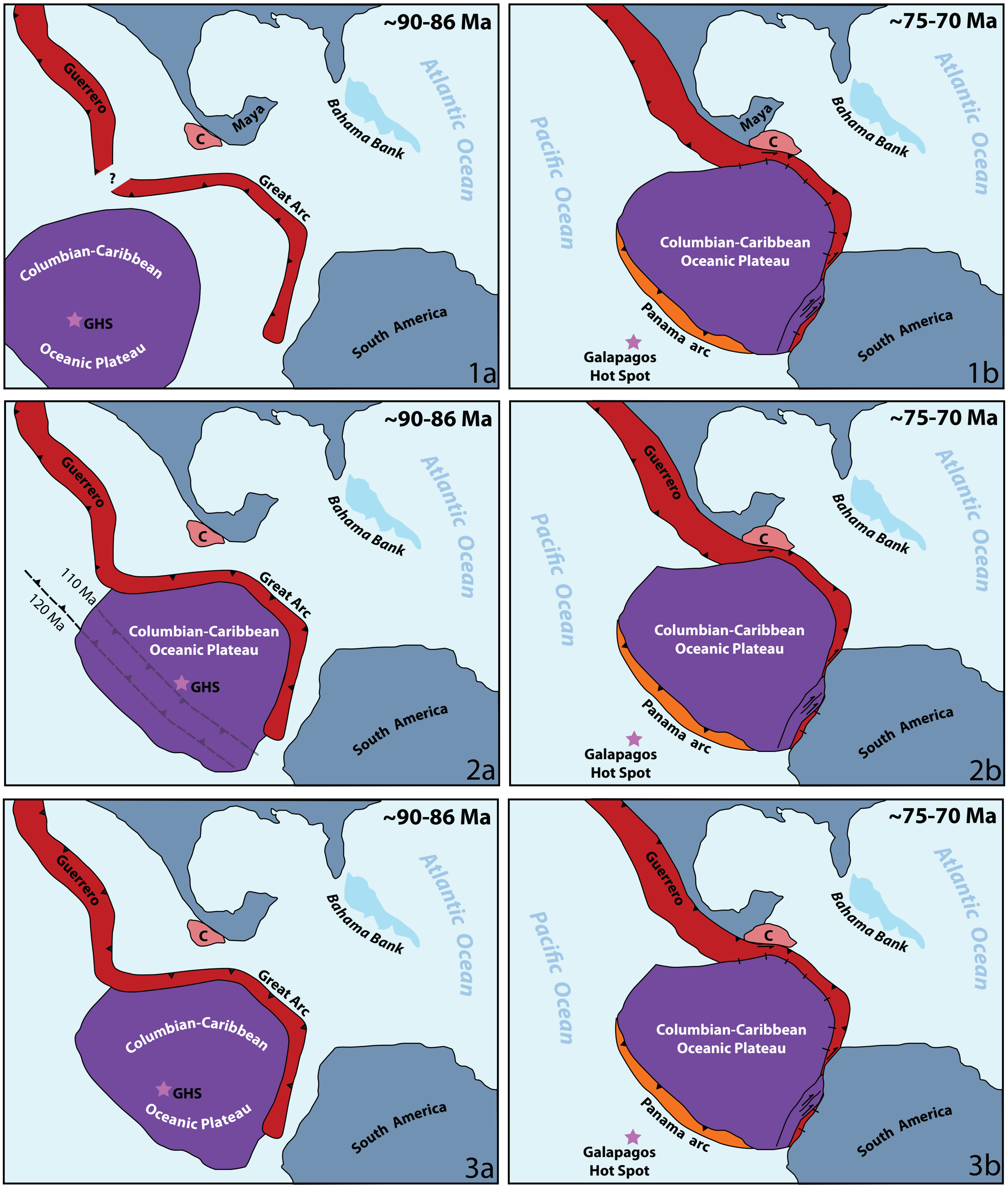


Figure 55. Three different models for the development of the Caribbean Sea. In (1), eastward-dipping subduction beneath the Great Arc leads to collision of the Columbian-Caribbean Oceanic Plateau (CCOP) and subduction reversal beneath the arc (Burke, 1988; Rogers et al., 2007; Altamira-Areyán, 2009), which then migrate together between North and South America into the Atlantic. In (2), the collision between the arc and CCOP takes place between 120 and 110 Ma and subduction flipped at that time (Pindell et al., 2005). The main problem with both (1) and (2) is that subduction beneath the Guerrero arc-bearing superterrane is eastwardly-dipping following collision with the CCOP, but in Mexico the superterrane was the upper plate above a westward-dipping subduction zone during the Laramide event, which would require an additional flip in subduction polarity. In (3), the model preferred here, subduction was always westward-dipping beneath the Guerrero superterrane and the Great Arc, which in part was built on the eastern margin of the CCOP; thus the arc didn't reverse its subduction polarity.

et al., 1995; McDowell et al., 2001; Henry et al., 2003; Valencia-Moreno et al., 2006, 2007; Ramos-Velázquez et al., 2008).

Large numbers of porphyry copper deposits occur in the Sonoran segment and along with their plutonic sources formed between ~75 and 55 Ma (Titley and Anthony, 1989; Titley, 1982; Damon et al., 1983; Barton et al., 1995; Barra et al., 2005; Valencia-Moreno et al., 2006, 2007). Many of the intrusions are not accurately dated by zircons, so their precise ages are not well known; nevertheless, the 65 ± 10 Ma age suggested to Hildebrand (2009) that the porphyry deposits in the Sonoran segment might be related to slab failure as they appear to have been in the Alpine belt (de Boorder et al., 1998), Central Range orogeny of Papua New Guinea (Cloos et al., 2005; McDowell et al., 1996), and elsewhere in the southwest Pacific (Solomon, 1990).

In the Coast plutonic complex of British Columbia, Jurassic to Late Cretaceous tonalitic-granodioritic plutons, deformed and metamorphosed to gneiss under amphibolite-granulite conditions, and generally considered to constitute the lower and middle crust of a Cordilleran-type magmatic arc, were rapidly exhumed after arc magmatism ceased (Armstrong, 1988; Hollister, 1982; van der Heyden, 1992; Crawford et al., 1999). Extension, which took place by at least 60 Ma (Armstrong, 1988), and involved at least 15 km of tectonic exhumation, was accompanied by a voluminous Late Cretaceous–early Tertiary intrusive bloom of plutons, including adakitic and A-type bodies, derived from multiple sources (Hollister and Andronicos, 2000, 2006; Hollister et al., 2008; Andronicos et al., 2003; Gehrels et al., 2009; Mahoney et al., 2009), the hallmark of slab-failure magmatism. The Coast Range magmatism is similar in terms of composition to magmatic rocks of the Tibetan Plateau, where postcollisional magmatism is varied, but dominated by adakitic and alkaline compositions that suggest Eocene Neo-Tethyan slab failure followed by progressive crustal uplift and northward shunting of hot mantle (Chung et al., 2003, 2005; Wang et al., 2010; Searle et al., 2011). The band of strongly alkaline and subalkaline rocks of the Montana alkalic magmatic province are intriguingly similar to young rocks on the north Tibetan Plateau, sit inboard of the main belt of slab-failure intrusions, and so may also reflect detachment of collisionally-thickened lithosphere and consequent upwelling of hot asthenosphere, as envisioned by Chung et al. (2005) for the Tibetan analog. The upwelling hot mantle might be a reasonable explanation for “fast” seismic velocities detected beneath the region (Schmandt and Humphreys, 2011).

The shutdown of arc magmatism just before the Late Cretaceous, the rapid exhumation of the central gneiss complex by at least 60–50 Ma, and the voluminous Late Cretaceous–early Tertiary magmatism are interpreted to represent the change from pre-collisional continental arc magmatism to syn- to postcollisional slab-failure magmatism and exhumation. Asthenosphere upwelling through the narrow tear as it formed best explains the highly focused nature of the elongate Late Cretaceous tonalitic magmas along the western margin of the belt (e.g., Barker and Arth, 1990). Overall, the linear belt, some 1500 km long, of Late Cretaceous–early Tertiary plutons within the Coast Range Complex may be the best-exposed example of slab-failure magmatism anywhere on Earth.

When the 80–70 Ma paleolatitude of the Coast plutonic complex is restored, its southern end matches up reasonable well with the Sonoran sector to create a continuous band of Late Cretaceous–early Tertiary slab-failure magmatism (Fig. 56). With such a reconstruction, the Swakane gneiss, now located at the southern end of the Coast plutonic complex, is colocated with the nearly identical Pelona and Orocopia schists. And it matches the strike-slip faults and 110–100 Ma basinal closure within the Peninsular Ranges, Idaho batholith, and Coast plutonic complex to suggest that they are segments of the same system. Thus, there are three independent geological lines of evidence that support such a restoration (Fig. 57).

The only Late Cretaceous–early Tertiary magmatism in the magmatic gap between the Phoenix and Orofino faults occurs in the Colorado Mineral Belt (Wilson and Sims, 2003), which trends SW-NE through the region (Figs. 8 and 22). The belt divides the Laramide basins into two distinct fields (Fig. 8) and appears to comprise magmas that Hildebrand (2009) suggested were generated by slab failure and likely mixtures of asthenospherically derived basalt and melted Proterozoic crystalline basement emplaced into the lower North American plate (Stein and Crock, 1990; Bailley and Farmer, 2007) through a tear in the North American lithosphere that developed along a long-standing lithospheric boundary (McCoy et al., 2005). More recently, with an excellent summary of basic data on the zone, Chapin (2012) suggested that magmatism of the belt was derived from a slab tear in an eastward-dipping subduction scheme, which is at odds with the marked lack of arc magmatism in North America at that time and the westward subduction model presented here. I suggest another possibility: that the magmatism of the Colorado Mineral Belt originated when the North American lithosphere cracked as the Canadian and Sonoran sectors were separated (Fig. 22). As Rubia had already collided with North America, the strong northward movement of the Canadian sector is envisioned to have dragged the Great Basin sector northward sufficiently to have opened a crack that allowed asthenospheric magmas access to the crust, where they melted and assimilated it to produce the observed magmatism.

Shutdown of Franciscan Subduction

The demise of Franciscan subduction at ~80 Ma, during or just after the Laramide event has gone largely unrecognized; yet three lines of evidence indicate that subduction must have stopped at that time: (1) the youngest known blueschist is 84 Ma; (2) there is an abrupt gap in sedimentation from ~80 Ma to 53 Ma; and (3) the coherent blueschists were exhumed after 84 Ma and at least locally at the surface by 67 Ma (Fig. 58). Thus, I divide the Franciscan into two parts: Franciscan 1 and 2, which were separated by a period of no subduction. During the intervening period, motion was largely northward along the margin, and both the Franciscan and the Great Valley rocks were

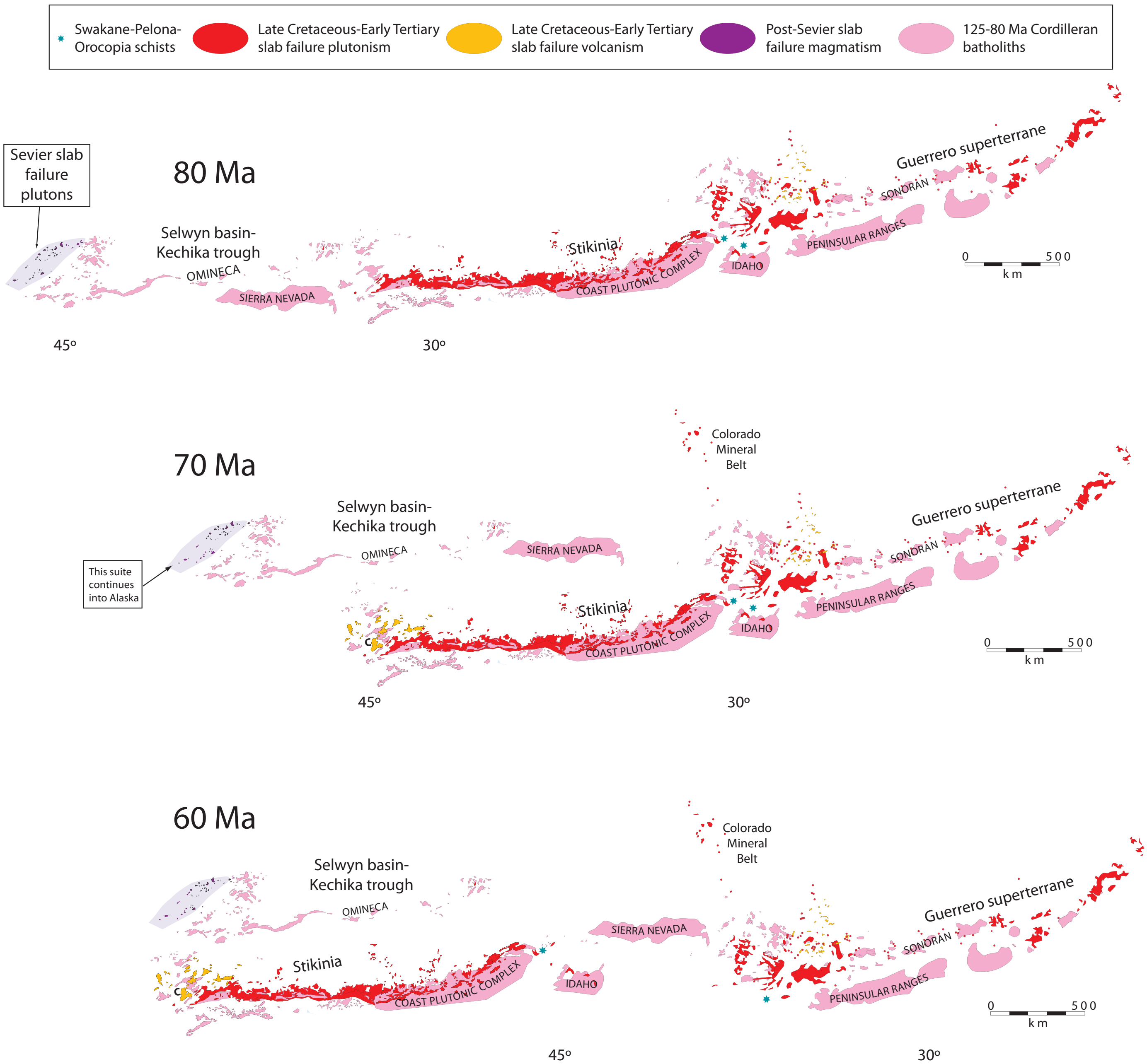


Figure 56. Tectonic model illustrating possible relative locations of various magmatic elements within the Canadian–Great Basin and Sonoran sectors during the Late Cretaceous–early Tertiary. The Omineca batholith and the Sierra Nevada batholith are different from the other Cordilleran-type batholiths in that they don't have postcollisional slab-failure magmatism and so are here considered to have been together; however, there is considerable uncertainty as to the paleogeographic location of the Sierra Nevada. c—Carmacks volcanics.

exhumed and uplifted. These ideas are similar to those developed by Jayko (2009).

Although the Coastal Franciscan is not well dated, it is most probably younger than ~52 Ma, the age of the youngest detrital zircons dated by Snow et al. (2010) in the San Bruno belt. Low-resolution data collected by Tagami and Dumitru (1996) support this finding. Similarly, most sedimentation within the Great Valley group appears to have waned or ceased during the Campanian–Maastrichtian between ~75 and 65 Ma (Ingersoll, 1979; DeGraaff-Surpless et al., 2002). Thus, the existing detrital zircon data from the accretionary prism, the absence of magmatism after 80 Ma, and the paucity of sedimentation with the Great Valley fore-arc basin indicate that eastward subduction stopped at ~80 Ma (Fig. 58). The shutdown of Franciscan subduction corresponds with the termination of magmatism within the Sierra Nevada, the end of Sevier thin-skinned thrusting, and the begin-

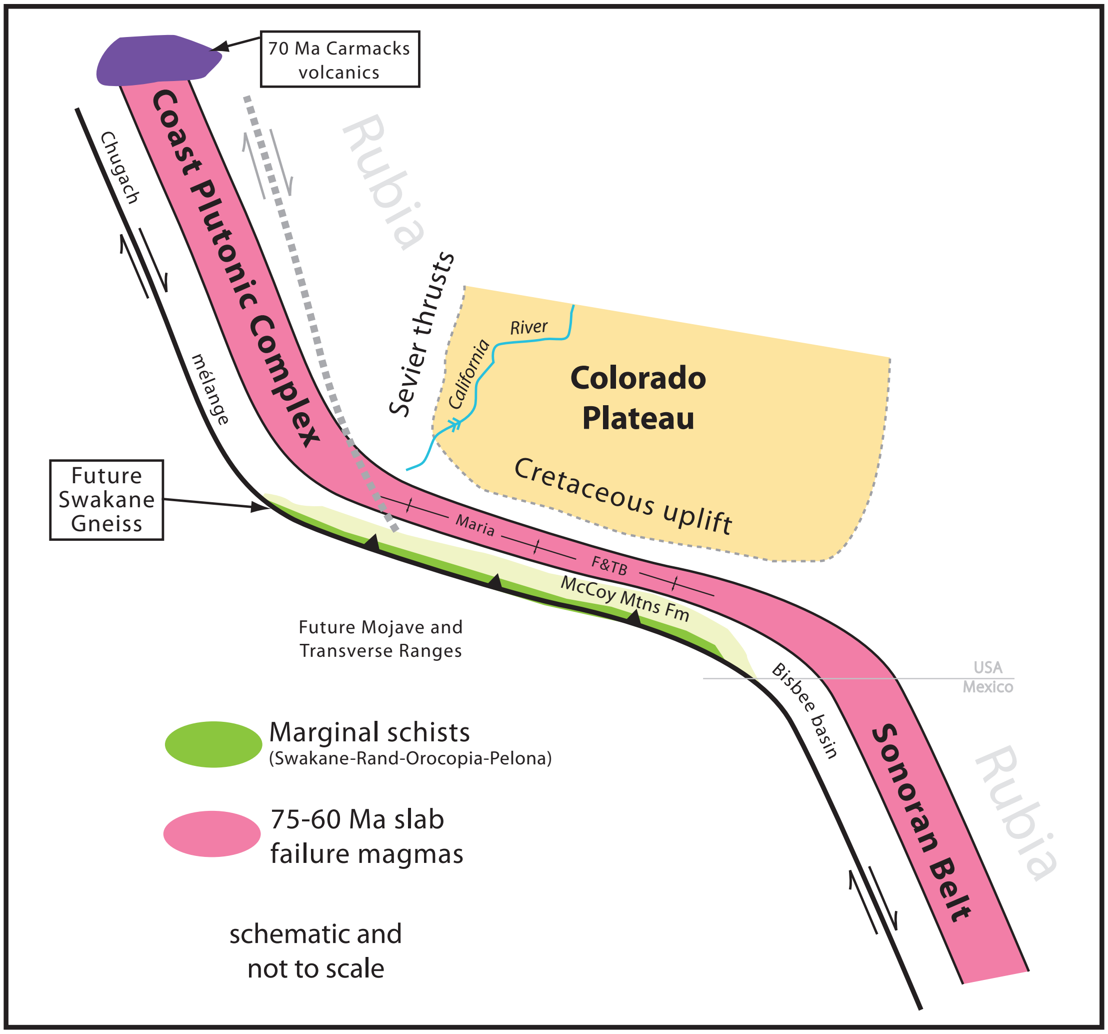


Figure 57. Tectonic model illustrating the original proximity of the Coast plutonic belt to the Sonoran belt and the sigmoidal nature of the plate margin within the southwestern United States at ~70 Ma. The arrival of Rubia caused shutdown of subduction along the margin and ultimately to slab failure leading to a continuous band of magmatism along the margin from the Coast plutonic complex into the Sonoran region. The Swakane gneiss, now outcropping in the High Cascades of Washington, is interpreted as part of the more or less E-W–trending band of marginal schists (Pelona-Orocopia) and thus, along with postcollisional slab-failure magmatism, constitutes a piercing point to restore younger latitudinal separation.

ning of Laramide deformation; so I relate it to the collision of the Laramide ribbon continent. The Chugach accretionary complex, now located in southern Alaska but formerly located much farther south as described earlier, also wasn't active during the period, 84–68 Ma, and so lends additional credence to the idea of a widespread subduction shutdown following Laramide collision. We now explore some of the implications of this shutdown and examine its possible cause.

Exhumation of Franciscan Complex

The contact between the Franciscan complex and the Coast Range ophiolite has long been recognized as a fault and originally considered to be a relict subduction zone (Bailey et al., 1970; Ernst, 1970). More recently, workers argued that because there is a substantial difference in metamorphic grade between the relatively unmetamorphosed rocks of the Great Valley Group and the blueschist-grade Franciscan rocks, there is a thick crustal section missing, so that the fault must be extensional and have a normal sense of movement (Suppe, 1973; Jayko and Blake, 1986; Platt, 1986; Jayko et al., 1987; Harms et al., 1992). Similarly, there is a 3 kb pressure gap across the fault between the Central and Coastal belts (Terabayashi et al., 1992) so that there is, in a way, a sandwich structure, where higher grade rocks are "sandwiched" in between rocks of lower grade (Suppe, 1973). This is not a new, or even a local problem,

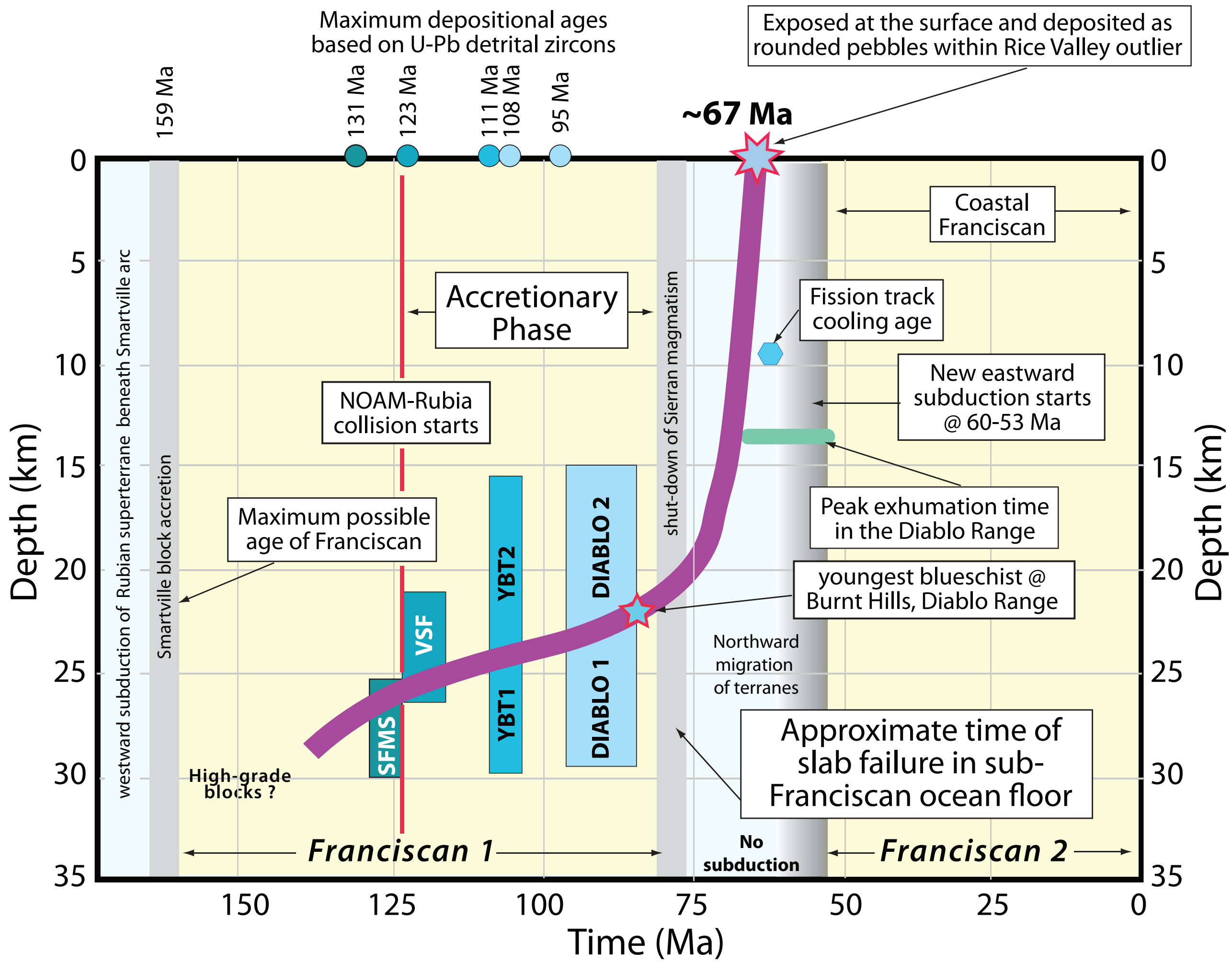


Figure 58. Depth-time plot for rocks of the Franciscan complex illustrating its development. I infer two distinct periods of development for the Franciscan complex, with a period of nonsubduction between them. On this plot one can readily see the rapid burial of packages of 131–95 Ma sediments, shutdown of Franciscan subduction by ~80 Ma, and rapid exhumation of blueschists, at least some of which must have been at the surface by ~67 Ma (Berkland, 1973). New subduction started sometime between 60 and 53 Ma and is represented by the Coastal Franciscan. Detrital zircon ages (small colored circles) are from Dumitru et al. (2010), except the 108 Ma age, which is from Unruh et al. (2007). The colored boxes, which represent estimated depth and metamorphic ages, were modified from Dumitru (1989), and match the colors of the detrital zircon ages for the same belts: SFMS—South Fork Mountain Schist (McDowell et al., 1984; Jayko et al., 1986); VSF—Valentine Springs Formation (Jayko et al., 1986); YBT1—jadeitic pyroxene-bearing zones of Yolla Bolly terrane; YBT2—lawsonite-aragonite-bearing areas of YBT (Suppe, 1973; Jayko et al., 1986); DIABLO1—jadeitic pyroxene-bearing zones of Diablo Range; DIABLO2—lawsonite-aragonite-bearing areas of Diablo Range (Suppe and Armstrong, 1972; Moore and Liou, 1979; Cloos, 1983). Fission track cooling age and peak exhumation time in the Diablo Range are from Unruh et al. (2007). Note that the youngest known blueschist metamorphic age in the Franciscan is Coniacian (Blake et al., 1985; Wakabayashi and Unruh, 1995; A. Jayko, 2010, personal commun.). Based on a slide by Jayko (2009).

as pressure gaps of 5–15 kb are typical of similar belts worldwide (Maruyama et al., 1996). Nevertheless, for the Franciscan, Platt (1986, 1993) argued that the blueschists were exhumed by normal faulting during a period of synsubduction extension; Ring and Brandon (1994) proposed that the exhumation was generated along an out-of-sequence, westerly-directed thrust; Cloos (1982) advocated a channel-flow model in which there is active flow in a narrow channel; Ring (2008) advocated erosional exhumation of an uplifted fore-arc high; Unruh et al. (2007) championed synsedimentary extension; Krueger and Jones (1989) wanted to generate the uplift by Laramide shallowing of the slab dip; Terabayashi et al. (1996) argued for a wedge extrusion process; and many workers (Cloos and Shreve, 1988a, 1988b; Dumitru, 1989; Jayko et al., 1987) pushed underplating as a mechanism to create the uplift. Each model failed because they attempted to generate the exhumation during ongoing subduction and did not consider the possibility that subduction shut down prior to exhumation.

The recognition that subduction stopped at ~80 Ma allows that slab failure caused by the Laramide collision might provide a mechanism for uplift and exhumation of the coherent blueschist terranes. The mechanics of slab failure and their applications to orogeny and magmatism are reviewed elsewhere (Price and Audley-Charles, 1987; Sacks and Secor, 1990; Davies and von Blanckenburg, 1995; Hildebrand and Bowring, 1999; Davies, 2002; Levin et al., 2002; Haschke et al., 2002, Cloos et al., 2005; Hildebrand, 2009) and therefore won't be repeated here. While it is difficult to absolutely estimate the strength of materials and the positive and negative buoyancy forces generated during break-off, modeling (Davies and von Blanckenburg, 1995) clearly shows that in the case of collision, or even slab stagnation, the dense deeper part of the slab will tear off, allowing the shallower, more buoyant upper part, whether it be of oceanic or continental material, to rise. In the case of oceanic material, the uplift of the relict slab might be driven largely by the upward flux of hot asthenosphere. Additionally, the failure and consequent exhumation and rise of material might produce a major normal fault that corresponds to the surface of exhumed material as shown experimentally by Chemenda et al. (1996). The idea of slab failure leading to exhumation of high-pressure accretionary wedge rocks is not new, as some time ago Maruyama et al. (1996) suggested a two-stage model of wedge extrusion caused by slab delamination, followed by domal uplift.

Overall, uplift caused by shutdown-induced slab failure appears to be the most viable mechanism for explaining the observed relations within the Franciscan accretionary complex, not only because it explains the rapid exhumation of the coherent blueschists, but also their relations with the Coast Range ophiolite (Fig. 59). This concept fits well with the model of Hildebrand (2009), who presented evidence that following the Laramide orogeny, there was no subduction along the western margin of the amalgamated collision zone until a new subduction zone formed at ~53 Ma and produced arc magmatism from Montana to the Yukon and from Arizona southward through western Mexico.

Northward Migration of Terranes

The idea that much of British Columbia, Alaska, and northern Washington migrated northward great distances is not new, for it has been around for decades (Beck and Noson, 1972; Irving et al., 1980) and was labeled the Baja-BC (British Columbia) hypothesis (Irving, 1985) because many rocks in the Canadian Cordillera had anomalously shallow paleomagnetic poles relative to similar age rocks of North America, implying that a major portion of the British Columbian Cordillera migrated northward some 2–3000 km sometime between ~90 and 60 Ma (Irving

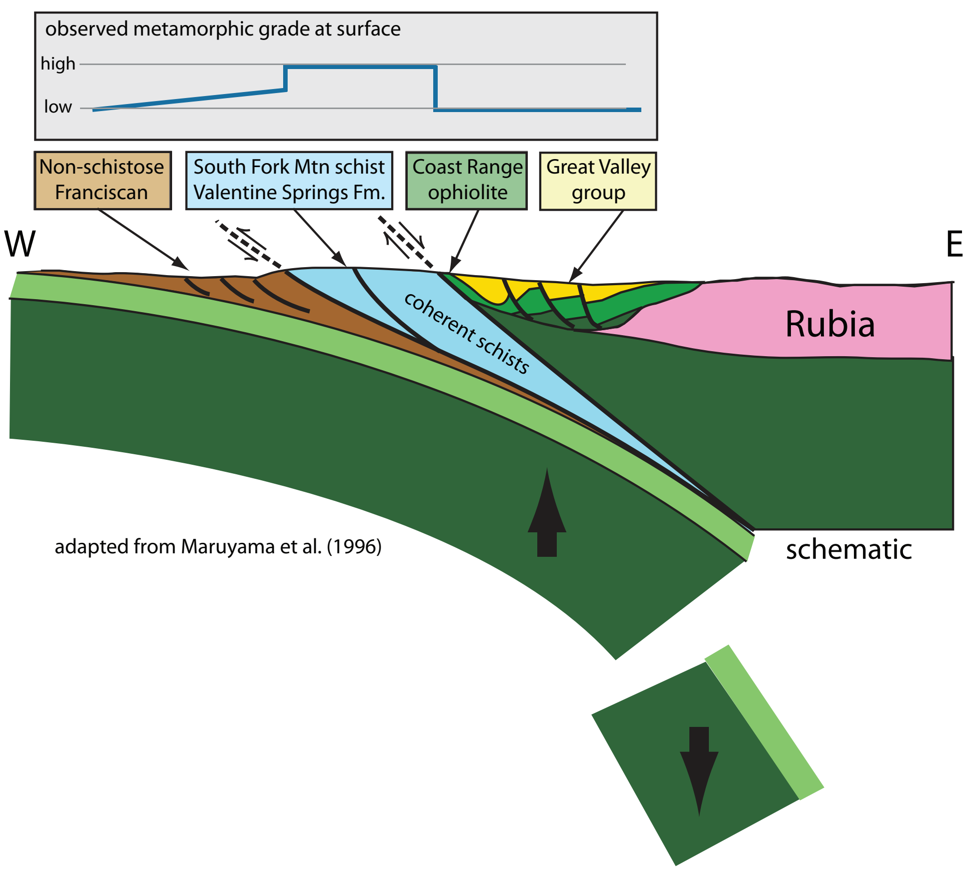


Figure 59. Tectonic model for exhumation of coherent blueschists within the Franciscan subduction complex. The cessation of subduction due to collision of the ribbon continent, consequent failure of the subducting slab, and buoyant uplift of the remaining oceanic lithosphere and its overlying accretionary–fore-arc complex is the simplest and most actualistic method to get the coherent blueschist belts to the surface by 67 Ma. Adapted from Maruyama et al. (1996).

et al., 1995; Wynne et al., 1995; Kent and Irving, 2010). It is now recognized that fragments of continental crust can move rapidly, say 1000 km, within 10–15 Myr (Umhoefer, 2011).

A new tectonic regime started in the Cordillera at ~80–75 Ma, when (1) Laramide thick-skinned deformation commenced in the foreland; (2) Cordilleran-type batholithic magmatism shut down; (3) high-grade metamorphism within the Franciscan complex stopped; (4) deep water sedimentation of the Great Valley Group shoaled; and (5) slab-failure magmatism started in both the Coast plutonic complex and the Sonoran block. These more or less coeval events suggest that subduction on both sides of Rubia shut down at about the same time.

In one example where opposed subduction occurs today, the Philippine archipelago, the interaction of oppositely subducting plates, linked by a transform boundary, is inadequately imaged and so poorly understood. Another region with opposed subduction occurs in the Solomon Islands of the SW Pacific. There the older westward-dipping Pacific slab with the thick Ontong-Java Plateau is imaged dipping beneath the much younger Woodlark slab (Mann and Taira, 2004). Exactly how the two slabs interact at depth is unresolved, largely because the east-dipping slab is very young, but one can imagine several scenarios: (1) the two slabs simply meet and descend steeply into the mantle; (2) the younger, and presumably weaker, slab gets bent back on itself and is carried down with the older slab; (3) rollback of both slabs such that they don't stay in contact for very long; (4) the leading edge of the Australian craton enters the trench and pushes the Pacific plate eastward; and (5) one slab fails and then it impinges on the other plate to delaminate the crust from the mantle so that its slab fails. One can readily conceive of scenarios where huge amounts of water are freed up from either the outer margin of the Australian margin and/or the Ontong-Java Plateau as well as variations in which slabs tear and fall away.

Another well-studied example occurs in New Zealand where opposed subduction zones are connected by the Alpine transform fault. There, Liu and Bird (2006) modeled the collision of the Australian and Pacific plates and showed that a wedging model where one plate impinges on the other, perhaps after slab failure, to delaminate it, best fits the topography, uplift rate, surface erosion, and deep seismicity. In both the Philippines and New Zealand, transpressional regimes took over as slabs interacted with one another and failed.

Whatever the precise cause for the North American situation at 80–75 Ma, the shutdown of both easterly and westerly subduction beneath Rubia, coupled with strong northwestward movement of the oceanic plates within the eastern Pacific basin and southwestward movement of North America (Doubrovine and Tarduno, 2008; Kent and Irving, 2010), appear to have caused parts of the Rubian superterrane to migrate northward relative to North America. In fact, geological constraints on terrane assembly, coupled with paleomagnetic data from a variety of rock types in the terranes, combine to suggest that a substantial portion of Rubia migrated northward starting at ~80 Ma (Fig. 53) and continued until at least ~58 Ma, when shortening stopped in the Rocky Mountain foreland, as marked by significant exhumation of the thrust wedge (Price and Mountjoy, 1970; Ross et al., 2005) and denudation of the giant Belt-Purcell megathrust sheet (Sears, 2001). Late-stage transpressional migration was apparently marked by movement along discrete high-angle faults such as the Tintina (Gabrielse et al., 2006), and continues today along faults such as the Denali (Fuis and Wald, 2003).

An interesting complication is the possible birth of the Kula-Farallon spreading ridge at ~85 Ma (Woods and Davies, 1982; Lonsdale, 1988). Although the location of the ridge at that time is highly speculative as it is now entirely destroyed, it may have been located as far south as present-day Central America (Engebretson et al., 1985). Some workers (Wallace and Engebretson, 1984) suggested that the strong northward migration of the Kula plate led to the generation of the Aleutian trench and the linear belts of Late Cretaceous–early Tertiary magmatism within the Alaska Range–Talkeetna and Kuskokwim Mountains belts. However, as we'll see later in this section, those rocks probably lay farther south and were oriented more northerly than at present.

As stated, the idea of northerly movement of a substantial region of the Canadian Cordillera (Beck and Noson, 1972), generally known as Baja British Columbia, or BajaBC for short (Irving, 1985), has been highly controversial. Brief histories of the controversy are provided by several workers (Umhoefer, 1987, 2000, 2003; Umhoefer and Blakey, 2006; Irving and Wynne, 1992; Cowan et al., 1997) and so needn't be repeated here. The newly calculated North American reference poles of Kent and Irving (2010) provide an up-to-date and robust framework in which to evaluate terrane motion relative to the North American craton. To me the strongest argument for large-scale northward migration of terranes lies in their consistency (Table 1), that is, Upper Cretaceous rocks within the various terranes all have discordant poles, displaced from the North America poles in a consistent manner that indicates dextral shear along the margin after ~80 Ma (Beck, 1991, 1992). Also, many of the modern studies do not use plutonic rocks, but instead have focused on bedded sections with both volcanic and sedimentary rocks, in order to mitigate, or at least reasonably evaluate, the effects of compaction and postdepositional deformation. In what follows, I utilize the most robust paleopoles from the paleomagnetic data and attempt to place them within a coherent and logical framework, largely to test the overall model and evaluate whether or not such large translations make sense in terms of the tectonic development of the Cordilleran orogeny (Table 1).

Data from Stikinia and Wrangellia (Fig. 53) suggest that they were joined, or at least quite close to one another in terms of latitude by the Late Triassic–Early Jurassic in that poles for the 225 Ma Wrangellian Karmutsen Formation (Irving and Yole, 1972, 1987; Schwarz et al., 1980; Yole and Irving, 1980) and the 210 Ma Stikinian Savage Mountain Formation (Monger and Irving, 1980) are quite close to one another as are the 195 Ma Wrangellian Bonanza Formation (Irving and Yole, 1987) and the 195 Ma Stikinian Hazelton Group (Monger and Irving, 1980). Late Pliensbachian ammonites collected on Vancouver Island

TABLE 1. PALEOMAGNETISM OF SELECTED CORDILLERAN UNITS

Unit	Age	Measured paleolatitude	Expected paleolatitude	Northward transport	Source
Silverado Formation	60 ± 2 Ma	25° ± 7° N	37° ± 3° N	12° ± 6°	Morris et al. (1986)
Silverado Formation	62 ± 2 Ma	26° ± 6° N	37° ± 3° N	11° ± 5°	Lund and Bottjer (1991)
Carmacks volcanics	70 ± 1 Ma	54.8° ± 4.1° N	72.1° ± 2.7° N	17.3° ± 5.5°	Enkin et al. (2006a)
Punta Baja Formation	70 ± 3 Ma	29° ± 13° N	34° ± 4° N	5° ± 11°	Filmer and Kirschvink (1989)
Pigeon Point Formation	~71 ± 7 Ma	21° ± 5° N	47 ± 2° N	24° ± 5°	Champion et al. (1984)
Point Loma Formation (N)	72 ± 2 Ma	22° ± 4° N	37° ± 5° N	14° ± 5°	Bannon et al. (1989)
Point Loma Formation (R)	72 ± 2 Ma	20° ± 12° N	37° ± 5° N	17° ± 10°	Bannon et al. (1989)
Rosario Formation (P Baja)	74 ± 6 Ma	26° ± 7° N	34° ± 5° N	8° ± 7°	Flynn et al. (1989)
Nanaimo Group	75 ± 8 Ma	35.7° ± 2.6° N	60.7° ± 3° N	25° ± 3.7°	Enkin et al. (2001)
Rosario Formation (P San Jose)	77 ± 3 Ma	25° ± 2° N	36° ± 5° N	11° ± 5°	Filmer and Kirschvink (1989)
MacColl Ridge Formation	80 Ma	53° ± 8° N	68° ± 6° N	15° ± 8°	Stamatakos et al. (2001)
Ladd and Williams formations	82 ± 8 Ma	27° ± 5° N	38° ± 5° N	11° ± 6°	Morris et al. (1986)
Valle Formation 1 (Vizcaino)	85 ± 1 Ma	22° ± 8° N	36° ± 4° N	13° ± 8°	Patterson (1984)
Valle Formation 2 (Vizcaino)	87 ± 1 Ma	20° ± 5° N	36° ± 4° N	16° ± 5°	Patterson (1984)
Valle Formation 3 (Vizcaino)	90 ± 2 Ma	25° ± 4° N	36° ± 4° N	11° ± 4°	Patterson (1984)
Valle Formation 4 (Cedros Island)	90 ± 2 Ma	22° ± 5°	37° ± 4° N	15° ± 5°	Patterson (1984)
Valle Formation 5 (Vizcaino)	90 ± 2 Ma	25° ± 2° N	35° ± 4° N	9° ± 4°	Patterson (1984)
Valle Formation 6 (Vizcaino)	90 ± 2 Ma	32° ± 6° N	36° ± 4° N	4° ± 6°	Patterson (1984)
Silverquick–Powell Creek Formations	90 ± 5 Ma	39.5° ± 2.2° N	59.8° ± 3° N	20.3° ± 2.7°	Enkin et al. (2006b)
Blue Mountains terranes	~93 Ma	39.2° ± 4.5°	55.1° ± 2.7° N	15.9° ± 4.1°	Housen and Dorsey (2005)
Valle Formation 7 (Vizcaino)	94 ± 2 Ma	20° ± 1° N	35° ± 4° N	14° ± 4°	Patterson (1984)
Valle Formation 8 (Vizcaino)	94 ± 8 Ma	24° ± 12° N	36° ± 4° N	12° ± 10°	Hagstrum et al. (1985)
Valle Formation 9 (Cedros)	95 ± 5 Ma	21° ± 3° N	37° ± 4° N	16° ± 4°	Smith and Busby-Spera (1993)

Key: P—peninsula; N—normal polarity; R—reverse polarity.

have Tethyan affinities and paleontological latitudes (Smith et al., 2001) that agree with the paleomagnetic data set. Kent and Irving (2010) pointed out that (1) from 225 Ma to 90 Ma, the combined Wrangellia-Stikine block moved ~20° northward, whereas the North American craton moved some 35° northward so that there was a net southward movement of the Wrangellian-Stikine block relative to North America; and (2) between 90 and 50 Ma, the Wrangellian-Stikine block moved 20° northward, while the North American craton had been migrating southward between ~145 and 90 Ma and an additional 5° from 90 to 50 Ma, so that the total relative movement was 25° in a dextral sense. These movements account for the early sinistral shear followed by dextral shear.

Cache Creek terrane, located just to the east of Stikinia, contains Tethyan fauna and DUPAL-anomaly basalts (Monger and Ross, 1971; Johnston and Borel, 2007; Johnston, 2008) and Tethyan fauna also exist in Late Triassic–Early Jurassic rocks of Vancouver Island (Tozer, 1982; Smith et al., 2001); so in addition to late northward displacement with the Rubian superterrane, those terranes migrated a considerable distance eastward from equatorial Tethys prior to incorporation within the Rubian superterrane.

The 91 Ma Mount Stuart batholith was analyzed (Beck and Noson, 1972), reanalyzed (Beck et al., 1981), analyzed again (Ague and Brandon, 1996), then studied in more detail (Housen et al., 2003) and found to have migrated northward some 24.5° ± 6.3° relative to North America. Similarly, after recognizing that there were structural complexities, the 109 Ma Duke Island layered ultramafic complex was re-analyzed and found to have been some 2350 km (21°) anomalous with respect to cratonic North America (Bogue and Grommé, 2004). Rocks of the 75 Ma Nanaimo Group on Vancouver Island were found by Enkin to have moved ~2750 ± 400 km northward (Enkin et al., 2001), and leaf margin studies from the Albian Winthrop Formation, located to the east within the Methow basin (Fig. 8), showed some 2200 km of northward displacement (Miller et al., 2006).

In more easterly terranes, Rees et al. (1985) found 94 Ma paleopoles in Quesnellia to be similar to the more westerly terranes in that they are 23° ± 10° anomalous to North America. And detailed paleomagnetic studies over 500 km of strike length within the foreland belt of the Canadian Cordillera indicate that Paleozoic carbonates within the Front Ranges have a steep Late Cretaceous remagnetization, with poles compatible with those from the North American craton (Enkin et al., 2000). If shallow inclinations measured in remagnetized carbonate rocks of the western Main Ranges also formed during the Late Cretaceous, then the differing inclinations between the two belts suggest that the suture between Rubia and North America is at or just west of the Kicking Horse Rim in the Main Ranges and that previous correlations across the carbonate-shale facies change are in error (Enkin, 2006).

As discussed earlier, rocks included in the 95–85 Ma Silverquick and Powell Creek formations are located in southern British Columbia, where they represent overlap successions on the Coast plutonic complex and the Intermontane superterrane; and—based on robust paleomagnetic results from both sedimentary and volcanic rocks—migrated some 2300 km northward (Wynne et al., 1995; Krijgsman and Tauxe, 2006; Enkin et al., 2006b; Enkin, 2006) since they accumulated. This is wholly consistent with results from 70 Ma volcanic rocks of the Carmacks Group, located at the north end of the Coast plutonic belt in the Yukon Territory at 62°N (Figs. 5 and 22), which were located in the western United States at the latitude of present-day Oregon when they were erupted at 70 Ma (Enkin et al., 2006a; Wynne et al., 1998). At that time, North America was located a bit farther north than at present such that San Francisco would have been located at ~45° (Kent and Irving, 2010). As the complex is ~1500 km long, stretching from the Yukon Territory to the High Cascades of Washington, its southern end would have been located more or less at the present-day latitude of southern California–northern Mexico. Presently, at its southern end in Washington is a peculiar body of rock known as the Swakane gneiss, which is similar to nearly every known attribute, including age, composition, rapid, deep burial and exhumation, and in tectonic setting to the Pelona and Orocopia schists of southern California and Arizona (Matzel et al., 2004). By restoring the Coast plutonic complex to its 70 Ma latitude, the Swakane gneiss is approximately restored with the Pelona and Orocopia schists into a single band (Fig. 56). Similar results were obtained from bedded rocks of the 78 Ma MacColl Ridge Formation in Alaska, which are part of the Wrangellian composite terrane presently located at 61°N, in that they show a paleolatitude 15° ± 8° lower than at present (Stamatakos et al., 2001) putting them in the area of the restored San Francisco region, consistent with the Carmacks results.

95-85 Ma Powell Creek and Silverquick fms sit on both Insular and Intermontane superterranes

26 sites in volcanic lava flows and **54 sedimentary sites**, along with several **positive contact, conglomerate, and tilt tests** combine to provide a reliable and robust record of the geomagnetic field inclination for the area and yield a paleolatitude of 39.5°±2.2°, which is about **20.3°±2.7° south of the expected paleolatitude of North America at that time** (Enkin et al., 2003; Enkin et al., 2006b; Kent and Irving, 2010)

Previous workers (Pope and Sears, 1997; Wernicke and Klepacki, 1988) developed models in which the Cassiar platform and Stikinia of the Canadian Cordillera (Fig. 5) escaped northward from the area of Idaho-Montana, but the linkages between terranes and the paleomagnetic data cited above indicate that terranes of the Canadian Cordillera were already assembled by 70 Ma and located much farther south. In fact, as stated earlier, the Carmacks volcanics, located today in the Yukon Territory at the north end of the Coast plutonic complex, were located near San Francisco, so it seems more likely that the terranes of the Canadian Cordillera were originally located at the south end of the Great Basin sector, not the north. Restoring the Cordillera southward also joins the 80–75 Ma slab-failure plutonic rocks of the Sonora-Mojave region with those of the Coast plutonic complex (Fig. 56) as discussed earlier (see Hildebrand, 2009).

Except for the Sierra Nevada, other terranes of the western United States show displacements that are consistent with the movement of the Canadian terranes. Rocks from two sites within the Salinian block along the central California coast, Point San Pedro (Fig. 28), where Paleocene sedimentary rocks sit in buttress unconformity with a Cretaceous granodiorite, and at Pigeon Point (Fig. 28), where thick sections of Campanian–Maastrichtian turbidites are exposed, were sampled for paleomagnetic analysis (Champion et al., 1984). Their results indicated that the rocks of both formations originated some 2500 km farther south than their present latitude at 25°N and 21°N, respectively.

The Sierra Nevada present an unresolved problem in paleogeographical reconstructions because existing paleomagnetic data indicate little movement (~1000 km) relative to North America since ~100 Ma (Frei et al., 1984; Frei, 1986). Samples from plutons in the 102 to 97 Ma range were sampled, and as we know that 100 Ma volcanic rocks were folded, it is hard to understand how the plutons might have escaped the folding. Thus, the magnetization measured might be secondary and postdate folding. Irrespective of the explanation, the Sierra Nevada needs additional paleomagnetic study. It is tempting to link the Sierra Nevada batholith with the Omineca belt batholith (Fig. 56) as neither contain the Late Cretaceous–early Tertiary plutons characteristic of both the Coast plutonic complex and the Sonora-Mojave region.

The White-Inyo Mountains block, located just to the east, is an isolated block, containing a 7-km-thick section of early Paleozoic rocks that continues to the NE into Nevada only in Esmeralda County, and it and the Sierran block may have migrated northward from northern Mexico, which is permissible given the paleomagnetic data from the Sierra and the southward migration of North America (Kent and Irving, 2010) following its emplacement, that collectively yield ~1000 km of relative displacement.

To the south in Baja California, several paleomagnetic studies have concluded that rocks of the Valles fore arc were deposited 10°–20° farther south than their current location (Hagstrum et al., 1985; Smith and Busby-Spera, 1993; Sedlock, 1993; Hagstrum and Sedlock, 1990, 1992). Some fixists (Butler et al., 1989, 1991; Butler and Dickinson, 1995; Dickinson and Butler, 1998) argued that shallowly inclined paleomagnetic results weren't faithful indicators of paleolatitude because those taken from sedimentary successions have compaction shallowing, whereas the plutonic samples were tilted after emplacement and/or affected by much younger remagnetization. Beck (1991) examined the various possibilities and concluded that the consistency of the data is remarkable and supports the far-traveled results. Also, Smith and Busby-Spera (1993) tested the effects of compaction, and analyzed slump blocks in an olistostromal unit to test the remagnetization hypothesis. They concluded that the rocks were not remagnetized and that northward displacement of 18° ± 7° best explained the data. More recently, Sedlock (2003) provided an excellent overview of the pros and cons of the paleomagnetic possibilities and concluded that northward migration of the western terranes, including the Valles Formation, was most likely.

Located along the west coast of Mexico, both north and south of Acapulco, is an apparent post–80 Ma truncated margin where Precambrian–Mesozoic metamorphic and Cretaceous plutonic rocks of the Xolapa terrane (Fig. 5) lie abnormally close to the Middle America Trench, and so might represent the original home of one of the slices found much farther north today (Karig et al., 1978). Based on detrital zircon populations, Wright and Wyld (2007) argued that rocks of the western Great Valley group originally were deposited near Oaxaca, Mexico, which might be a reasonable fit.

Jumping northward, sparse paleomagnetic data from rocks in northern Alaska suggest that many of those terranes were located in western Canada before migrating northward to form the Alaskan orocline (Johnston, 2001). Data cited earlier clearly show that more outboard terranes, such as Chugach and Prince William, were originally located along the western side of the Coast plutonic complex, and migrated northward (Farmer et al., 1993; Sample and Reid, 2003; Roeske et al., 2003, 2009; Housen et al., 2008). Rocks of the Arctic Alaska, Yukon-Tanana, and Selwyn basin terranes share similarities, such as detrital zircon suites (Beranek et al., 2010a, 2010b), with each other, and with rocks in western Nevada, such as the Roberts Mountain allochthon, in that they were deformed and metamorphosed in the Late Devonian–Early Mississippian, characteristics unknown from the western part of cratonic North America. Arctic Alaska was apparently located 12° ± 5° south of its present position at 130 Ma (Halgedahl and Jarrard, 1987); so it is reasonable to conclude that this family of terranes, and those accreted to it, were formerly located at the latitude of western Canada north of the Nevadan region.

By 50 Ma (Fig. 53), it appears as though the northward migration of Cordilleran terranes into the Canadian sector was largely complete as the Eocene Ootsa Lake volcanics of central British Columbia are more or less in place (Vandall and Palmer, 1990), as are the 50 Ma Flores volcanics of Vancouver Island (Irving and Brandon, 1990), and the late Neogene basalts of British Columbia (Mejia et al., 2002). After 50 Ma, both North America and its attached terranes moved a small amount southward (Kent and Irving, 2010). Note that the overall movement of these terranes—1000 km in 25–40 Myr—is well within modern estimates of terrane translation (DeMets et al., 2010; Umhoefer, 2011).

A major effect of the northward migration of Rubia was the development of the Rocky Mountain fold-thrust belt (Price and Mountjoy, 1970; Price, 1981) within rocks of the North American platform shelf. Recent work by Larson et al. (2006) demonstrated that lower Paleozoic rocks of the Belt-Purcell-Windermere block located on the east limb of the Purcell anticlinorium, were deformed by thrusting prior to emplacement of intrusions dated by $^{40}Ar/^{39}Ar$ at 108 Ma; whereas passive margin rocks immediately to the east weren't involved in thrusting until the Campanian, coincident with deposition of the thickest sedimentary wedge in the foredeep and containing metamorphic and igneous debris (Leckie and Smith, 1992). Larson et al. (2006) suggested that the fold-thrust belt might represent dextral transpression along the North American margin. Thus, within the Canadian sector, the suture between the Rubian superterrane and the North American margin is a relatively late feature that lies just west of the North American shelf edge and is represented by different thrusts depending on location. On both sides of the international border, the suture is the thrust fault that carries the Belt-Purcell supergroup—in places, the Lewis-Eldorado-Hoadley—out over the Cretaceous sedimentary rocks of the foredeep. In many places along the suture, there are half-gräben (Price, in press; Constenius, 1996), which presumably formed due to collapse of thickened crust where North American strata were pulled down beneath the leading edge of Rubia.

An interesting corollary of the transpressional model is that if time-temperature curves for the Monashee complex (Sevigny et al., 1990; Parrish, 1995) are correct, not only were the Rubian rocks transported northward, but locally so were some North American rocks which had earlier (125–108 Ma) been overthrust and metamorphosed by the Rubian terranes. Thus, the basement within the core of the Monashee complex might be a piece of North America severed from the craton much farther south and transported northward with its allochthonous, exotic Rubian cover. It is thus a somewhat isolated piece of crust analogous to the tiny piece of North America isolated north of the Maria fold-thrust belt and south of the Phoenix fault (Fig. 8).

Colorado Plateau and the Laramide Grand Canyon

The southwest corner of the Colorado Plateau appears to have been particularly affected by the Laramide collision and subsequent slab failure in that it was apparently high-standing with northward drainage by the Late Cretaceous–early Tertiary (Flowers et al., 2008; Hill and Ranney, 2008; Wernicke, 2011; Flowers and Farley, 2012). Paleocene–Eocene rim gravels located near the boundary of the Transition zone and the Colorado Plateau indicate NE paleoflow (Elston and Young, 1991; Potochnik, 2001) on a high-relief surface, as locally the gravels fill paleochannels as deep as 1200 m (Young, 1979). The gravels contain clasts of volcanic rock with ages of 80–64 Ma (Elston et al., 1989). The canyons and clast provenance combine to present a compelling case that a proto–Grand Canyon was carved mainly during the Campanian by the California River as it flowed down the surface to the N-NE (Wernicke, 2011). Fluvial strata of the Upper Cretaceous Wahweap and Kaiparowits formations in southern Utah yielded minimum age peaks of detrital zircons at 82, 77, and 73 Ma from bottom to top suggesting transport of progressively younger debris from the S-SW by the NE-flowing California River (Larsen, 2007; Jinnah et al., 2009; Larsen et al., 2010). The age peaks are too young to have been derived from the Cordilleran batholiths but fit well with the slab-failure rocks erupted and intruded to the south. Younger sedimentary rocks of Maastrichtian age have 105–100 Ma detrital zircons indicating that the Cordilleran batholiths or possibly rocks of the Delfonte volcanic field were being eroded (Link et al., 2007b; Larsen et al., 2010). The uplift, yet lack of folding and penetrative deformation of the Colorado Plateau, just to the north of the Maria fold-thrust belt suggest that the two were juxtaposed at a later date, most probably along the Phoenix fault (Hildebrand, 2009) after, or during, the Miocene collapse of the thickened fold belt. I suggest that the Colorado Plateau formed during the Laramide event when Rubia collided with both the western and southern sides of the region. To the north, the plateau is bounded by the Orofino fault and/or the Lewis and Clark fault system.

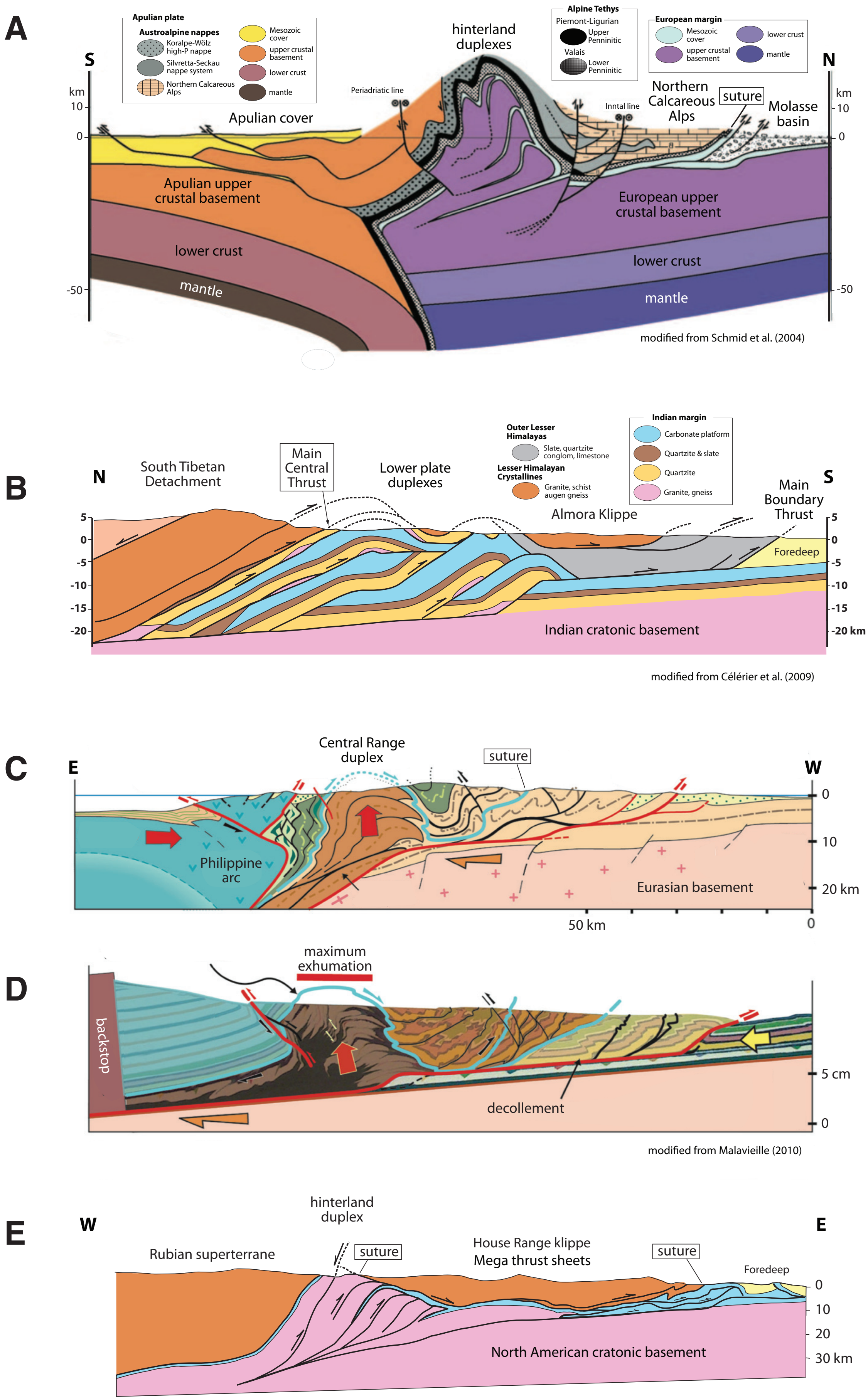
Metamorphic Core Complexes and Hinterland Duplexes

In western North America, there appear to be four age groups of metamorphic core complexes, one in each of the four main sectors of the orogen: (1) an Eocene band, located north of the Lewis and Clark lineament in the Canadian sector; (2) a dominantly Paleocene band located in the Great Basin sector from the Lewis and Clark lineament to Death Valley; (3) a Miocene band in the Sonoran sector that arcs eastward from the Mojave Desert southward into Sonora, Mexico; and (4) an Early Cretaceous group mostly located along the southern side of the Brooks Range in Alaska. This general scheme, outside of the Alaskan occurrences, was recognized by Coney and Harms (1984), but insufficient age dates were available to them at that time to define the temporal relations.

The Miocene complexes are best known in the Colorado River–southern Arizona region (Davis, 1980; Davis et al., 1980; Rehrig and Reynolds, 1980; Anderson, 1988; Dokka, 1989; Spencer and Reynolds, 1990; Foster and John, 1999). They continue southward in Sonora (Anderson et al., 1980; Nourse et al., 1994) and westward into the central Mojave (Dokka, 1989; Glazner et al., 1989, 2002; Walker et al., 1990a). Because the distribution of the core complexes so closely follows the zone of older Laramide deformational and related slab-failure magmatism, I relate the core complexes stretching from the Mojave eastward into Arizona and south into Mexico to collapse of this part of the Laramide belt during the Miocene as did Coney (1987). Walker et al. (1990a) related the extension within the Mojave region to a pulse of Miocene magmatism but didn't consider the total distribution of the complexes.

Hildebrand (2009) related the Paleocene–Eocene collapse of the hinterland belt within the Great Basin and Canadian sectors to slab failure and consequent uplift of North America beneath Rubia, but this may not be entirely correct, for the hinterland belt has many features in common with hinterlands of other collisional orogens, in some of which the slab hadn't yet failed when collapse initiated (Mattauer et al., 1983). The most obvious and pertinent features are the lower-plate antiformal duplexes that form in collisional orogens due to erosional exhumation and underplating (Malavieille, 2010). These duplex structures typically isolate a klippe of exotic rocks on the foreland side of the orogen (Fig. 60). In the Alps, a large crustal duplex that contains the Tauern window—where European basement is exposed beneath the suture—has isolated the Northern Calcareous Alps, which comprise low-grade sedimentary rocks of Apulian provenance that sit structurally upon European rocks, out in front of the duplexed core (Schmid et al., 2004). Similarly, recent work in Taiwan (Beyssac et al., 2007) suggested that much of the deformation and exhumation in the higher grade Central Range is sustained by underplating rather than solely by frontal accretion in a critical wedge. And in the Lesser Himalaya a similar duplex structure, also formed by underplating, isolated rocks above the Main Central thrust as the Almora klippe from its main mass to the north (Célérier et al., 2009; Bollinger et al., 2004). In the orogen of Oman, the Saih Hatah Mountains are underlain by a thick-rooted, duplex anticline that isolates a huge klippe of the Semail ophiolite to its foreland side (Hanna, 1990; Al-Lazki et al., 2002; Gray and Gregory, 2003; Searle et al., 2004; Searle, 2007). Within the North American Cordillera, the presence of probable North American crystalline basement within relict antiformal structures of the hinterland, such as occur within the Ruby, East Humboldt, Raft River–Albion, and Pioneer Mountains, the Priest River complex of Washington, and the Monashee complex of British Columbia (Howard et al., 1979; Journeay, 1992; Parrish, 1995; Doughty et al., 1998; Snoke and Miller, 1988; Link et al., 2007a; Gervais et al., 2010), suggest that the hinterland belt originally formed as an underplated duplex region and that an isolated klippe of exotic rocks sits to the east today on rocks of the North American platform. I informally refer to the klippe as the House Range klippe after exposures in the House Range of

Figure 60. Comparative cross sections showing typical exotic klippen sitting to the foreland side of well-developed lower-plate duplex structures developed in the hinterland from erosional exhumation in a deforming wedge developed above a basal detachment: (A) cross section through the eastern Alps (Schmid et al., 2004); (B) cross section through Himalayas after Célérier et al. (2009); (C) cross section through the Taiwan arc-continent collision zone (Malavieille, 2010); (D) cross section of a scaled model (Malavieille, 2010); and (E) a pre–Basin and Range model through the Great Basin showing the isolated klippe of Rubian rocks lying to the east of the hinterland belt. Other examples include the Saih Hatat of the Oman Mountains, where the Semail ophiolite forms an isolated klippe (Searle et al., 2004); the Brooks range of Alaska, where the Doonerak Fenster sits atop the duplex stack and exposes lower-plate rocks (Moore et al., 1994); and Wopmay orogen, where a large klippe of exotic Hottah terrane sits isolated to the east of a lower-plate duplex formed from slabs of the Archean Slave craton (Hildebrand et al., 2010a).


A
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Apulian plate
Austroalpine nappes
Koralpe-Wölz high-P nappe
Silvretta-Seckau nappe system
Northern Calcareous Alps
Mesozoic cover
upper crustal basement
lower crust
mantle
hinterland duplexes
Alpine Tethys
Piemont-Ligurian
Upper Penninitic
Valais
Lower Penninitic
European margin
Mesozoic cover
upper crustal basement
lower crust
mantle
Periadriatic line
Inntal line
Northern Calcareous Alps
suture
Molasse basin
Apulian cover
Apulian upper crustal basement
lower crust
mantle
European upper crustal basement
lower crust
mantle
modified from Schmid et al. (2004)
B
N
S
South Tibetan Detachment
Main Central Thrust
Lower plate duplexes
Outer Lesser Himalayas
Slate, quartzite conglom, limestone
Lesser Himalayan Crystallines
Granite, schist augen gneiss
Indian margin
Carbonate platform
Quartzite & slate
Quartzite
Granite, gneiss
Main Boundary Thrust
Almora Klippe
Foredeep
Indian cratonic basement
5
0
-5
-10
-15
-20
-20 km
modified from Célérier et al. (2009)
C
E
W
Central Range duplex
suture
Philippine arc
Eurasian basement
0
10
20 km
50 km
0
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maximum exhumation
backstop
decollement
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modified from Malavieille (2010)
E
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hinterland duplex
Rubian superterrane
suture
House Range klippe
Mega thrust sheets
suture
Foredeep
North American cratonic basement
0
10
20
30 km

western Utah. A similar hinterland welt occurs within the Brooks Range of Alaska, where the Doonerak fenster (Fig. 5) sits along the crest of the Mount Doonerak antiform (Moore et al., 1997; Fuis et al., 2008) and was interpreted as a window into duplexed basement (Oldow et al., 1987, 1989).

The intense erosion and exhumation combine to provide an ideal place for both syn- and postcompressional extensional collapse. Thus, these regions can locally preserve both thrusts and normal faults with complex interplay between the two. For example, Cretaceous normal faults are known to have developed along the south side of the Brooks Range in the so-called Schist belt coincident with uplift (Gottschalk, 1990; Gottschalk and Oldow, 1988; Gottschalk et al., 1998; Miller and Hudson, 1991; Law et al., 1994; Little et al., 1994). Argon mica ages increase in age southward from ~90 Ma to ~100 Ma (Vogl et al., 2002), whereas data collected by Toro et al. (2002) indicate that peak extension took place before 112 Ma with rapid cooling in the range of 98 to 90 Ma. Overall, the presence of erosional duplexes containing both thrust and normal faults that isolate klippen of exotic rocks to the foreland side of the orogen appear to be common, and perhaps diagnostic, features of collisional orogens.

Basin and Range Extension

I earlier suggested (Hildebrand, 2009) that the Basin and Range extensional province developed in areas where the Rubian superterrane sat atop the North American craton. Within the Great Basin a set of Consortium for Continental Reflection Profiling (COCORP) deep seismic lines from western Utah to eastern California showed strong horizontal reflectors in the deep crust westward from central Nevada to eastern California (Allmendinger et al., 1987) that may represent the suture zone and possibly a thin veneer of Paleozoic siliciclastic metasedimentary rocks of the autochthon beneath Rubia. As much of the Great Basin region has crust of "normal" 30–35 km thickness (Heimgartner et al., 2006) and has undergone more or less 100% extension (Gans and Miller, 1983; Wernicke, 1992), it follows that prior to normal faulting the crust was approximately double normal thickness. Because the distribution of the overthrust region coincides with the extended and collapsed region, I postulate that the early to mid-Tertiary extension resulted directly from the collapse of the region with doubled Rubian–North American crust formed during the collisional event. While topography could be the main control on extension, it may be that following collision the hot, and possibly molten, lower crust of the upper plate was more likely to have flowed laterally when sandwiched between cool lower-plate crust and its own cool upper levels (Wernicke, 1992; Burov and Watts, 2006).

An excellent analogue is the Tibetan Plateau, which is a region of double-thickness crust formed as a result of convergence between India and Eurasia (Molnar and Tapponnier, 1975). At least part, and perhaps all, of the area of thickened crust can be directly related to subduction of Indian lithosphere beneath Eurasia (Searle et al., 1987). A variety of geophysical data, such as seismic reflectors, highly conductive and low-velocity zones, high heat flow, and strong attenuation of seismic waves, are collectively interpreted to indicate that a partially molten zone exists at depths of 15–20 km beneath the plateau (Nelson et al., 1996; Schilling and Partzsch, 2001).

Driven by the gravitational potential of the thick crust in high plateaus, the thickened region could have thus flowed outward along the partially molten layer, causing the sheetlike region above it to extend (England and Houseman, 1988; Teyssier et al., 2005). For North America, I envision that a rheological gradient existed where the lower part of the Rubian plate was able to flow laterally above the suture zone, whereas the uppermost crust simply broke up in brittle fashion.

The Mexican Basin and Range province extends southward from the Phoenix fault through Mexico to at least the Trans-Mexican volcanic belt (Stewart, 1978; Henry and Aranda-Gomez, 1992; Henry et al., 1991). While deep seismic data are not available for the Mexican Basin and Range province, it occupies the same tectonic setting as that of the Great Basin in that it lies immediately west of the fold-thrust belt of eastern Mexico, and so it seems reasonable to assume that lower-plate crust continues well to the west in the subsurface.

Basin and Range type extension didn't occur within the Canadian sector. The reason for this is probably because the Rubian superterrane is not sitting atop nearly as much North American crust as it is farther south. The Rocky Mountain fold belt was created by transpression as the Rubian terranes migrated northward, so barely sits on North America.

SUMMARY OF CORDILLERAN ASSEMBLY

One of the major results of this study is that the Rubian ribbon continent grew by accretion on both its eastern and western margins and was nearly fully developed by the time it impinged on the western passive margin of North America—first in the Great Basin sector during the Sevier event, and more completely during the Laramide event. North America did not grow westward by incremental accretion (Fig. 39).

Rocks of the Roberts Mountain allochthon were emplaced upon the Rubian margin during the Late Devonian–Early Mississippian Antler orogeny, and coarse debris was shed eastward to form a clastic wedge over the pre-collisional Antler shelf. Upper Devonian to earliest Triassic chert-argillite sequences with intercalated lenses of pillow basalt of the Golconda allochthon were emplaced over the modified western margin of the Roberts Mountain allochthon during the Early Triassic Sonoman orogeny. If there was an arc located on the western part of the Roberts Mountain allochthon, it was removed by faulting prior to the arrival of the Golconda allochthon. It could be located within the Canadian sector as the Kootenay terrane.

Within the Canadian sector, Wrangellia and Alexander terranes were stitched together by a 309 ± 5 Ma pluton, and those terranes developed together during the late Paleozoic and Mesozoic. Along the eastern margin of Rubia, subduction was continuously westward-dipping as a series of arc-bearing blocks and their accretionary complexes were added to the superterrane along west-dipping sutures. Between 260 and 253 Ma, the western margin of the Cassiar platform–Selwyn basin was pulled down beneath Yukon-Tanana terrane on the Inconnu thrust and rocks of the oceanic Slide Mountain terrane were telescoped and sit structurally between the two terranes.

Between 187 and 173 Ma, the western edge of Kootenay terrane was pulled beneath Quesnellia to form an eastward-vergent fold-thrust belt, and soon afterwards early southwest-verging structures such as the Scrip nappe were overprinted by northeast-verging folds and thrusts between ~173 and 168 Ma and were intruded by a swarm of plutons. The second phase of deformation corresponds to the attempted westward subduction of the Belt-Purcell-Windermere block beneath Kootenay terrane at 173 Ma. In Alaska, Wrangellia was pulled beneath the oceanic Talkeetna arc at 170 Ma and generated a northward-vergent fold-thrust belt and foredeep along its northern margin.

The broad temporal relations of the Cordilleran orogeny are shown on Figure 61 with events from the Sierra Nevada westward displayed on Figure 62. In the model favored here, a separate ribbon continent, or composite arc terrane, consisting of small Neoproterozoic–Paleozoic blocks such as the Shoo Fly, Redding, Trinity, and Yreka, along with Permo-Triassic McCloud arc terranes, formed a basement for Late Triassic–Jurassic arc magmatism until it collided with Rubia at 160 Ma (Fig. 40). Until the collision, there were westward-dipping subduction zones on each side of the older Sierran-Klamath block. Along its western side, there was westward subduction beneath the Triassic–Jurassic Slate Creek–Combie-Hayfork arc (Fig. 62) until it collided with the western margin of the western Sierran–Klamath block at 169–164 Ma (Wright and Fahan, 1988; Day and Bickford, 2004). Postcollisional plutons intruded both the Sierra (159–150 Ma) and Klamaths (162–156 Ma) and are attributed here to slab failure during the collision.

To the east, westwardly subduction beneath the Black Rock–Klamath–Sierra Nevada–Mojave-Sonora arc and its dominantly Neoproterozoic–Paleozoic basement led to the development of a Triassic–Jurassic continental magmatic arc extending from northern Nevada southeastward through the Mojave-Sonoran region. This arc collided with the western margin of Rubia and its carbonate-dominated passive margin at ~160 Ma, which created the thin-skinned easterly-vergent Luning-Fencemaker thrust belt and associated thrusts to the southeast (Fig. 40). During the collision, the subducting slab failed and hot asthenosphere was able to upwell through the tear and into the crust where it led to crustal melting and the emplacement of a linear band of magmatism including the Independence dike swarm and the bimodal, alkaline Ko Vaya plutonic suite between ~150 and 145 Ma.

At ~159 Ma, the Smartville arc collided with the western margin of the Sierran block to create another collisional belt (Figs. 40 and 62). The polarity of the subduction that led to this collision was westerly, beneath the Smartville arc, and the western edge of the Sierran block was partially subducted beneath the arc so that the arc now sits above continental crust of the Sierran block.

In Alaska, the Brookian orogeny apparently occurred during the latest Jurassic–Neocomian when the Angayucham ocean closed and the Koyukuk terrane, which is interpreted to represent an arc terrane on the upper plate, collided with Arctic Alaska. Many smaller terranes of central Alaska, such as Farewell, Ruby, and Kilbuck were probably already attached to Arctic Alaska before the Brookian orogeny. There is still some uncertainty as to the exact age of the Brookian orogeny, and it could have started as late as Aptian. Similar-age ophiolites, such as Angayucham, Ingalls, Josephine, and Coast Range, over the length of the orogen suggest they were formed in the same marginal sea between 170 and 160 Ma.

The Franciscan subduction complex developed after 159 Ma, and maybe not until just after 131 Ma, the age of the youngest detrital zircons in the oldest and farthest inboard coherent unit, the South Fork Mountain Schist (Fig. 62). Even the oldest facies of the Great Valley group, the Stoney Creek, contains zircons as young as 135 Ma. Therefore, it is possible that easterly subduction on the west side of Rubia didn't even start until the Sevier collision. Only the exotic blocks suggest an older age for the Franciscan, and they are polycyclic in that they were generally encased in serpentinite prior to incorporation in the mélange. The only magmatism that could be related to eastward-dipping subduction prior to the pulse of Cordilleran batholiths is a small group of 140 Ma plutons in both the Klamaths and western Sierran metamorphic belt, but even those plutons could be slab-failure magmatism related to slightly older collisions as mentioned earlier. Nevertheless, the presence of such large plutons document that the accreted terranes sit atop continental crust.

The subduction complex was nonaccretionary until 123 Ma when the eastern side of Rubia collided with western North America in the Great Basin area (Fig. 63). Rocks of the Great Valley fore-arc basin were disrupted, faulted, and warped. This occurred just after the North American continent with its west-facing passive margin and cratonic terrace rode up and over the outer swell to the west-dipping trench on the eastern, or Panthalassic, side of the Rubian superterrane, and shed sedimentary debris eastward to form an extensive sheet of gravels and pebbly conglomerate. The cratonic terrace was then pulled down into the trench, deformed, and detached from its basement along a basal décollement, while the gravels and conglomerates were buried by orogenic debris of the depressed foredeep.

The major period of thrusting in the Sevier fold-thrust belt of the Great Basin segment occurred from ~123 Ma to ~108–105 Ma and resulted in the accretion of the large Proterozoic–Cambrian clastic megathrust sheets. Farther west, within what would later become the orogenic hinterland, the Rubian ribbon continent was detached from rocks to the east and migrated southward relative to North America (Fig. 63).

A period of slab-failure magmatism started by ~96 Ma and led to the intrusion of linear belts of small, metal-rich alkaline plutons into the Rubian crust (Fig. 63). The plutons now outcrop in

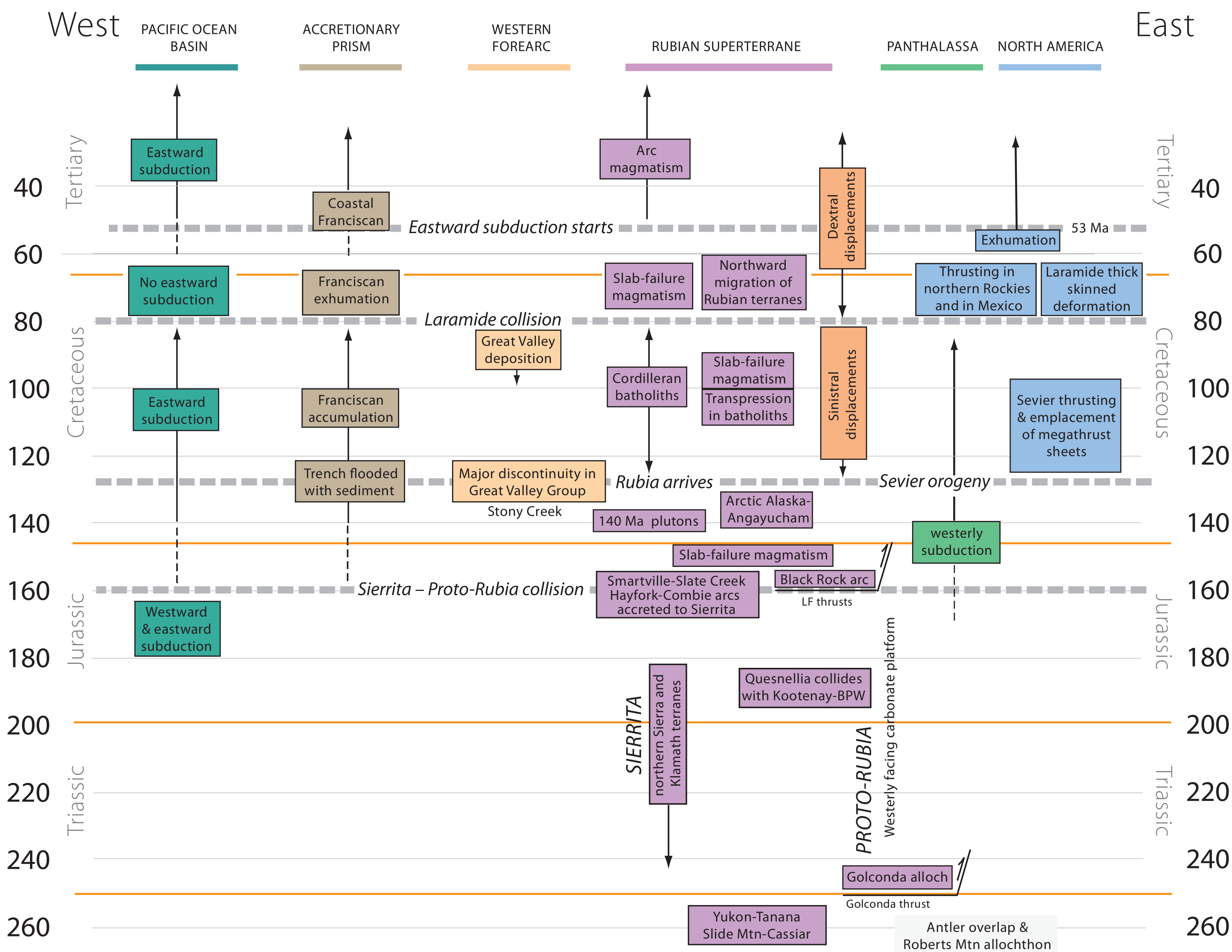

Figure 61. Time-space diagram for various regions within the Cordilleran realm. BPW—Belt-Purcell-Windermere; LF—Luning-Fencemaker.

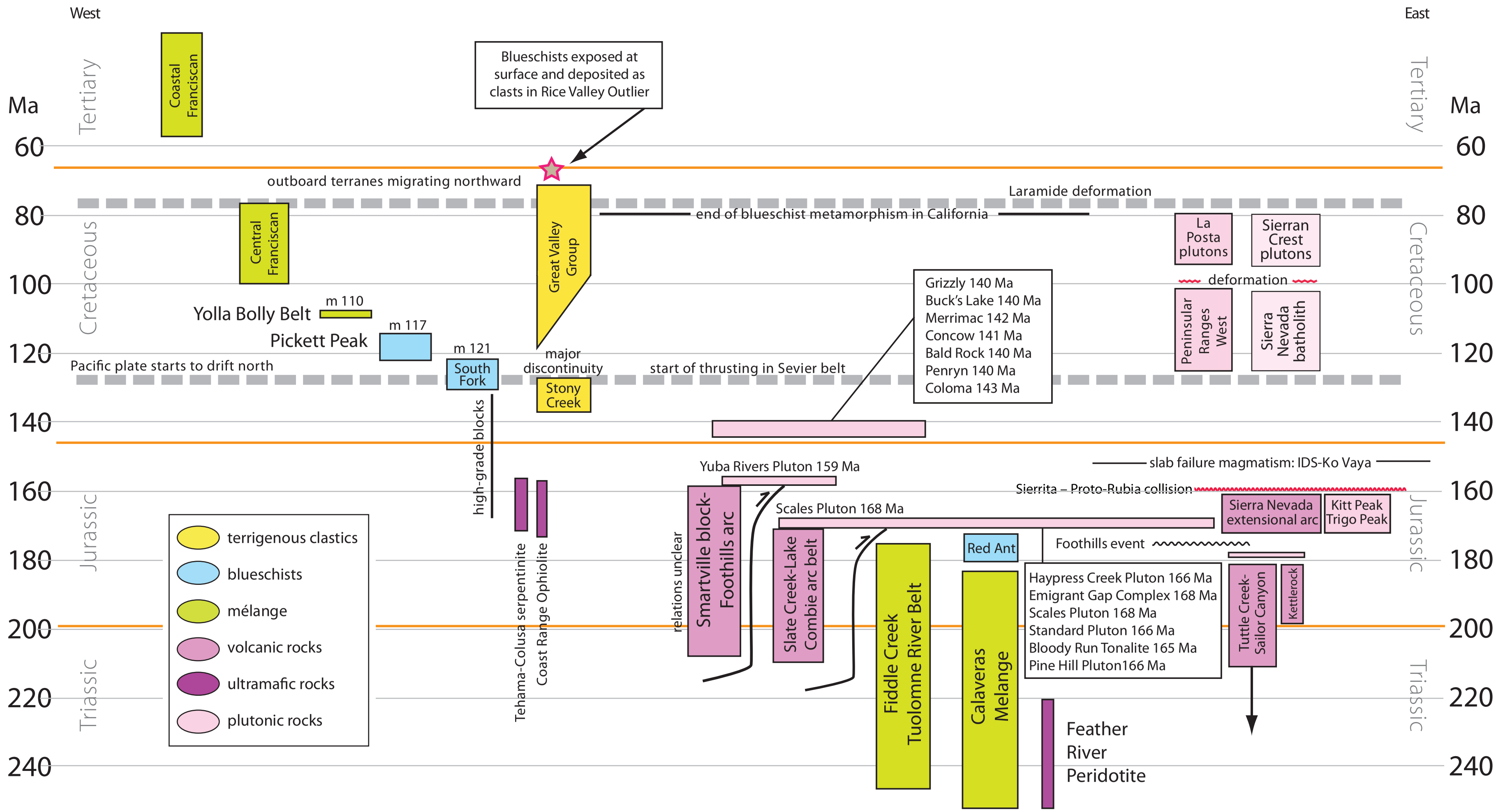

Figure 62. Time-space diagram illustrating the main tectonic, sedimentary, and igneous events from the Sierra Nevada westward to the Pacific coast. Based in part on figures in Dickinson (2008). IDS—Independence dike swarm; Ko Vaya—Ko Vaya intrusions; m—age of metamorphism.

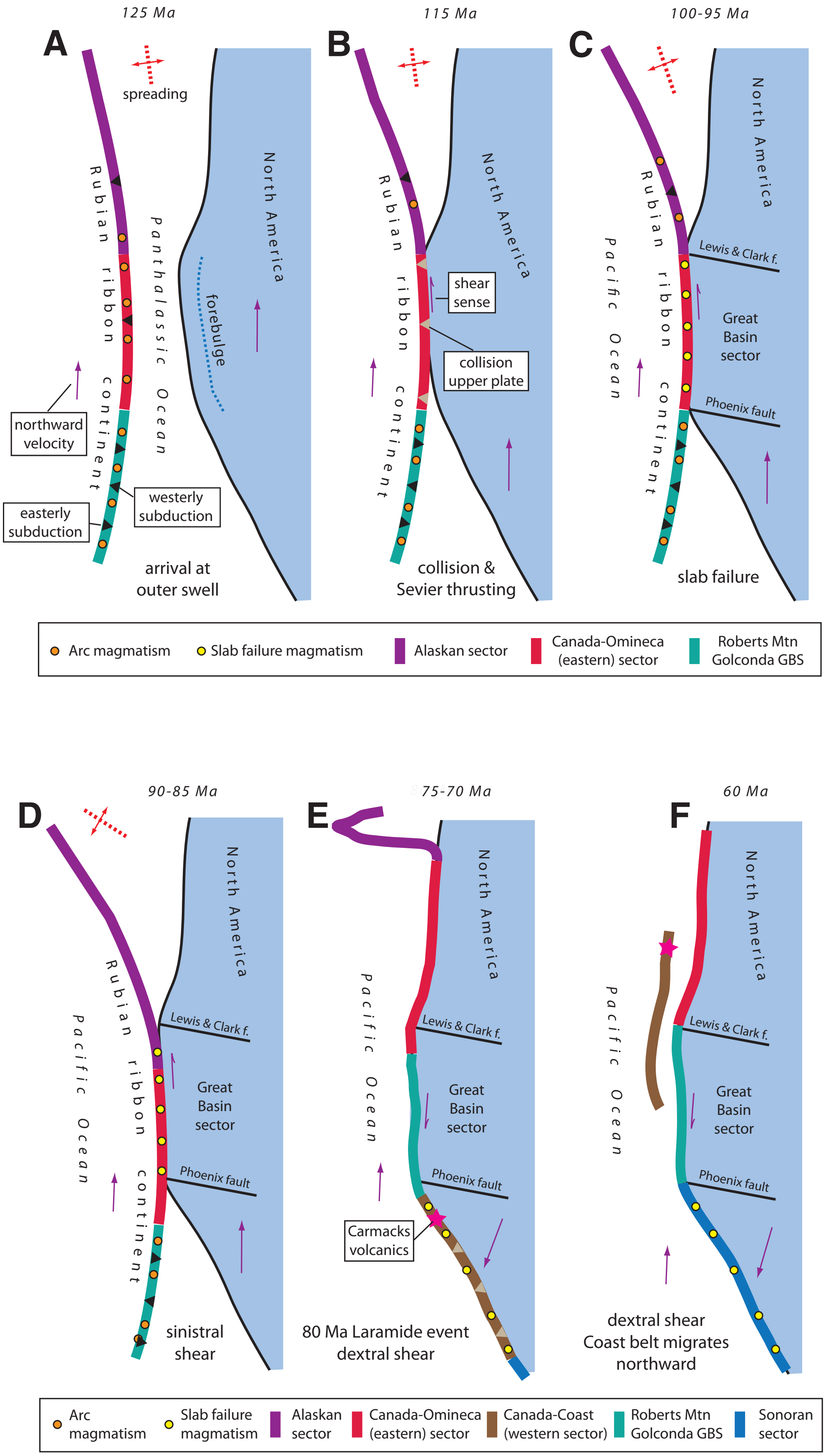

Figure 63. Cartoon illustrating various time slices of the Rubian–North American interactions. (A) ~125 Ma, Rubia arrived in the vicinity of North America; (B) the Rubian ribbon continent collided with the Great Basin sector (GBS) of North America; (C) the westwardly-dipping slab of North America failed at ~100–96 Ma, leading to a pulse of slab-failure magmatism; (D) both North America and Rubia were moving northward at this time, but because the Rubian ribbon continent was moving more slowly, the shear between the two was sinistral; (E) at ~80 Ma North America started to move southward and the entire length of the Rubian superterrane collided with it, the oceanic slab failed to generate slab-failure magmatism, and the Rubian terranes began to move northward with the Pacific ocean plates; and (F) Wrangellia and Stikinia migrated north together, probably on a fault along the margin of Quesnellia, to double the Cordilleran batholith terranes in the Canadian Cordillera.

northern Canada and eastern Alaska where they are known as the Livengood-Tombstone-Tungsten-Fairbanks-Salcha suites. The slab-failure magmatism ended at ~90 Ma and because the collision only took place in the Great Basin sector, slab failure only took place there. The failure of the west-dipping slab in that sector at ~100 Ma explains the lack of slab-failure magmas there after the Laramide event. STEP faults developed in the lower North American craton along both the northern and southern margins of the sector and were active until the final Laramide collision.

At around 100 Ma, diachronous closure of oceanic basins along the western margin of Rubia, such as the Gravina-Nuzotzin of Canada, the unnamed basin east of the Alisitos arc in Baja California, and the cryptic basin located within the Sierra Nevada batholith that also divides it into two halves, led to major transpressional shortening within the Cordilleran batholithic zone (Fig. 43). Presumably, oblique subduction on the western margin of Rubia, like that of present-day Sumatra, was partitioned into an orthogonal subduction component and a strike-slip component with a major fault coinciding more or less with the magmatic front. Pre-collisional subduction with the intervening basin could have been westerly directed as magmatism appears to young eastward in the western blocks. Post-collisional plutons, emplaced dominantly to the east, but locally across the suture, such as those of the 98–92 Ma La Posta suite in Baja California and the 98–85 Ma Sierran Crest magmatic event, might be slab failure magmas or possibly a mixture of subduction and slab failure magmas.

At about 82–80 Ma, during the Laramide event, nearly the entire length of the Rubian ribbon continent collided with North America and a cratonic-vergent thrust belt with associated foredeep developed on the western margin of North America, except within the Great Basin segment, where the collision took place much earlier. During the collision, arc magmatism shut down as the subducting slab connected to the North America craton tore and broke off, taking the rifted margin, and an extensive amount of the miogeoclinal succession with its cratonic basement into the mantle to be recycled. Slab failure readily explains the obvious lack of rift deposits on the passive margin of North America. Compression between North America and the Rubian superterrane led to the thick-skinned thrust belt developed in the Great Basin sector of the orogen. The shutdown of easterly-directed subduction on the western margin of Rubia led to exhumation of the coherent blueschist terranes of the Franciscan complex (Fig. 61).

To the south in the Sonoran block, where the Canadian terranes were residing at that time, the slab failure led to a band of slab-failure magmatism that extended from the Phoenix fault southward through the Coast plutonic complex, the southern end of Quesnellia, and the Belt-Purcell allochthons into the Mojave-Sonoran region, which was apparently connected to those terranes at the time (Figs. 56 and 63). The only magmatism of this age within the Great Basin segment occurred in the Colorado Mineral belt.

The cessation of subduction along the Rubian margins coupled with strong northward motion of the Pacific ocean floor allowed a large coherent piece of Rubia, at that time located within the Sonoran sector along the south side of the Great Basin sector (Figs. 56 and 63), to be captured by the Pacific plates and migrate northward relative to North America, which was moving southward (Kent and Irving, 2010). By 80–75 Ma part of it was impinging on the North American cratonic terrace within the Canadian sector, where it generated the Rocky Mountain fold-thrust belt and the thick clastic wedge located just to the east.

By ~58 Ma, the large-scale northward migration of the main bulk of the Rubian superterrane had ceased and thrusting within the Rocky Mountain fold-thrust belt ended, although some strike-slip motion continued well into the Eocene on discrete faults such as the Tintina and is still active today on the Denali fault. Exhumation of the Canadian sector of the orogen started at ~58 Ma as documented by uplift of the Belt allochthons and erosion of the thick Campanian foredeep within the Western Canada Basin.

Localized gravitational collapse of the orogen happened at different times depending on the timing of collision and mode of thickening. Within the Great Basin sector, core complexes formed during the Paleocene and reflect underplating and thickening during the Sevier phase of shortening; whereas within the Canadian sector, collapse occurred during the Eocene, and is a consequence of thickening during the Sevier event and the Laramide event. The collapse within the Sonoran sector occurred during the Miocene due to thickening during the Laramide event. And in Alaska, the collapse occurred during the Cretaceous following and partly coincident with thickening during the Brookian orogeny.

Regional gravitational collapse that led to the formation of the Basin and Range also occurred during the Miocene and appears to reflect the region where crustal thicknesses were doubled in the area where North American crust was pulled beneath the Rubian superterrane. The resultant middle crust was likely hot and plastic such that rock above it flowed laterally to generate the upper crustal brittle deformation characteristic of the region.

Overall, the amalgamation of Rubia and its interactions with North America outlined here demonstrate that the orogen is readily explained as a typical collisional belt characterized by multiple arc-continent and arc-arc collisions. There is no obvious need to invoke a Cordilleran-type model for its origin.

CARRYING THE OROGEN SOUTHWARD

In northern South America, significant portions of the Great Arc of the Caribbean and its oceanic plateau collided with the South America craton in Venezuela, Colombia, and Ecuador and were accreted to the continent above the west-dipping subduction zone during the Campanian between 73 Ma and 70 Ma (Luzieux et al., 2006; Vallejo et al., 2006; Altamira-Areyán, 2009). Obvious exotic rocks continue southward into northwestern Peru (Feininger, 1980, 1987).

In 2002, Moores et al. presented a speculative model in which they suggested that much of the South American margin was deformed during the Late Jurassic–Cretaceous by arc collision, perhaps to include the allochthonous Mesoproterozoic Arequipa massif of Peru and Bolivia (Ramos, 2008), and closure of the Rocas Verdes ophiolitic basin of Patagonia and Tierra del Fuego. Although large areas of the Scotian marginal basin to the south are poorly known and might represent fragments of Mesozoic arcs and microcontinents (Barker, 2001), it is clear that just as with the Antillean arc, the Scotian arc (Barker et al., 1991) represents a Pacific realm that migrated into the Atlantic (Moores, 1970; Pugh and Convey, 2000), leaving scattered traces strewn along its transform margins with South America and Antarctica (Garrett et al., 1987).

While to some the model of Moores et al. (2002) might seem outrageous, I believe it to have great merit as it readily explains the crustal thickening and Late Cretaceous–early Tertiary foredeeps (DeCelles and Horton, 2003; Arriagada et al., 2006), thrust belts, and proposed basement megathrusts (McQuarrie, 2002) of the central Andes—all of which I find to be poorly explained in current models because they call for huge slices of mid- to upper-crustal rocks to be separated from lower-crustal rocks, which are apparently not shortened, and to be transported hundreds of km inboard by frictional forces on the base of the crust by the subducting slab. Such deformation is more readily accomplished by attempted subduction of the cratonic margin. One might look at the Arequipa block (Ramos, 2008) as the upper plate during a Late Cretaceous collision.

In the south-central Andes of Chile and Argentina, the eastward-vergent, Late Cretaceous Agrio fold-thrust belt and associated foredeep rocks of the Neuquén Group (Cobbold and Rossello, 2003; Ramos and Kay, 2006)—as well as the collision with, and attempted subduction of, the western margin of South America beneath the arc now represented by the Patagonian batholith, at ~75 Ma (Maloney et al., 2011)—provide additional evidence of a major Late Cretaceous collisional orogen extending throughout westernmost South America.

Although much remains to be learned from the complicated geology of the Cordillera of South America, a cursory glance at magmatism in the Coastal batholith of Peru shows striking similarities with those of North America, both in development and timing (Fig. 64). As Cordilleran plutons and Laramide deformation occur down the western margin of South America, just as they do along western North America, the Rubian ribbon continent may once have extended along the entire coast of the Americas. Thus, long-lived hypotheses that major mountain chains can be created entirely by subduction of oceanic lithosphere without accretionary collisions (Hamilton, 1969a, 1969b; Dewey and Bird, 1970; Dalziel, 1986) may need revision.

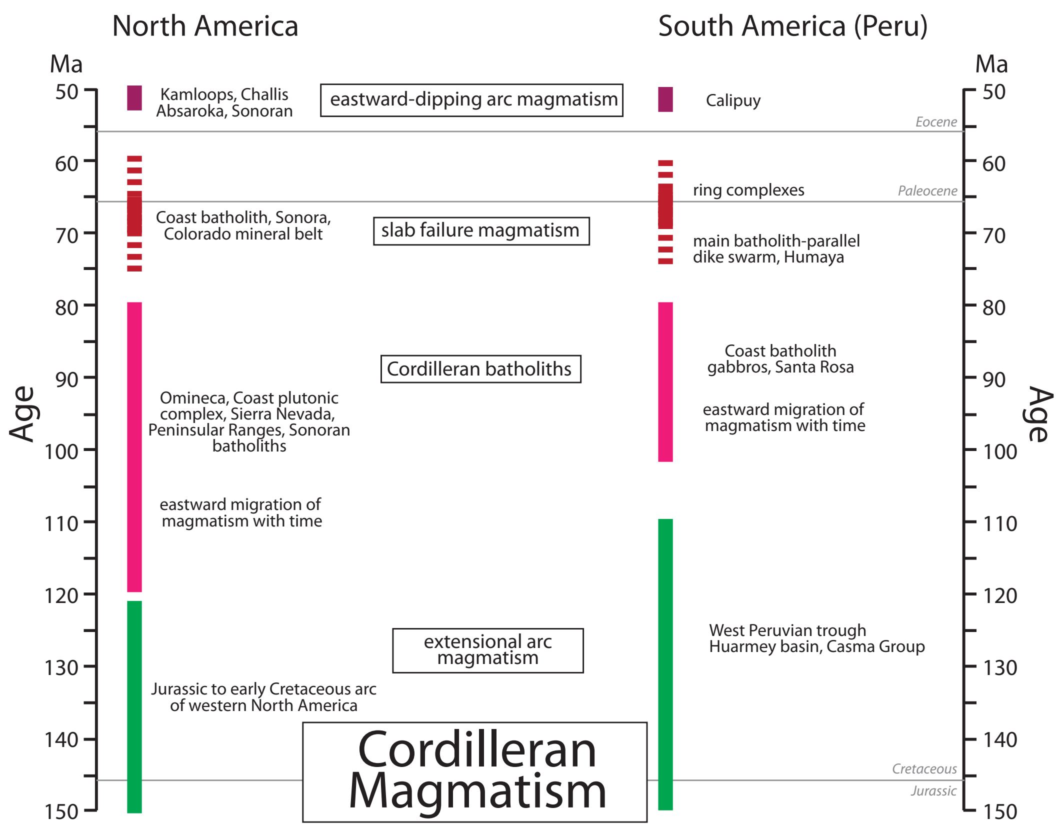

Figure 64. Comparison between North American magmatism and that of Peru for the Cretaceous–Paleocene illustrating the overall similarity in timing and expression. ATL—Atlanta lobe.

PROBLEMS AND DIRECTION FOR FUTURE RESEARCH

Based on the summary presented here, it seems reasonable to conclude that much of the Cordillera migrated southward relative to North America following its collision with North America during the Sevier event, then northward during the Laramide event between 80 and 50 Ma. Two areas in the western United States are poorly studied paleomagnetically: the Sierra Nevada and the Great Basin, and so are difficult to place in proper context. As discussed earlier, existing data from 102 to 97 Ma plutons within the Sierra Nevada show a maximum displacement relative to North America of ~1000 km (Frei, 1986; Kent and Irving, 2010), but the plutons might be folded and remagnetized due to subsequent reheating as younger plutons were emplaced.

Detailed paleomagnetic study of Jurassic and Cretaceous plutonic rocks within the Great Basin area is also important as they are also understudied. Sheet-like, concordant plutons of Jurassic age, such as the one exposed on the west side of the House Range in Utah, are excellent candidates for study as the floor and roof rocks are well bedded and dip gently.

The nature of the basin that originally lay between the halves of the Cretaceous Cordilleran batholiths is poorly known, as is the polarity of subduction within the basin. Allied with these problems is the question of whether post-collisional magmatism and exhumation was generated by slab failure, subduction, or both.

Where were the Franciscan, related accretionary complexes, and the Great Valley fore-arc basin located during the Laramide event? Severe transpressional deformation affected the western and central Sierra Nevada at around 100 Ma, yet sedimentary rocks purported to have been located just to the west, show no evidence of such a deformation. Wright and Wyld (2007) suggested that rocks of the Great Valley group were deposited well to the south in Mexico, but the idea remains to be tested.

Questions involve the timing of displacement on the major faults that trend largely transverse to the orogen and bound the major sectors. These include the Lewis and Clark, Orofino, Snake River Plain, and Phoenix faults. Are the three northern faults confined to the Rubian superterrane? None have conspicuous offsets of North American rocks, yet they appear to have affected deformation and sedimentation there. For example, faults of the Lewis and Clark lineament clearly affected sedimentation in the foredeep (Wallace et al., 1990), and eastward projections of the zone appear to mark the northern limit of Laramide thick-skinned deformation. Similarly, southeastward projections of the Orofino fault appear to coincide with the northern end of the Colorado Plateau. Are the faults now currently cutting the Rubian superterrane manifestations of lower-plate STEP faults?

When was the hypothesized Snake River dextral fault active? It appears to offset 15.5 Ma Lovejoy basalt of northern California's Sierra and Great Valley from the similar-age Steen's basalt of southeastern Oregon (Garrison et al., 2008).

The Phoenix fault, which I earlier suggested (Hildebrand, 2009) to be a transform fault, clearly separates many prominent features such as the Basin and Range, the Oligocene ignimbrite flare-up of Nevada, and the Sierra Madre Occidental; so some might argue that it is instead a young transcurrent fault. Seemingly forming a barrier to this interpretation is the 18.7 Ma Peach Springs tuff, which is interpreted to cross the possible trace of the fault (Glazner et al., 1986), but the fault could be early Miocene in age. Understanding the temporal and spatial relations of these faults is a fundamental precursory requirement for constrained fault trace reconstructions of individual throughgoing faults presumed to have been active during the Late Cretaceous–early Tertiary (Wyld et al., 2006).

Detailed work in the hinterland belt is necessary to better understand the Jurassic deformation. It is possible, as mentioned in the main body of text, that a fragment of Kootenay terrane is located within the area.

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