The Tectonic Setting and Origin of Cretaceous Batholiths within the North American Cordillera

The Case for Slab Failure Magmatism and Its Significance for Crustal Growth

By Robert S. Hildebrand and Joseph B. Whalen

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by

Robert S. Hildebrand
1401 N. Camino de Juan
Tucson, Arizona 85745, USA

Joseph B. Whalen
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario K1A 0E8, Canada

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Cover: Mount Whitney (14,505 ft, 4421 m) and the Sierran escarpment framed by a granitic arch in the famed movie backdrop of the Alabama Hills, just west of Lone Pine, California. The granitic arch developed in the 85 Ma Alabama Hills granite, whereas Mount Whitney and the subjacent escarpment comprise 87–83 Ma granodiorites of the Mount Whitney intrusive suite. The rocks are all interpreted to be the products of slab failure. Photograph by Robert S. Hildebrand.

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Robert S. Hildebrand
1401 N. Camino de Juan, Tucson, Arizona 85745, USA

Joseph B. Whalen
Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8, Canada

ABSTRACT

In the standard model, Cordilleran-type batholiths form beneath volcanic arcs in thickened crust, but our survey of modern and ancient continental arcs revealed most to be regions of normal to thinned crust, not zones of crustal thickening. This suggested to us that the standard batholithic paradigm is flawed.

In order to better understand the batholiths, we explored (1) the 100–84 Ma La Posta and Sierran Crest magmatic suites of the Peninsular Ranges and Sierran batholiths, which formed after the 100 Ma Oregonian event due to closure of the Bisbee-Arperos seaway; (2) plutons and batholiths emplaced into the metamorphic hinterland of the 124–115 Ma Sevier event, which occurred in the Great Basin sector of the United States but, due to younger meridional transport, are now exposed in the Omineca belt and Selwyn Basin of Canada; and (3) Late Cretaceous–early Cenozoic intrusive rocks, such as the Coast, Idaho, and Boulder batholiths, which intruded a metamorphic hinterland during and after the Laramide event. The dominance of syn- to postdeformational emplacement and the distinctive slab failure–type geochemistry indicate that most, but not all, Cretaceous plutons within Cordilleran batholiths formed during and after arc-continent collision as the result of slab failure.

We interpret whole-rock geochemistry, as well as radiogenic and stable isotopes, to indicate that slab failure magmas involve only minor amounts of crust and are derived mainly from plagioclase-absent melting of garnet-bearing rocks in the mantle. Some suites, such as the <100 Ma Oregonian Sierran and Peninsular Ranges batholiths, have evolved Nd and Sr isotopes compatible with old enriched subcontinental lithospheric mantle. The well-known 0.706 $^{87/86}Sr_i$ isopleth appears to separate rocks of Oregonian slab failure from rocks of older arc magmatism and is probably unrelated to any obvious crustal break; instead, it reflects involvement of old subcontinental lithospheric mantle in the slab failure magmas. To expand our findings we examined the geochemistry of Cenozoic slab window and Precambrian tonalite-trondhjemite-granodiorite suites and found them to share many similarities with the Cretaceous slab failure rocks.

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Because most Cretaceous plutons in the North American Cordillera appear to represent juvenile additions to the crust, we argue that substantial volumes of continental crust are formed by slab failure magmatism. Slab failure rocks, especially those emplaced within the epizone, are richly metalliferous and make excellent exploration targets.

INTRODUCTION

The North American Cordillera constitute an intricate geological tapestry developed from complex plate interactions over many eons. Within the kaleidoscope of basins, blocks, and sutures, huge volumes of magma were emplaced into the crust to form linear belts of gregarious plutons, collectively known as Cordilleran batholiths. The great paradox of batholithic petrogenesis is that although experimental studies demonstrate that the requisite magmas can be derived from the crust, neither temperature nor water content are sufficiently high to cause crustal melting without the input of mantle-derived heat, generally inferred to be in the form of mafic magmas rising from the mantle (Dickinson, 1970; Pitcher, 1979; Clemens and Vielzeuf, 1987). Thus, we might better understand Cordilleran batholiths by erecting a general space-time framework that relates the magmatism to mantle melting, regional deformations, and plate kinematics—all driven by mantle dynamics. In this way, we might use magmatism to better understand the development of ancient orogenic belts as well as mantle contributions to crustal growth.

In our recent essays, we presented a model in which Cordilleran-type batholiths were generated by combinations of two fundamentally different kinematic processes: subduction and slab break-off, each of which commonly yields unique and distinct magmas derived from different mantle sources and emplaced in contrasting tectonic regimes, either before or during and after deformation: Arc magmas predate collisional deformation, whereas magmas generated by slab break-off are syn- to postdeformational (Hildebrand, 2013; Hildebrand and Whalen, 2014a, 2014b).

The typical temporal progression relates directly to ocean closing and progresses from subduction-related to slab failure magmatism, with a period of deformation between the two. During arc-continent collision, continental margins are pulled beneath arcs, so magmas related to the failure of the subducting slab are emplaced into rising, tectonically thickened and exhumed crust, in great contrast to the older arc magmatism, which erupted onto and/or was emplaced into normal to extensionally thinned crust. Thus, in their simplest form, we envision Cordilleran batholiths to be composed of two parts that may or may not overlap spatially: (1) an older belt of mesozonal to epizonal arc plutons commonly emplaced into their own cover, followed closely by (2) a parallel belt of generally more deeply seated plutons emplaced during crustal thickening and exhumation. Both are dominantly upper-plate phenomena, but they have very different causes and sources. More complex scenarios involve younger arc magmatism due to subduction reversal following collision and possible dismemberment and translation along orogen-parallel strike-slip faults (Hildebrand and Whalen, 2014a, 2014b; Hildebrand, 2015).

In this overview, we greatly expand our initial treatment with additional examples from the Cretaceous of the North American Cordillera to show how a better knowledge of magmatism can contribute to our overall understanding of the orogen. We start with an overview of the two contrasting tectonic regimes and then present an overview of the Peninsular Ranges batholith, which we use as a template to explore and characterize the origin of several other Cretaceous batholiths within the North American Cordillera. We then utilize our general findings to explore their implications for the development of the Cordilleran tapestry, the process of ridge subduction, the importance of tonalite-trondhjemite-granodiorite (TTG) suites, and the origin of continental crust.

Arcs

Hamilton (1969a, 1969b) hypothesized that because the only modern volcanic fields of comparable size to the great Mesozoic batholiths of the western Americas are Andean-type volcanic fields, batholiths must exist beneath volcanic arcs. The idea took hold, and ever since, most geologists have assumed that Cordilleran batholiths were generated above subduction zones in high-standing regions of great crustal thickness generated by voluminous arc magmatism (Bateman, 1992; Ducea, 2001; Ducea and Barton, 2007; Grove et al., 2008; Mahoney et al., 2009; DeCelles et al., 2009, 2015; Ducea et al., 2015a, 2015b). This reasoning was based largely on the abundant volcanism (Siebert et al., 2011), the high general elevation of the Andes, the deep Sierran and Andean crustal roots, their linear nature, and the apparently continuous magmatism through time (Lawson, 1936; Bateman and Wahrhaftig, 1966; Bateman et al., 1963; Dodge and Bateman, 1988; Christensen, 1966; Beck and Zandt, 2002; Beck et al., 1996; Yuan et al., 2002; McGlashan et al., 2008; Bianchi et al., 2013; Cao et al., 2015). Recently, some researchers suggested that arcs contain voluminous pulses of magmatism—so-called arc flare-ups—that follow crustal thickening and are generated either by retro-arc thrusting or magmatic underplating (Ducea, 2001; DeCelles et al., 2009, 2015; Ducea and Barton, 2007; Ducea et al., 2015b; Lee and Lackey, 2015).

Our evaluation suggests that those ideas are largely incorrect. When we examine all of today's active continental arcs, we find that the majority are—contrary to the batholithic paradigm—characterized by sequences of rocks erupted and deposited within subsiding basins, not volcanoes sitting on high-standing thick crust (Levi and Aguirre, 1981; Hildebrand and Bowring, 1984; Busby-Spera, 1988; Busby, 2012). Young examples include both the modern and ancestral Miocene Cascades, where the volcanoes and their debris aprons subsided in graben (Fiske, 1963; Williams and McBirney, 1979; Taylor, 1981; Mooney and Weaver, 1989; Smith et al., 1987); the low-standing Alaskan Peninsula (Burk, 1965; Fliedner and Klemperer, 2000), where volcanoes such as Augustine are partially submerged in Cook Inlet (Power et al., 2010); the Kamchatka Peninsula of easternmost Russia, where towering stratovolcanoes erupt in extensive fault-bounded troughs close to sea level (Erlich, 1968, 1979; Levin et al., 2002); the North Island of New Zealand, where the Taupo zone sector of the arc is actively extending as calderas and stratocones erupt (Houghton et al., 1991; Harrison and White, 2006; Downs et al., 2014); the Central American arc, where volcanoes are aligned in a long, linear, low-standing depression (Williams et al., 1964; Williams and McBirney, 1979; Burkart and Self, 1985; MacKenzie et al., 2008), and the Hellenic arc, where volcanoes form islands in the Aegean Sea (Druitt et al., 1989; Druitt and Francaviglia, 1992).

Furthermore, the stratigraphy within pendants and wall rocks of Cordilleran batholiths provides no evidence for thick crust, as the volcanic rocks are commonly intercalated with shallow-marine sedimentary rocks, and therefore sat at sea level or below during volcanism. The Sierra Nevada arc was low standing during volcanism, because it contains intercalated marine rocks deposited as late as 100 Ma (Nokleberg, 1981; Saleeby et al., 2008; Memeti et al., 2010a). Likewise, in the western Peninsular Ranges, arc volcanic rocks of the Santiago Peak and Alisitos groups are interbedded with sedimentary rocks containing marine fossils (Fife et al., 1967; Allison, 1974; Phillips, 1993; Griffith and Hobbs, 1993; Wetmore et al., 2005; Busby et al., 2006; Centeno-García et al., 2011). In South America, the 9-km-thick Casma arc volcanics, in part the wall rocks for the younger Coastal batholith of Peru, are dominantly marine (Cobbing, 1978, 1985; Atherton et al., 1985), as are many of the thick Jurassic–Cretaceous arc rocks within the Ocoite arc of northern Chile (Levi and Aguirre, 1981; Åberg et al., 1984). Even Precambrian continental arcs, such as the Paleoproterozoic Great Bear magmatic zone of the Wopmay orogen, were low-standing zones of subsidence during magmatism (Hoffman and McGlynn, 1977; Hildebrand and Bowring, 1984). Thus, where arcs are concerned, the high-standing Andes are the outlier—simply atypical of modern and ancient arcs—likely because they were built atop a Late Cretaceous–Paleogene collisional belt (Hildebrand and Whalen, 2014a). Overall, the low-standing nature of arcs differs from collision zones, which are typically regions of abnormally thick crust, rapid exhumation, and a compositionally different suite of magmatic rocks.

Slab Failure

Because the continents are very old, and oceanic lithosphere is young, every collision that entails the closure of an ocean basin wide enough to drive collision must involve break-off of the subducting slab; the alternatives are for continents to be subducted or for slabs to dangle off continental margins into the mantle. Neither is observed. Therefore, slab failure and break-off are integral components of plate tectonics and natural consequences of subduction (Roeder, 1973; Price and Audley-Charles, 1987; Sacks and Secor, 1990; Davies and von Blanckenburg, 1995; Davies, 2002; Atherton and Ghani, 2002; Cloos et al., 2005). This is because the buoyancy forces resisting the subduction of continental lithosphere are as large as those pulling oceanic lithosphere downward (Cloos et al., 2005). Eventually, the greater density of the oceanic lithosphere causes the lower plate to break or tear—presumably by viscous necking (Duretz et al., 2012)—at its weakest point and sink into the mantle. The detachment can be relatively rapid, within 1 m.y. (Bercovici et al., 2015), consistent with geological observations (Dewey, 2005; Hildebrand et al., 2010; Hildebrand and Whalen, 2014b).

Once the subducting slab tears, and the lower plate is detached from its oceanic anchor, rocks of the partially subducted continental margin rise and are exhumed due to buoyancy forces (Duretz et al., 2011; Bercovici et al., 2015). The failure also allows asthenosphere to upwell through the tear, melt adiabatically, and rise into the collision zone, where it interacts especially with subcontinental mantle lithosphere and possibly crust of the upper plate. The resulting magmas, which form linear arrays above tears in the descending slab, are generated by adiabatic melting as mantle upwells through the breach in the slab (Macera et al., 2008). They commonly overlap the terminal stages of deformation. As the magmas dominantly intrude the upper plate, they commonly form a linear belt atop or alongside the old arc and appear temporally continuous with the older magmatism, and so they can be readily confused with it. Magmas might also invade rocks of the foredeep and/or the shortened passive margin of the lower plate (Hoffman, 1987; Hildebrand and Bowring, 1999; Hildebrand et al., 2010).

Several factors are important in slab failure and govern where and when during the collision the slab will rupture (Fig. 1). The age of the subducting lithosphere is perhaps the most important factor because young lithosphere is weaker, and therefore break-off will be fast, commonly less than 1 m.y. after initiation of collision (Duretz et al., 2012), and it will occur at shallow levels, whereas with older, thicker, and stronger lithosphere, break-off is slower, and the continental edge is subducted deeper into the mantle. The depth of break-off largely controls the width of the orogen, for it is the rebound of the partially subducted continent that will lead to the region of intense uplift and exhumation (Duretz et al., 2011, 2012; Duretz and Gerya, 2013). Thus, shallow break-off creates narrow orogens, lower-grade metamorphism, and intense, rapid and higher rates of exhumation, whereas deep break-off creates broad orogens with higher grades of metamorphism and slow, more subdued rebound and exhumation (Duretz et al., 2011). In cases of deep break-off, the resultant uplift and exhumation might cause the arc of the upper plate to be eroded such that only vestiges of the upper plate and

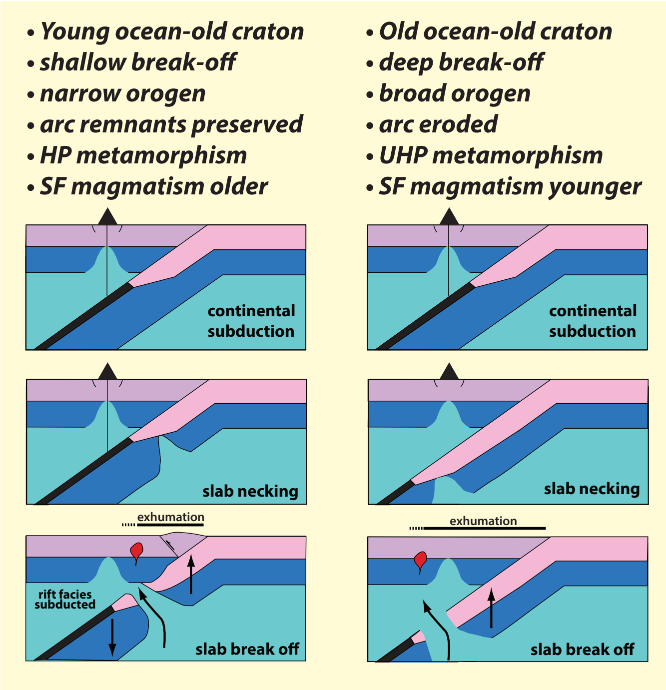


Figure 1. Slab failure controls many aspects of orogenic belts. Old lithosphere is thick and strong, and so it leads to deeper break-off than weak, young lithosphere. Deeper break-off creates broader orogens with younger slab failure magmatism. Because the continental edge is carried deeper, it is metamorphosed to higher grade than if break-off were shallow. Similarly, deeper break-off leads to slab failure magmatism farther from the collisional suture. HP—high pressure; UHP—ultrahigh pressure; SF—slab failure.

its arc remain, and the bulk of arc material might reside in adjacent basins.

The depth of break-off likely also controls the volume of slab failure magmatism because in cases of shallow failure, the asthenosphere upwells to shallower depths, which will generate greater volumes of melt due to adiabatic melting (McKenzie and Bickle, 1988). Additionally, during shallow failure, the upwelling asthenosphere, say at ~100 km depth, creates an advective heat source capable of generating melts in the lithospheric mantle and possibly the crust (van de Zedde and Wortel, 2001). Thus, shallow break-off will create larger quantities of magma than deep break-off, and they might be compositionally varied.

If continents are moving, such as during the post-Jurassic westward migration of North America as the North Atlantic widened, then magmatism might occur farther inboard with time and could mimic arc migration due to slab flattening. Also, the cooled mantle generated by the torn and descending slab causes dynamic subsidence of the overlying lithosphere, and so an older relatively narrow foredeep basin generated by flexural subsidence is supplanted by a broad dynamic basin caused when the continent passes over the cooler mantle (Gurnis et al., 1998; Gurnis and Müller, 2003). We turn now to the Peninsular Ranges batholith, which we argue provides clear-cut examples of both arc and slab failure magmatism.

PENINSULAR RANGES BATHOLITH: SETTING, GEOCHEMISTRY, AND PROOF OF CONCEPT

Our previously published synthesis of the Peninsular Ranges batholith (Hildebrand and Whalen, 2014b) provides a robust template to resolve the complexities of magmatism elsewhere because: (1) the geochemical, isotopic, and geochronological data are modern and of high quality (Lee et al., 2007; Gastil et al., 2014); and (2) the geology is reasonably well known (Morton and Miller, 2014). Here, we provide a synopsis of the geology, present our findings, and then augment this template with modern geochemical data from much younger belts worldwide to substantiate our findings from the Peninsular Ranges batholith.

The Cretaceous Peninsular Ranges batholith is exposed from the Peninsular Ranges of southern California to the state line midway down the Baja Peninsula (Fig. 2), but on the basis of aeromagnetic data, it extends all the way down the Baja Peninsula beneath younger cover (Langenheim et al., 2014). Recently dredged and dated samples from the southern Gulf of California, as well as geological mapping, U-Pb dating, and detrital zircon studies on the Mexican mainland, demonstrate that prior to rifting and spreading in the Gulf of California (Lonsdale, 1989; Umhoefer, 2011), the batholith and its wall rocks extended eastward onto the mainland at least as far south as Zihuatanejo (Duque-Trujillo et al., 2015; Henry et al., 2003; Centeno-García et al., 2011) and formed a continuous magmatic belt constructed across various terranes within the Guerrero superterrane (Fig. 2).

Rocks of the batholith intrude 128–106 Ma volcanic and intercalated shallow-marine to subaerial sedimentary rocks of the Santiago Peak–Estelle Mountain and Alisitos groups, generally interpreted to represent rocks of a magmatic arc (Larsen, 1948; Tanaka et al., 1984; Buesch, 1984; White and Busby-Spera, 1987; Gorzolla, 1988; Anderson, 1991; Wetmore et al., 2003; Schoellhamer et al., 1981; Busby et al., 2006; Herzig and Kimbrough, 2014). Basement to the arc rocks is a wide variety of Mesozoic, Paleozoic, and Proterozoic metasedimentary and plutonic rocks (Brown, 1980, 1981; Todd et al., 1988; Gastil et al., 1991; Campbell and Crocker, 1993; Leier-Engelhardt, 1993; Kimbrough et al., 2014a; Shaw et al., 2003, 2014; Hildebrand and Whalen, 2014b).

Most researchers divided the batholith into western and eastern sectors on the basis of distinctive geochemistry, stable and radiogenic isotopic composition, and the occurrence of different opaque minerals (Gastil, 1975; Silver et al., 1979; DePaolo, 1981; Gromet and Silver, 1987; Taylor and Silver, 1978; Gastil et al., 1990; Kimbrough et al., 2001; Tulloch and Kimbrough, 2003). We chose a different path and divided plutons into groups (Fig. 3) purely on the basis of whether they predate or postdate a regional period of deformation dated as ca. 100 Ma (Premo and Morton, 2014; see also discussion in Hildebrand and Whalen, 2014b, p. 411). We grouped plutons older than 100 Ma into the Santa Ana

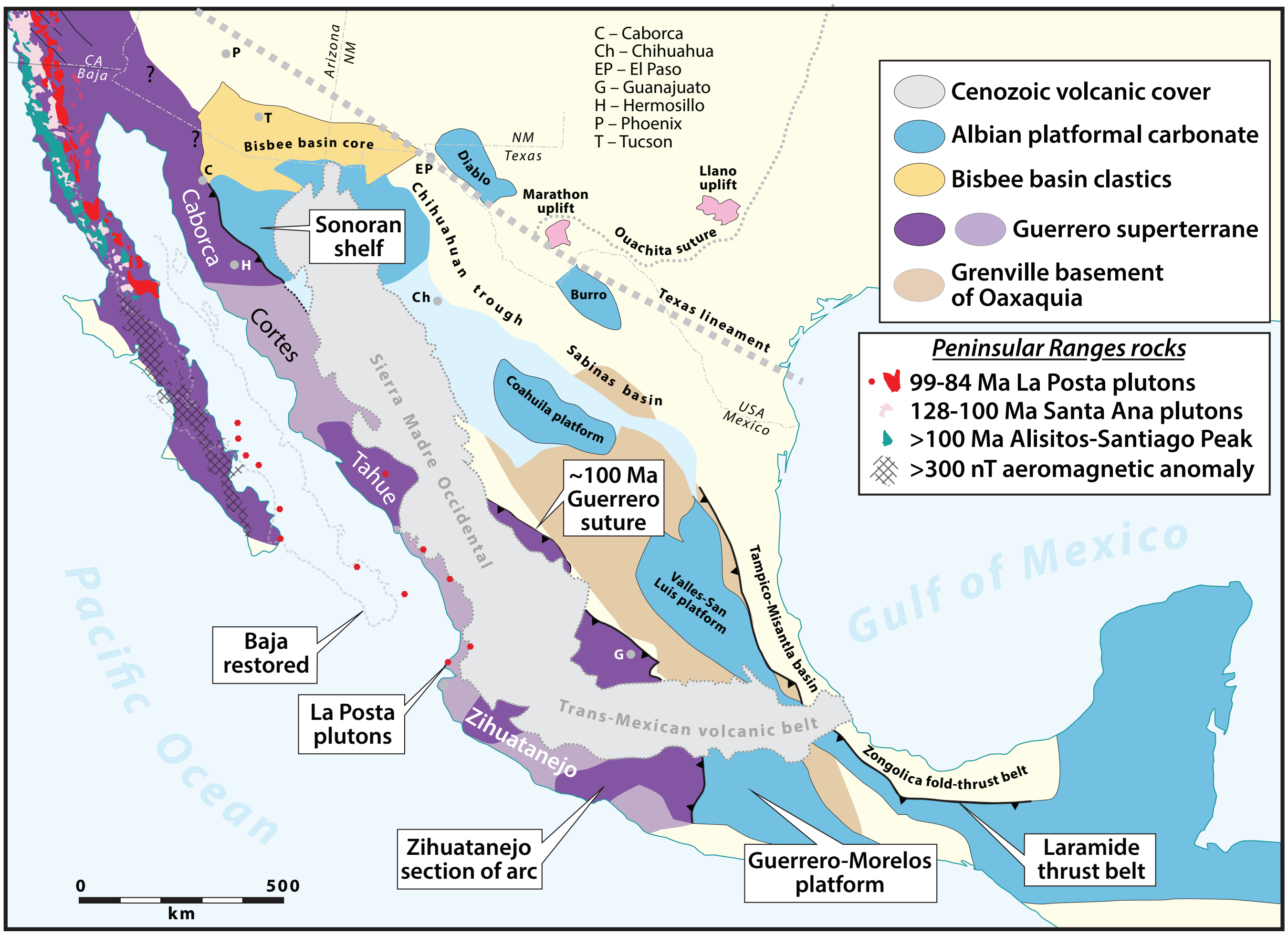


Figure 2. Sketch map modified from Hildebrand and Whalen (2014b) illustrating key geological units of the Peninsular Ranges batholith and Aptian–Albian volcano-sedimentary rocks of the Alisitos–Santiago Peak arc, various terranes of the Guerrero superterrane, and Albian carbonate platforms, mostly located west of the younger Laramide suture and its related fold-and-thrust belt. The Peninsular Ranges batholith continues the length of Baja California, as indicated by a conspicuous aeromagnetic anomaly (Langenheim et al., 2014), but it is buried by younger volcanic rocks south of the state line. Red dots represent drilled and dated core from La Posta plutons (Duque-Trujillo et al., 2015). Rocks of similar age and lithology to those of the Peninsular Ranges batholith crop out in Zihuatanejo (Centeno-García et al., 2011). Westward-facing Albian carbonate banks of the Sonora and Guerrero-Moreles platforms were pulled westward beneath rocks of the Guerrero superterrane at 100 Ma during closure of the Bisbee-Arperos seaway.

suite and assigned younger bodies, which are 99–85 Ma, to either the La Posta or Santa Rosa suites, depending on their location. Rocks of the Santa Rosa suite are slightly younger (Fig. 3) and occur farther east than do those of the La Posta suite, and they sit structurally above the Eastern Peninsular Ranges mylonite zone, but their age and composition suggest to us they are analogous to rocks of the La Posta suite.

The plutonic rocks of the Santa Ana suite constitute a variably deformed, calcic I-type suite ranging in composition from gabbro to granite, and they are 128–100 Ma in age (Johnson et al., 1999, 2002; Tate and Johnson, 2000; Schmidt and Paterson, 2002; Schmidt et al., 2009; Wetmore et al., 2005; Shaw et al., 2014; Premo et al., 2014; Clausen et al., 2014; Morton et al., 2014; Silver and Chappell, 1988; Todd et al., 2003). The plutons are both normally and reversely zoned, isotropic to foliated or protomylonitic, sheeted plutonic complexes. They are highly variable in composition, ranging from tonalite through quartz diorite and granodiorite to leucomonzogranite, locally with abundant wall-rock screens and mafic inclusions, and they contain varying proportions of clinopyroxene ± orthopyroxene ± biotite ± hornblende (Todd et al., 2003; Todd, 2004). Morton et al. (2014) noted that the westernmost plutons are isotropic, whereas those farther east are foliated, so that there is a megascopically visible deformational gradient from west to east.

Plutonic rocks of the La Posta–Santa Rosa suite young eastward (Ortega-Rivera, 2003; Shaw et al., 2014) and are dominated

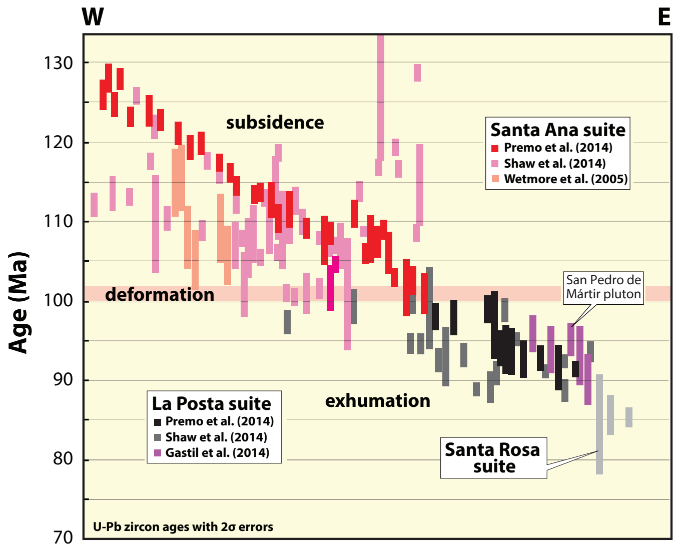


Figure 3. U-Pb zircon ages with 2σ errors for the Peninsular Ranges batholith plotted versus general longitude. The light salmon–colored horizontal line represents the time of deformation. Rocks older than 100 Ma were emplaced during subsidence, whereas rocks younger than 100 Ma were emplaced during exhumation. Data are from Premo et al. (2014), with additional ages from Shaw et al. (2014), Gastil et al. (2014), and Wetmore et al. (2005). The pluton ages prior to 100 Ma are not aligned by geography, but by age. Most workers have recognized that the western Santa Ana suite did not migrate with time (Silver and Chappell, 1988; Shaw et al., 2014), whereas some workers have argued that the post–100 Ma plutons are younger eastward (Ortega-Rivera, 2003).

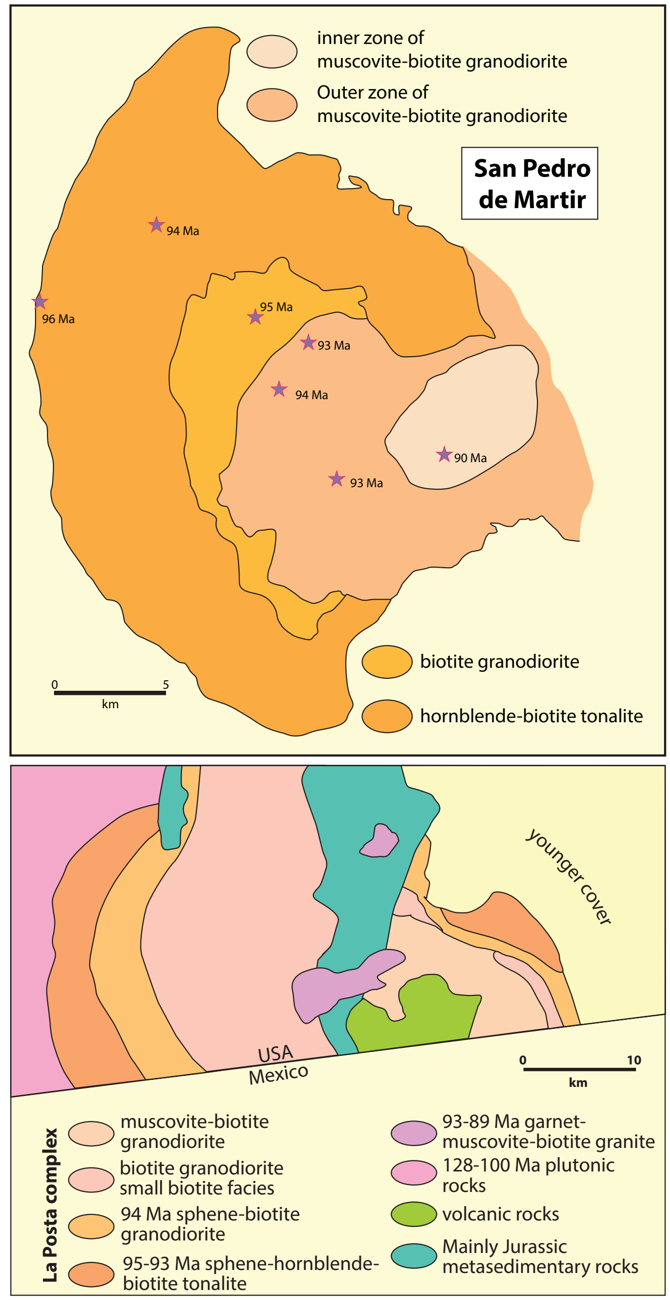


Figure 4. Geological sketch maps showing the geology of two intrusive complexes emplaced between 100 and 90 Ma within the Peninsular Ranges batholith: the Sierra San Pedro de Mártir intrusive complex in Baja California, with U-Pb zircon ages (Gastil et al., 2014), and the La Posta intrusive complex of Southern California (Walawender et al., 1990).

by large, concentrically zoned, mostly weakly foliated, complexes (Fig. 4) comprising biotite- and hornblende-bearing tonalitic marginal phases grading inward over several tens of meters to granodiorite and cored by granodiorite or granite, in places containing both biotite and muscovite (Hill, 1984; Silver and Chappell, 1988; Walawender et al., 1990). Euhedral titanite is characteristic of the La Posta plutons (Silver and Chappell, 1988), and members of this suite were emplaced both above and below the basal unconformity and in part overlap in space with the Santa Ana suite (Johnson et al., 1999; Schmidt and Paterson, 2002; Shaw et al., 2014), although most were emplaced farther east at greater depth within the basement complex.

Rocks of the La Posta–Santa Rosa suite appear to extend eastward across the Gulf of California into Sonora and Sinaloa, where a compositionally similar suite of 101 ± 2 to ca. 90 Ma foliated to nonfoliated tonalite plutons occurs within 50 km of the coast (Henry et al., 2003). They were intruded into Jurassic and older greenschist-grade rocks common to the eastern side of the Gulf of California (Gastil and Krummenacher, 1977; Gastil, 1979; Ramos-Velázquez et al., 2008). K-Ar biotite and hornblende ages are also in the range 100–90 Ma (Henry et al., 2003).

Overall, plutons of the La Posta and Santa Rosa suites represent an intense magmatic pulse ranging in age from 99 to 85 Ma (Silver and Chappell, 1988; Walawender et al., 1990; Kimbrough et al., 2001; Premo et al., 2014). The plutons were emplaced at depths of 5–23 km into upper-greenschist-grade, but mainly amphibolite-grade, wall rocks that are in many places migmatitic (Gastil et al., 1975; Ague and Brimhall, 1988; Todd et al., 1988, 2003; Grove, 1993; Rothstein, 1997; Rothstein and Manning, 2003; Gastil et al., 2014). The plutons were intruded during a period of exhumation when rocks at depths of 15–23 km were brought rapidly to the surface by detachment faulting and collapse (Krummenacher et al., 1975; George and Dokka, 1994; Ortega-Rivera et al., 1997; Ortega-Rivera, 2003; Grove et al., 2003; Miggins et al., 2014). Large volumes of rock were eroded from the uplifted terrane, as documented by abundant 100–90 Ma detrital zircon and feldspar deposited as part of a voluminous pulse of early Cenomanian to Turonian coarse clastic sedimentation in basins located to the west (Lovera et al., 1999; Kimbrough et al., 2001).

In our overview of the Peninsular Ranges batholith (Hildebrand and Whalen, 2014b), we argued that the regional 100 Ma deformation took place when the Santiago Peak–Alisitos arc collided with a Lower Cretaceous passive margin, located to the east and topped by a west-facing Albian carbonate platform terrace, which was uplifted and eroded as it passed over the outer swell, and then pulled into the trench, buried by orogenic flysch, and overthrust by exotic allochthons containing slices of the arc and its basement. The west-facing platform (Fig. 2) is intermittently exposed from Sonora (Warzeski, 1987; Lawton et al., 2004; Lapierre et al., 1992a; Monod et al., 1994; Centeno-García et al., 2008; González-León et al., 2008; Martini et al., 2012), where it is known as the Sonoran shelf, to Zihuatanejo, where it is called the Guerrero-Morelos platform (Monod et al., 1994, 2000). The polarity of the subduction was westward as the western edge of the passive margin was partially subducted beneath the arc. The basin, which we named the Bisbee-Arperos seaway, was apparently a marginal basin open for ~30 m.y. and was of unknown width, although it must have been sufficiently wide to be floored by oceanic crust in order to drive the 100 Ma collision.

Within 1–2 m.y. of collision, the earliest plutonic complexes of the 99–85 Ma La Posta–Santa Rosa suite (Kimbrough et al., 2001; Premo et al., 2014) were emplaced into the forearc and arc regions of the upper plate during uplift and exhumation (Fig. 3). They were emplaced contemporaneously with deposition of thick Cenomanian–Turonian clastic successions to the west, which involved a rapid change from flysch-type sedimentation to coarse cobbly and bouldery molasse during the early Cenomanian (Kimbrough et al., 2001). These are all characteristics of shallow slab break-off, as described earlier, so we argued (Hildebrand and Whalen, 2014b) that the pre–100 Ma Santa Ana suite was a product of arc magmatism, whereas the post–100 Ma La Posta–Santa Rosa suite was generated by slab break-off and adiabatic melting of asthenosphere as it rose through the tear.

Our study of the geochemical variations in rocks of the Peninsular Ranges batholith showed that rocks of the Santa Ana arc suite span a broad silica range, from 48 to 77 wt% SiO_2, whereas the range of the La Posta–Santa Rosa postcollisional suite is much more limited, as nearly all samples contain 60–70 wt% SiO_2 (Fig. 5). In general, mainly metaluminous compositions and amphibole-bearing mineralogy indicate that both suites are I-type (Silver and Chappell, 1988; Chappell and Stephens, 1988). Rocks of the Santa Ana suite define the type calcic suite (Whalen and Frost, 2013), whereas La Posta samples exhibit a calcic to calc-alkalic affinity. For samples with silica contents lying between 60 and 70 wt% SiO_2, rocks of the La Posta suite tend to exhibit significantly higher Ba and Sr and lower Y contents than samples from the Santa Ana suite (Figs. 5C, 5D, and 5E). While both suites have similar Rb contents, the suites are readily divisible on a Rb-Sr plot (Fig. 5F) due to contrasting Sr contents (Walawender et al., 1990; Tulloch and Kimbrough, 2003). Rocks of the Santa Ana and La Posta suites also display marked differences in La/Yb, Gd/Yb, Nb/Y, and Sr/Y values, and these are readily apparent on Harker-type variation plots (Fig. 6).

Although we firmly believe that the best indicator of slab failure magmatism is its syn- to postcollisional timing, we tried to discriminate between arc and slab failure magmas by exploiting the differences in trace-element concentrations (Fig. 7) following the empirical method of Pearce et al. (1984). We did this by first observing the separation of trace-element concentrations on histograms (Fig. 6) to arrive at values of La/Yb, Gd/Yb, Nb/Y, and Sr/Y that separate the largest numbers of pre- and postdeformational samples (Hildebrand and Whalen, 2014b). As can be seen from Figure 8, the pre- and postdeformational rocks of the Peninsular Ranges batholith fall predominantly into two discrete groups or fields on all three plots. Derivative plots (Fig. 9) avoid the use of Sr, which is commonly mobile during alteration, especially at SiO_2 >70%. We also found that plots involving Sm/Yb, which emphasize the differences in depth of melting (Putirka, 1999), do a good job of separating the two suites (Fig. 10).

We plotted the elemental concentrations from both suites on the Nb/Y and Ta/Yb discrimination plots of Pearce (Pearce et al., 1984; Pearce, 1996) but found that, while the two suites formed discrete clusters, those clusters did not coincide with Pearce's originally designated fields. We therefore decided to create new fields based on our well-constrained data set (Fig. 11) and then test the general applicability of the revised fields on those, as well as on our other discrimination diagrams, with Quaternary arc and slab failure examples (Fig. 12). For these plots, we chose representative analyses from relatively recent eruptions that had major elements with minor and rare earth elements (REEs) analyzed by inductively coupled plasma–mass spectrometry (ICP-MS). We also examined a few additional suites, considered by others on the basis of geological relations to be syn- to postcollisional magmatism (Figs. 13 and 14). Both the Quaternary and older examples support our geochemical discrimination based on samples from the Peninsular Ranges batholith, and therefore we are confident that they will work for other belts. Our plots are unlikely to work well with cumulates and rocks with SiO_2 greater than 70%–73%, as they are unlikely to represent liquid compositions.

SIERRA NEVADA BATHOLITH

Plutonic rocks of the 125–84 Ma Sierra Nevada batholith intrude Paleozoic and Mesozoic metasedimentary, metavolcanic, and older plutonic rocks (Stern et al., 1981; Chen and Moore, 1982; Bateman, 1992; Irwin, 2003). Some workers have

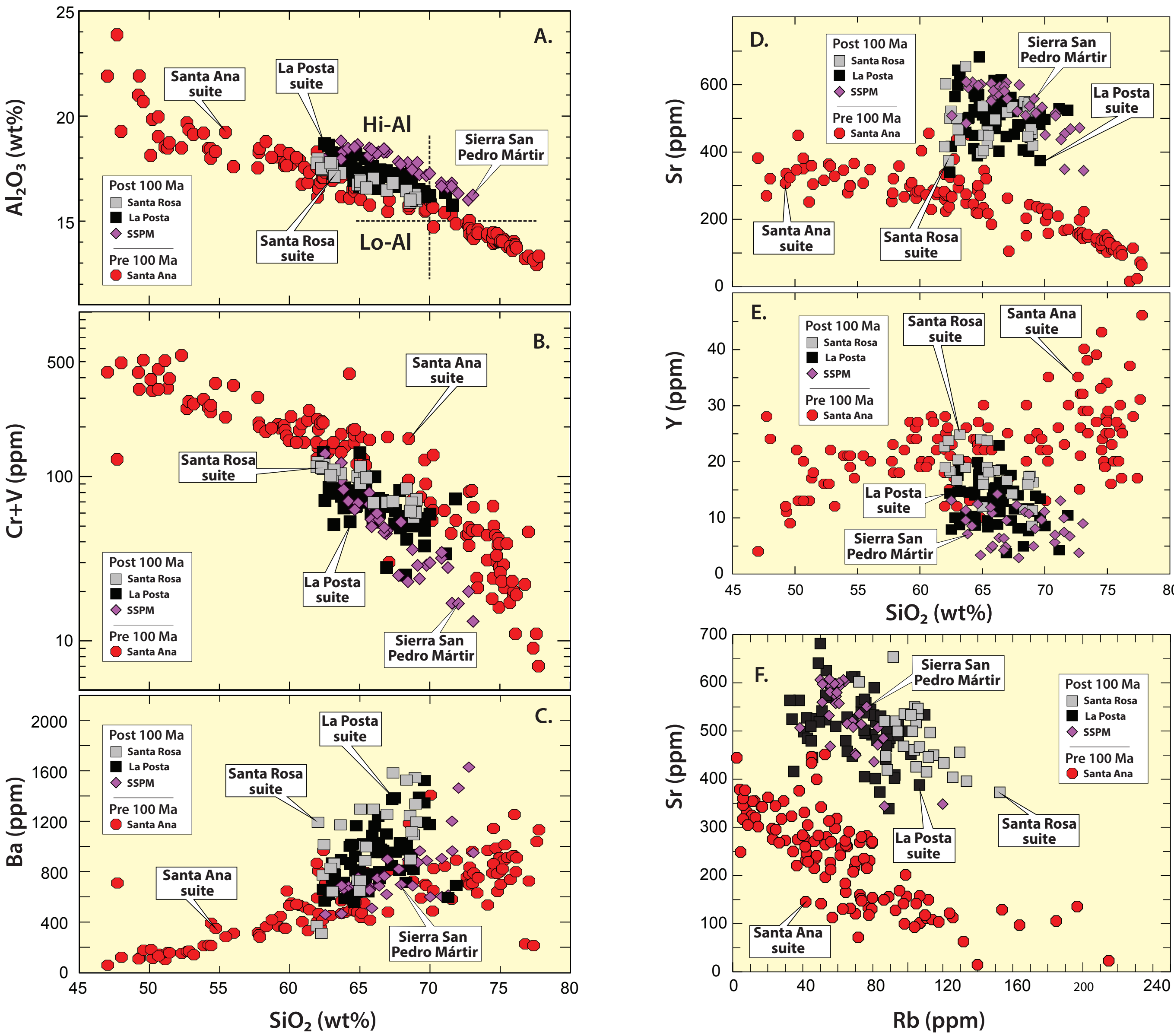


Figure 5. Harker variation diagrams for (A) Al_2O_3; (B) Cr + V; (C) Ba; (D) Sr; (E) Y; and (F) Sr-Rb for Peninsular Ranges batholith sample groups. In A, the dividing line of Barker (1979) between high-Al (>15 wt%) and low-Al tonalite suites at SiO_2 = 70 wt% is shown. Sample subdivision is shown in the symbol legend and discussed in the text. Note that on most of these plots, pre- and postdeformational plutons are compositionally distinct. Figure is modified from Hildebrand and Whalen (2014b). SSPM—Sierra San Pedro de Mártir.

suggested that Jurassic, and even Triassic, plutonic rocks should also be included in the batholith (Paterson et al., 2014; Ducea et al., 2015a, 2015b; DeCelles et al., 2009), but those appear to have been formed by arc and slab failure magmatism related to the accretion of exotic arc terranes to the narrow Paleozoic core during the Jurassic and earliest Cretaceous (Irwin, 2003; Hildebrand, 2013). Pulses of magmatism in the accreted terranes of the Sierra Nevada, as well as terranes of the Klamath Mountains (Fig. 15), followed each accretionary event, are slightly younger than the collisional events, and are spread over several different terranes (Irwin, 2003; Allen and Barnes, 2006; Barnes et al., 2006; Snoke and Barnes, 2006), so they are logically interpreted to represent slab failure magmatism (Hildebrand, 2013).

The plutonic rocks within the main mass of the Sierran batholith (Fig. 16) range in composition from gabbro to leucogranite, but the most common rock types are hornblende- and

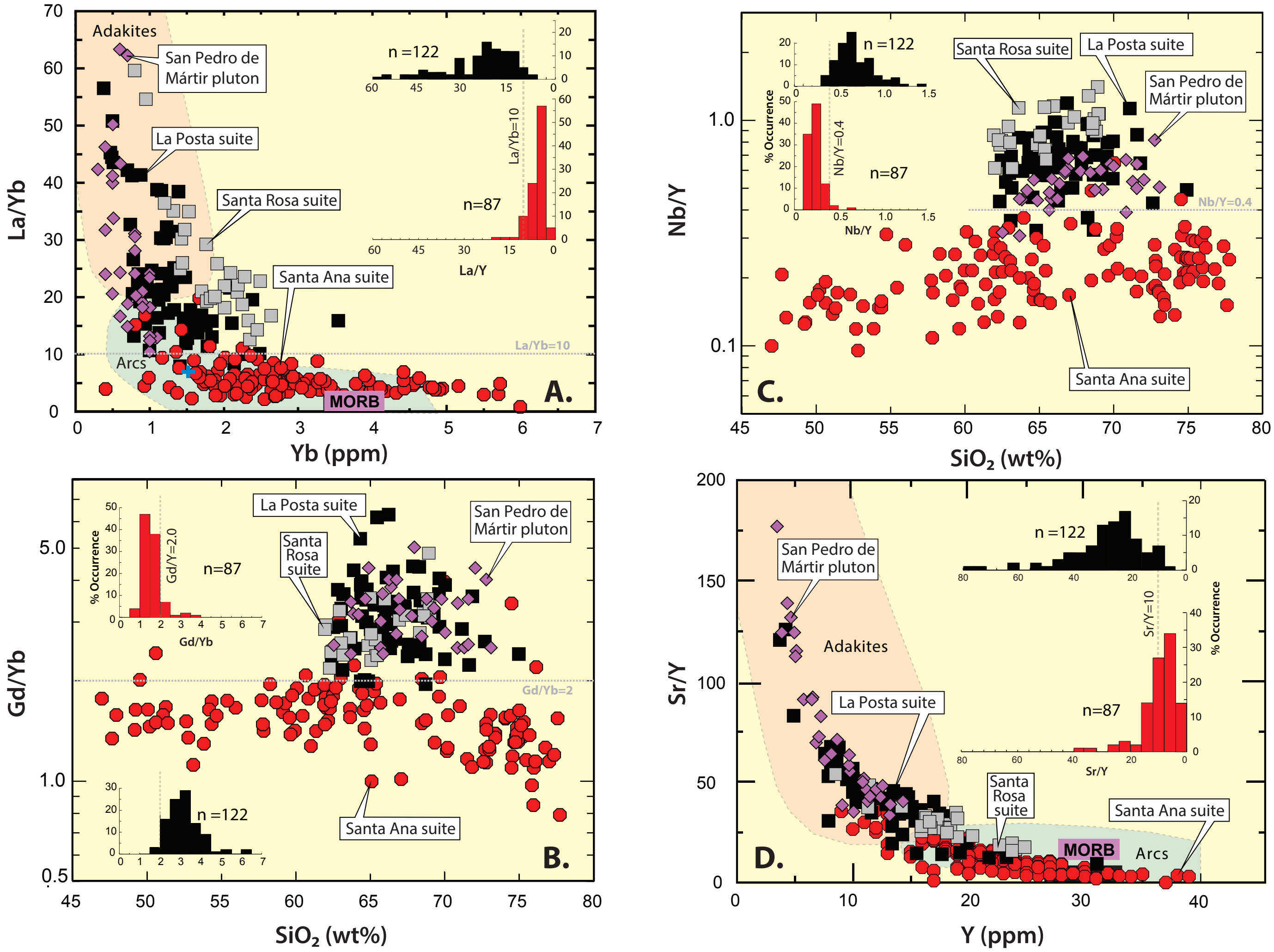


Figure 6. Some trace-element ratio plots for Peninsular Ranges plutonic suites. (A) La/Yb vs. Yb plot and La/Yb histograms; (B) Gd/Yb vs. SiO_2 and Gd/Yb histograms; (C) Nb/Y vs. SiO_2 plot and Nb/Y histograms; (D) Sr/Y vs. Y plot and Sr/Y histograms. In all inset histograms and *x-y* plots in A and D, only samples with >60% silica are plotted, whereas in B and C *x-y* plots, all samples are plotted. Fields are shown for adakite, arc andesite–dacite–rhyolite, and mid-ocean-ridge basalt (MORB) from Richards and Kerrich (2007), as is the average medium-K arc andesite (blue cross in A). In the histograms, vertical lines best separate pre- and postdeformational granitoid rocks. Figure is modified from Hildebrand and Whalen (2014b).

biotite-bearing tonalite, quartz diorite, and granodiorite, accompanied by lesser amounts of 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 one another or are separated by generally concordant screens of older metamorphic rock (Bateman, 1992; Bartley et al., 2012).

Just as Larsen (1948) recognized that gabbros were confined to the western part of the Peninsular Ranges batholith, Moore (1959) realized that the more mafic plutons within the Sierran batholith lay west of more intermediate-composition bodies. Many workers have since confirmed that the plutons of the Sierra Nevada can be divided into older western and younger eastern parts on the basis of geochemistry, magnetic susceptibility, age, and radiometric and stable isotopes (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 100–98 Ma, and they are deformed (Peck, 1980; Nokleberg and Kistler, 1980; Bateman et al., 1983; Saleeby and Busby-Spera, 1986; Memeti et al., 2010a; Wood, 1997; Saleeby et al., 1990), as are plutons such as the ca. 100 Ma Tehachapi intrusive suite (Wood, 1997). The bulk of this deformation apparently occurred prior to the emplacement of 98–83 Ma plutons of the Sierra Crest magmatic event (Coleman and Glazner, 1998; Davis et al., 2012) and, on evidence from metamorphic studies, prior

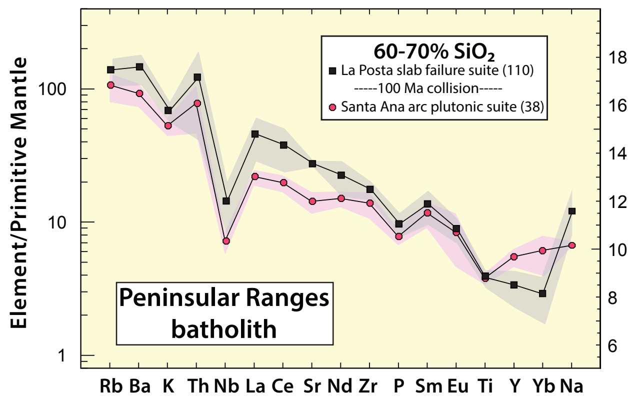


Figure 7. Primitive mantle–normalized extended element plots for plutonic rocks of Peninsular Ranges batholith showing compositional averages and ranges for 60–70 wt% SiO_2 plutonic rocks older and younger than 100 Ma. Primitive mantle–normalizing values are from Sun and McDonough (1989).

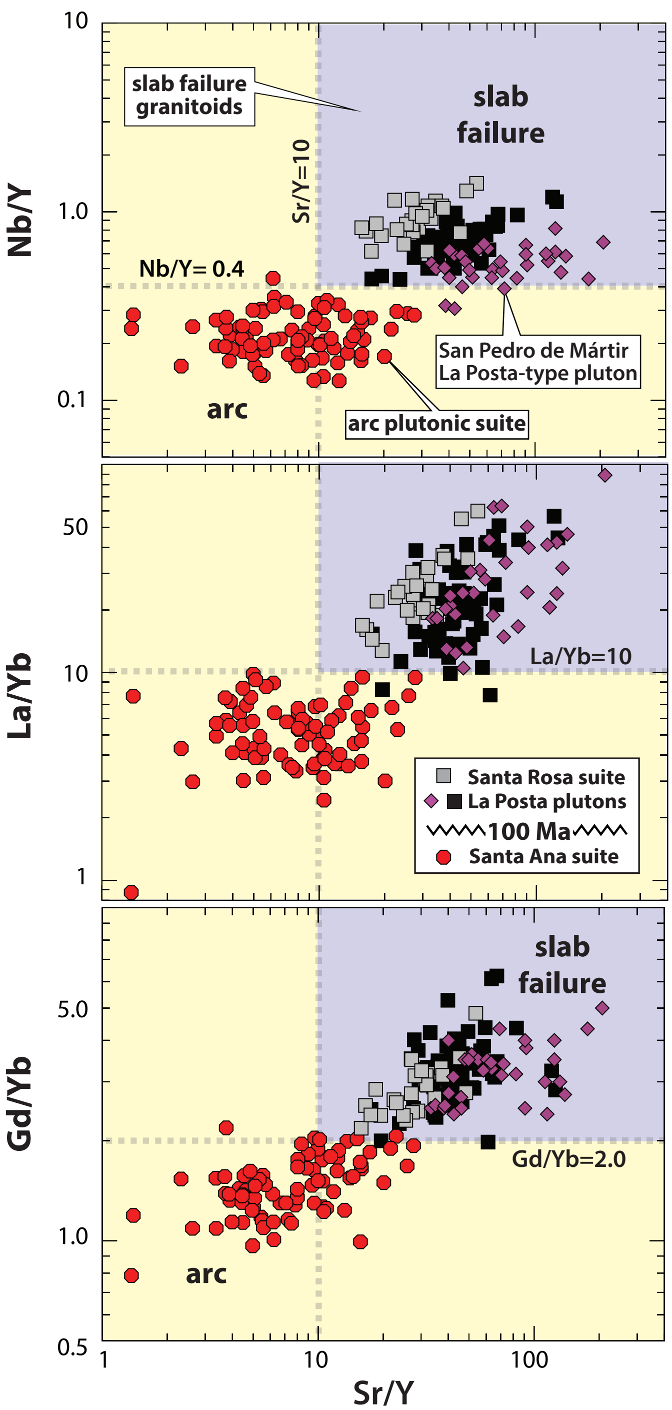


Figure 8. Samples with SiO_2 >60% from the Peninsular Ranges batholith plotted on Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y diagrams. Dashed lines are ratio values obtained from histograms in Figure 6 that separate most Santa Ana arc-type plutonic samples (pre–100 Ma) from plutonic rocks younger than 100 Ma of the La Posta–Santa Rosa suites (modified from Hildebrand and Whalen, 2014b). Analyses from the younger than 100 Ma Sierra San Pedro Mártir pluton, a La Posta–type plutonic complex in Baja California, are also plotted using data from Gastil et al. (2014). Rocks older than 100 Ma are interpreted to represent arc magmatism, whereas rocks younger than 100 Ma postdate a major deformational event and are best interpreted as slab failure rocks.

to 95 Ma in the southernmost Sierran batholith (Saleeby et al., 2007, 2008).

Similar to the Peninsular Ranges batholith, the western sector of the Sierra Nevada batholith contains a 125–100 Ma arc (Bateman, 1992) constructed upon crust assembled mainly during the Jurassic, whereas the eastern half contains large TTG complexes, such as the Sonora, Tuolumne, Whitney, John Muir, and Domelands (Fig. 16), emplaced during the 98–83 Ma Sierran Crest magmatic event (Coleman and Glazner, 1998; Saleeby et al., 2008). Also, like the Peninsular Ranges batholith, rocks of the Sierra Nevada contain evidence of a ca. 100 Ma deformational event (Fig. 17) that postdated all known sedimentary and volcanic wall rocks within the western arc (Peck, 1980; Nokleberg and Kistler, 1980; Bateman et al., 1983; Saleeby et al., 1990; Bateman, 1992; Wood, 1997; Memeti et al., 2010a; Hildebrand, 2013) yet predated, or was partly coeval with, the compositionally zoned plutonic complexes of the Sierran Crest magmatic event (Greene and Schweickert, 1995; Coleman and Glazner, 1998; Davis et al., 2012). The plutonic complexes were emplaced more or less simultaneously with development of mylonitic shear zones, a rapid increase in cooling rates between ca. 90 and 87 Ma, and increased sedimentation in the basin to the west during the Turonian (Mansfield, 1979; Renne et al., 1993; Tobisch et al., 1995). Thus, by analogy with the Peninsular Ranges batholith, we suggested (Hildebrand and Whalen, 2014b) that the western half of the Sierran batholith was an arc generated by westward subduction, and that the arc collided at ca. 100 Ma with a west-facing platform terrace on the eastern margin of the Bisbee-Arperos seaway. In our model, slab failure of the partially subducted eastern block led to voluminous magmatism of the Sierran Crest magmatic event.

Analyses from three post–100 Ma intrusive complexes were chosen to represent the postdeformational suite as they have relatively modern geochemistry and Sr and Nd isotopic data, and are

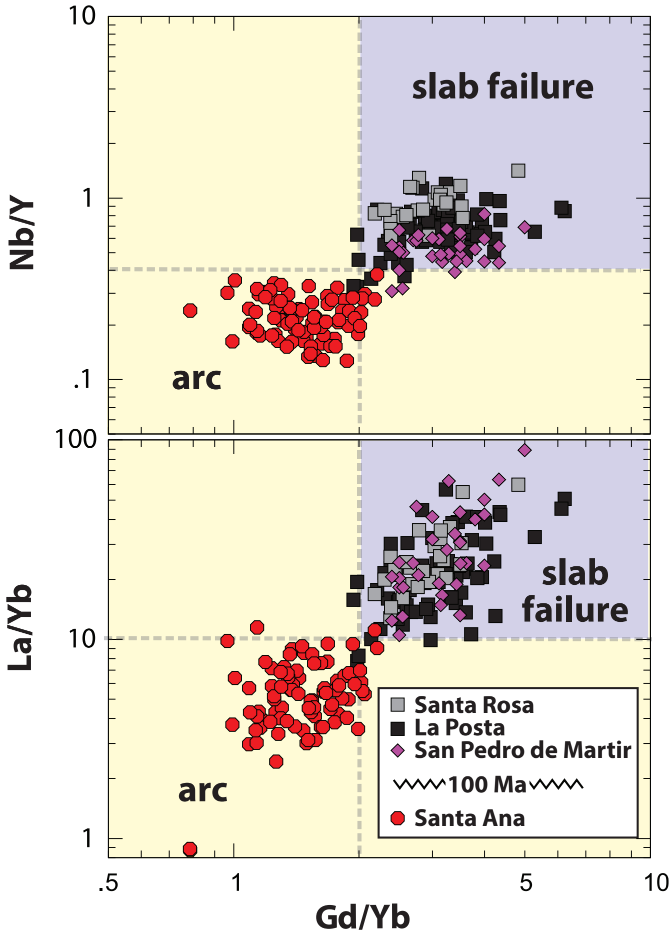


Figure 9. Plutonic samples with SiO_2 >60% from the Peninsular Ranges batholith plotted on Nb/Y and La/Yb vs. Gd/Yb diagrams. Although high Sr/Y is a common characteristic of slab failure magmas, we find Sr to be especially subject to alteration, and so we prefer these two plots to discriminate between suites. Data sources are as in previous plots.

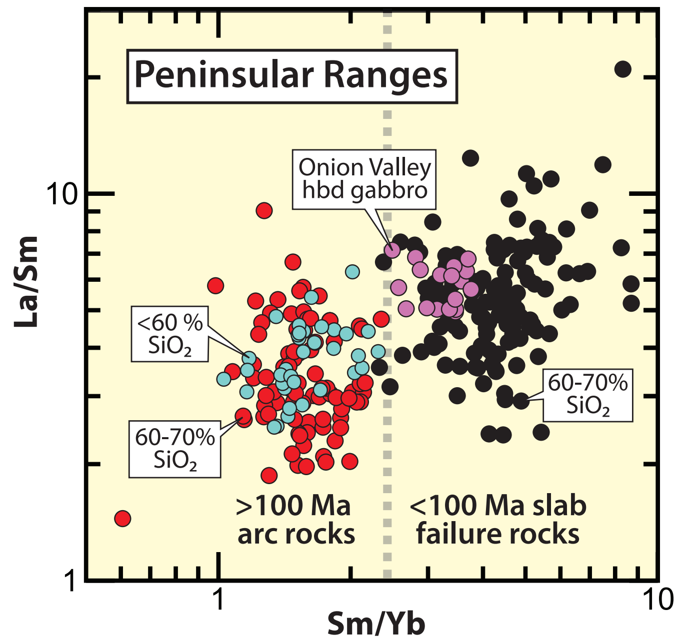


Figure 10. Sm/Yb ratios are one measure of partial melting depth in the mantle (Putirka, 1999). Rocks from both pre– and post–100 Ma suites from the Peninsular Ranges batholith are plotted in Sm/Yb vs. La/Sm space. Because there are no mafic rocks in the post–100 Ma La Posta suite, we plotted Sierran gabbros from the 92 Ma Onion Valley hornblende (hbd) gabbro (Sisson et al., 1996). Rocks older than 100 Ma have Sm/Yb values <2.5, whereas younger rocks have Sm/Yb >2.5. The differences presumably reflect depth of melting of the original source magmas and thus whether garnet was stable in the source. The data plotted here suggest that Sm/Yb ratios are useful for discriminating slab failure rocks from arc rocks, and so we have drawn a boundary between the two suites at Sm/Yb = 2.5.

mapped in some detail. Analyses from the Tuolumne intrusive suite (Fig. 18), likely the most studied intrusive complex in the world (Bateman and Chappell, 1979; Kistler et al., 1986; Huber at al., 1989; Coleman et al., 2004; Coleman et al., 2005; Žák and Paterson, 2005, 2009; Matzel et al., 2006b; Žák et al., 2007, 2009; Burgess and Miller, 2008; Gray, 2003; Gray et al., 2008; Paterson et al., 2008; Solgadi and Sawyer, 2008; Paterson, 2009; Memeti, 2009; Memeti et al., 2010b); the Mount Whitney intrusive suite (Fig. 19), another well-studied example (Hirt, 2007) of a mainly granodioritic intrusive suite in the eastern Sierra Nevada; and the Sahwave suite, an analogous intrusive complex located in northwestern Nevada (Van Buer and Miller, 2010), were plotted on our standard discrimination plots (Figs. 20 and 21). For some plots, we augmented the data from the intrusive complexes with analyses of the 92 Ma Onion Valley hornblende gabbro (Sisson et al., 1996), the ca. 95 Ma Sentinel granodiorite (Fulmer and Kruijer, 2009), the 92 Ma Lamarck Granodiorite (Frost and Mahood, 1987; Coleman et al., 1995), and 103–80 Ma plutons from western Nevada (du Bray, 2007), which appear to be a continuation of the Sierran-Sahwave belt (Fig. 22). As an example of the pre–100 Ma suite, we used analyses and isotopic data from the ca. 120 Ma Stokes Mountain complex (Clemens-Knott, 1992; Clemens-Knott and Saleeby, 1999).

Rocks of the Sierra Nevada batholith mimic those of the Peninsular Ranges batholith, not only in terms of age and geochemistry, but also in terms of Sr and Nd isotopes (Fig. 23). These criteria led us to argue that the post–100 Ma plutons in the Sierra Nevada were generated by slab failure similar to the post–100 Ma plutons of the Peninsular Ranges batholith. We turn now to other data from the Sierra Nevada (unavailable for the Peninsular Ranges), which strengthen our model of Oregonian slab failure involving a westward-dipping slab.

Ultramafic xenoliths collected from much younger volcanic rocks that erupted within the Sierra Nevada suggest that the xenoliths are dominantly residual cumulates remaining after extraction of partial melts from both upper mantle and subcontinental lithosphere at pressures of at least 32–18 kbar (Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1998; Chin et al., 2012),

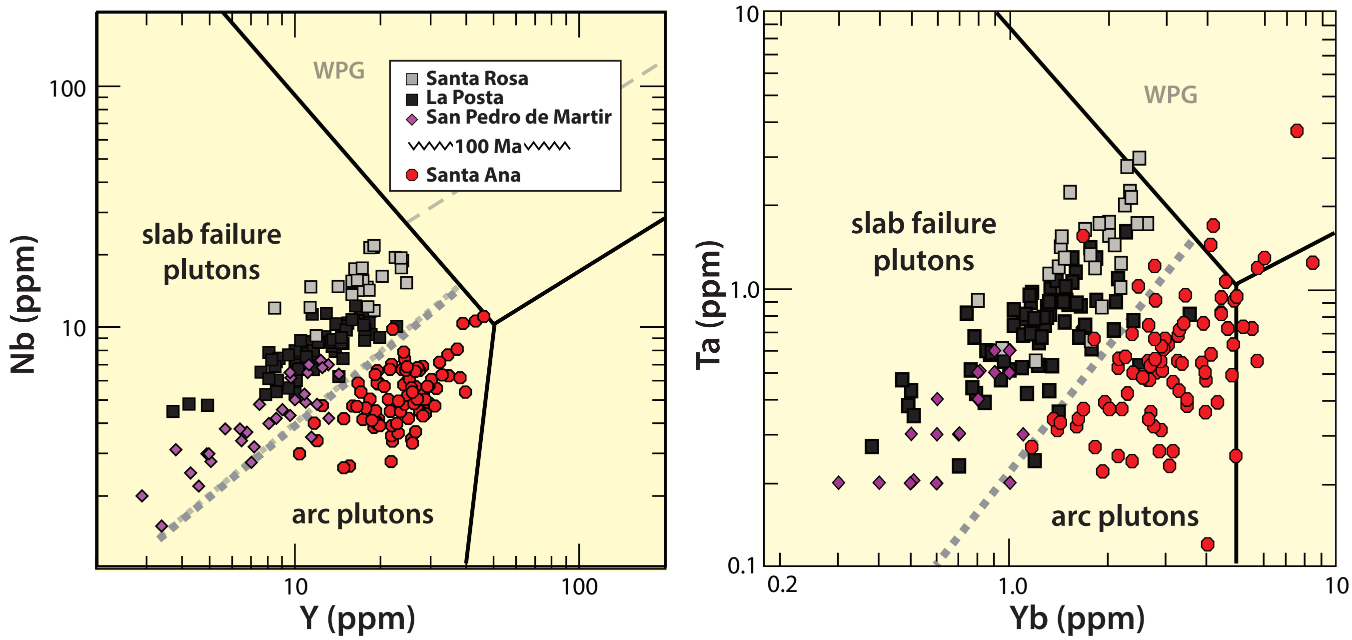


Figure 11. Plutonic samples with SiO_2 >60% from the Peninsular Ranges batholith plotted on Nb vs. Y and Ta vs. Yb discrimination diagrams, modified from Pearce et al. (1984) by addition of fields for slab failure and arc plutons based empirically on samples from the Peninsular Ranges batholith. WPG—within-plate granite.

consistent with upwelling mantle and adiabatic melting resulting from slab failure. Ducea and Saleeby (1998) showed that the cumulate rocks are broadly the same age as post–100 Ma Sierran granitoid magmatism and have similar Nd and Sr isotopic ratios.

Our concept of upwelling mantle and adiabatic melting was corroborated by two other detailed studies of peridotite xenoliths (Fig. 24), the first by Lee et al. (2001a, 2001b), who found two types of xenoliths with contrasting thermal histories and concluded that hot asthenosphere and cold Proterozoic lithospheric mantle "were suddenly juxtaposed, a feature consistent with the aftermath of rapid lithospheric removal or sudden intrusion of asthenospheric mantle into the lithosphere" during the Mesozoic. The second, by Chin et al. (2012), showed that peridotite underwent shallow melt depletion at 1–2 GPa and was refertilized at ~3 GPa, which could readily be accomplished by taking partially subducted, depleted lithosphere and then refertilizing it from upwelling asthenosphere.

Our model for westward-dipping subduction beneath the arc also resolves the difficulties of Chin et al. (2013) when trying to interpret granulitic quartzite xenoliths (Fig. 25) from a Miocene diatreme in the central Sierra Nevada. They were baffled because their extensive data from the quartzite xenoliths—T = 700–800 °C and P = 7–10 kbar, a ca. 103 Ma mean metamorphic age from U-Pb analyses in zircon rims, Proterozoic and Archean U-Pb crystallization ages for the cores of detrital zircon grains, and Hf isotopic ratios like those from Proterozoic basement of the eastern Sierra Nevada—appeared to indicate that rocks of what they thought to be the North American passive margin were transported deep beneath the arc and were metamorphosed at ca. 100 Ma; yet, they had no viable mechanism for getting them there in an eastward-directed subduction model because rocks of that age and type are unknown to the west. Alternatively, a westward-dipping subduction model could readily explain the presence of the metasedimentary xenoliths, as in that model, they could have been derived from basement to the west-facing platform margin of the Bisbee-Arperos seaway. Thus, the quartzite and peridotite xenoliths, as well as their features and histories—so difficult to explain in static eastward subduction models—are readily accounted for in an actualistic westward subduction–collision–slab failure model.

The age of deformation in the Sierran batholith is within error of that in the Peninsular Ranges (Memeti et al., 2010a; Chin et al., 2013), and plutons from the Sierra Nevada are similar in age, composition, and Nd-Sr isotopes to those of the Peninsular Ranges. In fact, individual plutons in both the Sierra and Peninsular Ranges batholiths are so similar in terms of geochemistry (Fig. 26) that, even in the cases of what traditionally were interpreted to be concentrically zoned complexes such as the La Posta, the San Pedro de Martir, the Tuolumne, and the Mount Whitney, the individual bodies cannot be related to one another by any fractional crystallization scheme, although some workers tried to do so (Bateman and Chappell, 1979; Hirt, 2007). Instead, they must be individual, but remarkably similar, batches of magma that were emplaced and then evolved more or less separately (Kistler et al., 1986; Memeti et al., 2014). We argued (Hildebrand, 2013; Hildebrand and Whalen, 2014b) that the plutons

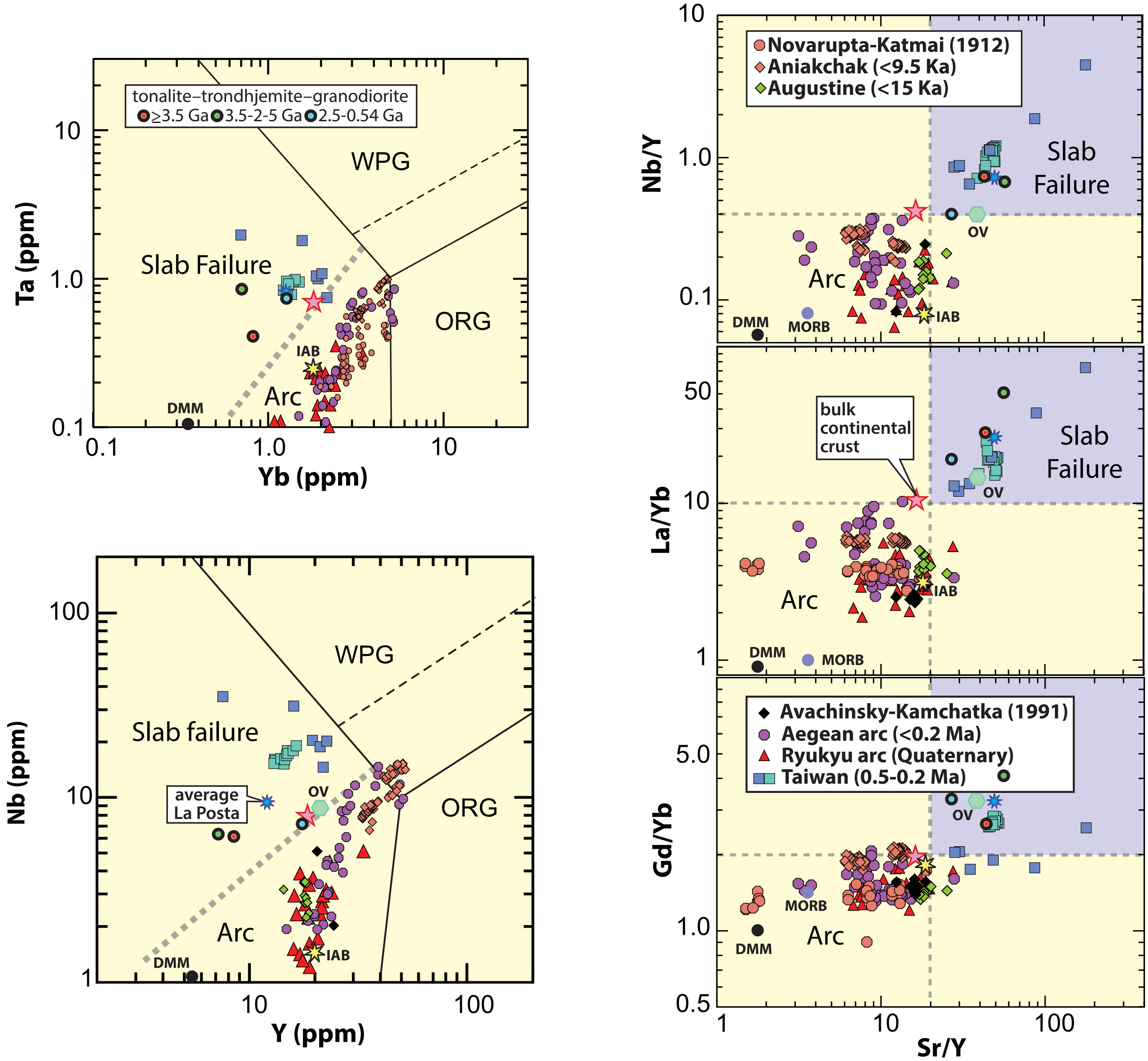


Figure 12. In order to test our discrimination plots developed from rocks of the Peninsular Ranges batholith, we plotted trace elements from young arc and slab failure settings on Nb vs. Y and Ta vs. Yb and Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y discrimination diagrams for: (A) Novarupta-Katmai from Hildreth and Fierstein (2012); Aniakchak from Bacon et al. (2014b); Augustine from Johnson et al. (1996); Avachinsky from Viccaro et al. (2012); Aegean arc from Bailey et al. (2009); Ryukyu from Shinjo et al. (2000); postcollisional lavas (Tsaolingshan and Kuanyinshan) of northernmost Taiwan from Wang et al. (2004); average compositions of different-aged tonalite-trondhjemite-granodiorite (TTG) suites (Martin et al., 2005); average compositions for the La Posta suite from Hildebrand and Whalen (2014b); mid-ocean-ridge basalt (MORB), depleted mantle (DMM), and island-arc basalt (IAB) from Kovalenko et al. (2010), and Onion Valley (OV) hornblende gabbro from Sisson et al. (1996). The orange and red stars indicate values for bulk continental crust from Rudnick and Gao (2003). WPG—within-plate granite; ORG—ocean-ridge granite.

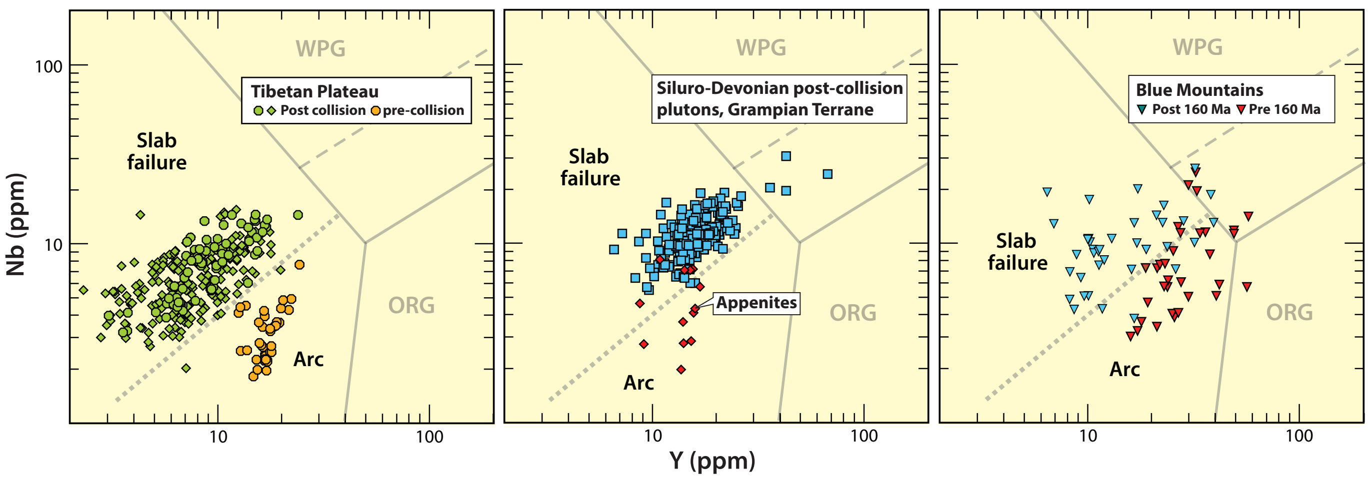


Figure 13. Nb vs. Y discrimination diagram for three arc and slab failure plutonic suites: (A) Tibetan Plateau pre- and postcollisional igneous rocks (Wang et al., 2008; Xu et al., 2010; Zeng et al., 2011; Zhao et al., 2009; Zhu et al., 2008; Wen, 2007; Zhang et al., 2014a, 2014b; Li et al., 2011); (B) postcollisional plutonic rocks from the Silurian–Devonian Grampian terrane, Caledonian orogen (Neilson et al., 2009; Kokelaar, 2015, personal commun.); and (C) pre– and post–160 Ma plutonic rocks from the Blue Mountains superterrane (Schwartz et al., 2011, 2014). WPG—within-plate granite; ORG—ocean-ridge granite.

in both areas are folded and were originally sheet-like in overall form.

One major difference between the Sierran and Peninsular Ranges batholiths is the presence of numerous small mafic bodies associated with the post–100 Ma Sierran Crest suite of plutons (Coleman et al., 1992; Coleman and Glazner, 1998; Sisson et al., 1996), which appear to be entirely absent in the post–100 Ma La Posta suite of the Peninsular Ranges (Kimbrough et al., 2014b). Sisson et al. (1996) suggested that magma that formed the hornblende gabbro sill complex at Onion Valley was able to ascend to high crustal levels because it contained 4%–6% H_2O, whereas drier basaltic magmas, likely abundant at depth, were unable to ascend to the same crustal level. If so, any mafic magmas associated with the La Posta suite might have been much drier. However, if such bodies exist there, they must be in the lowermost crust, as some exposed plutons of the La Posta suite have paleopressures as high as 5 kbar (Ague and Brimhall, 1988; Todd et al., 2003), and gabbros younger than 100 Ma are unknown there.

The overall similarity of rocks within the Sierran and Peninsular Ranges batholiths suggests to us, as it has to others (e.g., Ducea, 2001; Barbeau et al., 2005; Chapman et al., 2014), that they once formed part of a more extensive belt, but their precise original relationship and relative locations to one another are unclear, as they are now separated by the complex and incompletely resolved extensional/strike-slip Mojave block. Nevertheless, it appears reasonable, based on all the data, that subduction during the 130–100 Ma period was westward beneath the Sierra Nevada, just as it was for the Peninsular Ranges batholith. This model countermands the deeply entrenched hypothesis that subduction was eastward beneath the Sierran batholith and North American cratonic basement since at least the Early Jurassic.

The initial concepts for easterly subduction were largely based on the asymmetry of magnetic anomalies within the eastern Pacific Basin, which implied that North America had overridden several thousand kilometers of the Pacific plate (Hamilton, 1969a); the presence of rocks of the Franciscan complex, which could readily be interpreted to represent an ancient accretionary prism (Blake, 1984); and Jurassic–Cretaceous sedimentary rocks of the Great Valley Group (Ojakangas, 1968), interpreted as rocks deposited within a forearc basin (Dickinson, 1971). As these rocks lay west of the Sierra Nevadan batholith, the westerly progressing triad of batholith, forearc basin, and accretionary prism provided a sensible model to early workers (Burchfiel and Davis, 1975; Dickinson, 1981) and has come down to the present as the orthodox view: a paradigm of Cordilleran geology.

However, the model is not as robust as commonly portrayed, for there are problems that remain unresolved. First, within the Sierra Nevada, rocks older than 100 Ma were strongly deformed into westerly vergent folds and thrusts at ca. 100 Ma (Bateman et al., 1983; Wood, 1997; Hildebrand, 2013), whereas there appears to have been no commensurate deformation of Lower Cretaceous rocks within the Great Valley Group at that time (Constenius et al., 2000). Given that pre–100 Ma rocks of the Great Valley Group are not anywhere in contact with rocks of the Sierra Nevada, it appears possible, if not likely given the differences in degree of deformation, that the two elements were not adjacent to one another at, and before, 100 Ma.

Second, Wright and Wyld (2007) studied detrital zircon suites from rocks of the Great Valley Group and concluded that they were not derived from rocks exposed in the Sierra Nevada

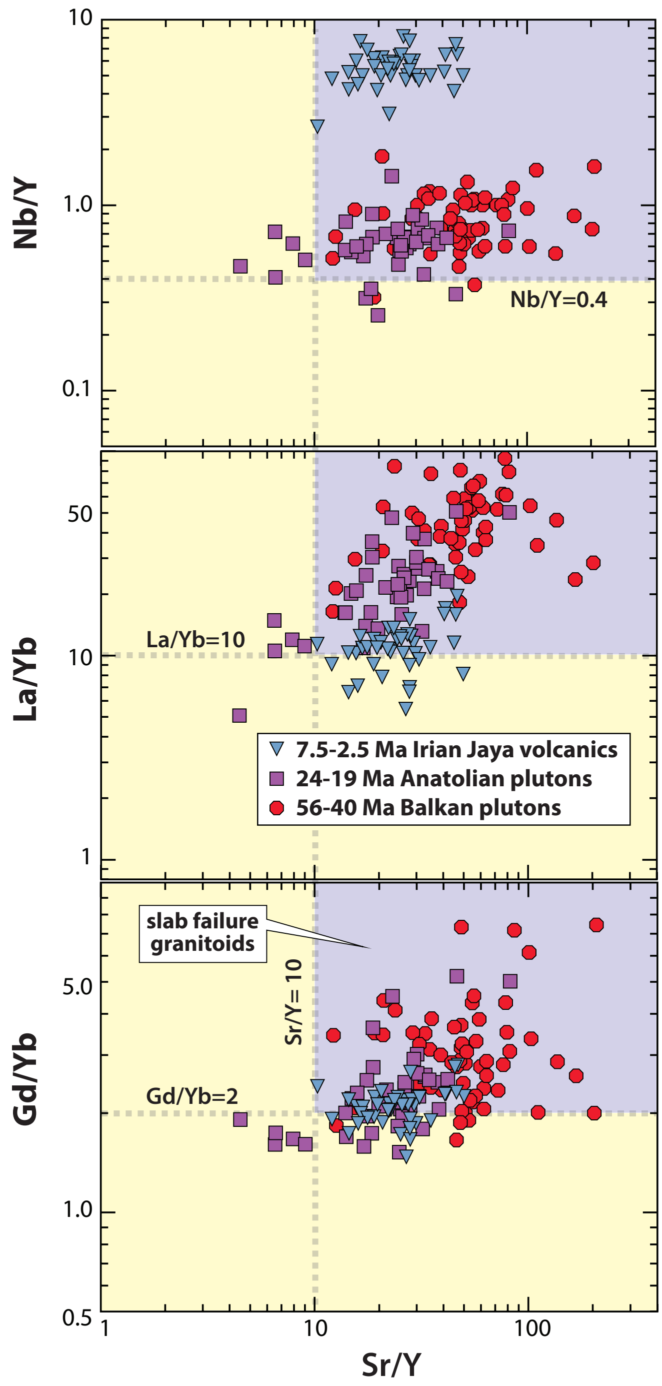


Figure 14. Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y discrimination diagrams for rocks from 7.5–2.5 Ma postcollisional magmatism in Irian Jaya from McMahon (2000, 2001); 24–19 Ma postcollisional plutonic rocks from western Anatolia, Turkey (Altunkaynak et al., 2012); and analyses from the 56–40 Ma postcollisional Rhodope Massif of southern Bulgaria–northern Greece (Marchev et al., 2013). All three suites are Cenozoic and are interpreted by the respective authors to represent slab failure magmatism.

as expected but instead were likely derived much farther south in the Oaxacan region of Mexico, as rocks there provide better sources for the observed zircon suites. Thus, if rocks of the Great Valley Group, and by association, the Franciscan complex, which, on the basis of paleomagnetic studies, also might have been far traveled (Champion et al., 1984), were not adjacent to the Sierran arc prior to 100 Ma, there may not have been either accretionary prism or forearc west of the Sierra Nevada during the 130–100 Ma period of arc magmatism, and so subduction might not have been eastward. We have already seen how a westward subduction model best explains the presence of high-grade quartzite xenoliths beneath the Sierra Nevada (Chin et al., 2013), so we now turn to the east and see what evidence exists for westerly subduction.

If our model for the Peninsular Ranges batholith can be applied to the Sierra Nevada as we suggest, then—despite the difficulties of tracing contacts through the multiply deformed and strongly extended Basin and Range Province to the east—we might locate the suture representing the original location of the Bisbee-Arperos seaway, although it could be difficult to find, because similar prerift rocks would exist on both sides of the basin. Nevertheless, the search might be narrowed by locating an easterly vergent thrust belt to the east of the Sierra Nevada–White-Inyo Mountains. Thrust faults, if they exist there, should be ca. 100 Ma, and not the 122–115 Ma thrusts of the Sevier event, nor the 82–75 Ma Laramide event, both of which would also have eastward vergence (Hildebrand, 2014).

Thrusts are well mapped and described in the area of the Spring Mountains, Nevada, and just to the south in eastern California (Burchfiel et al., 1974a, 1974b, 1998, 2010; Axen, 1984; Walker et al., 1995; Page et al., 2005). There, thrust faults are documented to be of at least two different ages. The well-known, and spectacularly exposed, Keystone thrust (Hewitt, 1931; Longwell, 1926) sits structurally atop a conglomerate that lies unconformably upon the structurally lower Wilson Cliffs–Red Spring–Contact thrust plate (Fig. 27). Clasts within the conglomerate are unlikely to have been derived from the Keystone plate, although they are similar to rocks within the thrust sheet, because cross-beds indicate that it was derived from the east, not the west, where the Keystone plate occurs (Burchfiel et al., 1998). This, and other evidence for erosion of the Wilson Cliffs–Red Spring–Contact thrust plate, along with balanced cross sections and geometric relations that show the Keystone thrust cutting and carrying the older thrust plates, different deformational styles, and an intervening period of high-angle faulting, led workers (Longwell, 1926; Davis, 1973; Burchfiel et al., 1998) to conclude that there were two periods of thrusting and that the Keystone thrust is younger and "out-of-sequence."

Another conglomerate, known as the conglomerate of Brownstone Basin (Fig. 27), sits structurally beneath the Red Spring thrust and contains cobbles and pebbles apparently derived from the Wheeler Pass thrust plate to the west (Axen, 1987) and detrital zircons as young as 103–102 Ma (Wells, 2016). The Wheeler Pass thrust sheet itself (Fig. 27), where exposed in

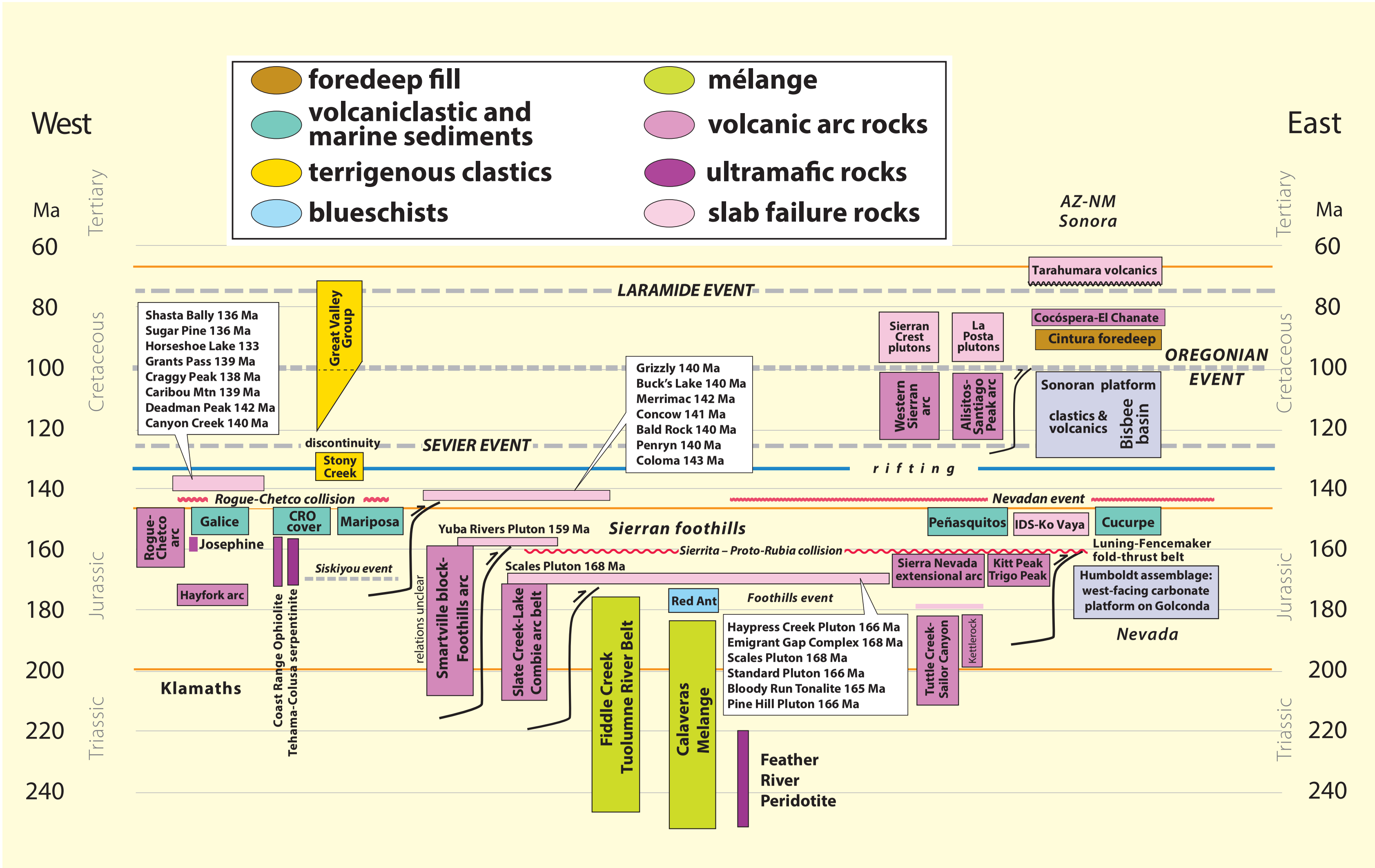


Figure 15. Geological time line for Sierra Nevada and environs, illustrating the occurrence of several postcollisional slab failure suites emplaced during the Jurassic and Cretaceous. The Jurassic suites are described in Hildebrand (2013), whereas the Arizona–New Mexico (AZ-NM)–Sonoran rocks are discussed in Hildebrand and Whalen (2014b). Figure is based in part on figures in Dickinson (2008). CRO—Coast Range ophiolite; IDS—Independence dike swarm.

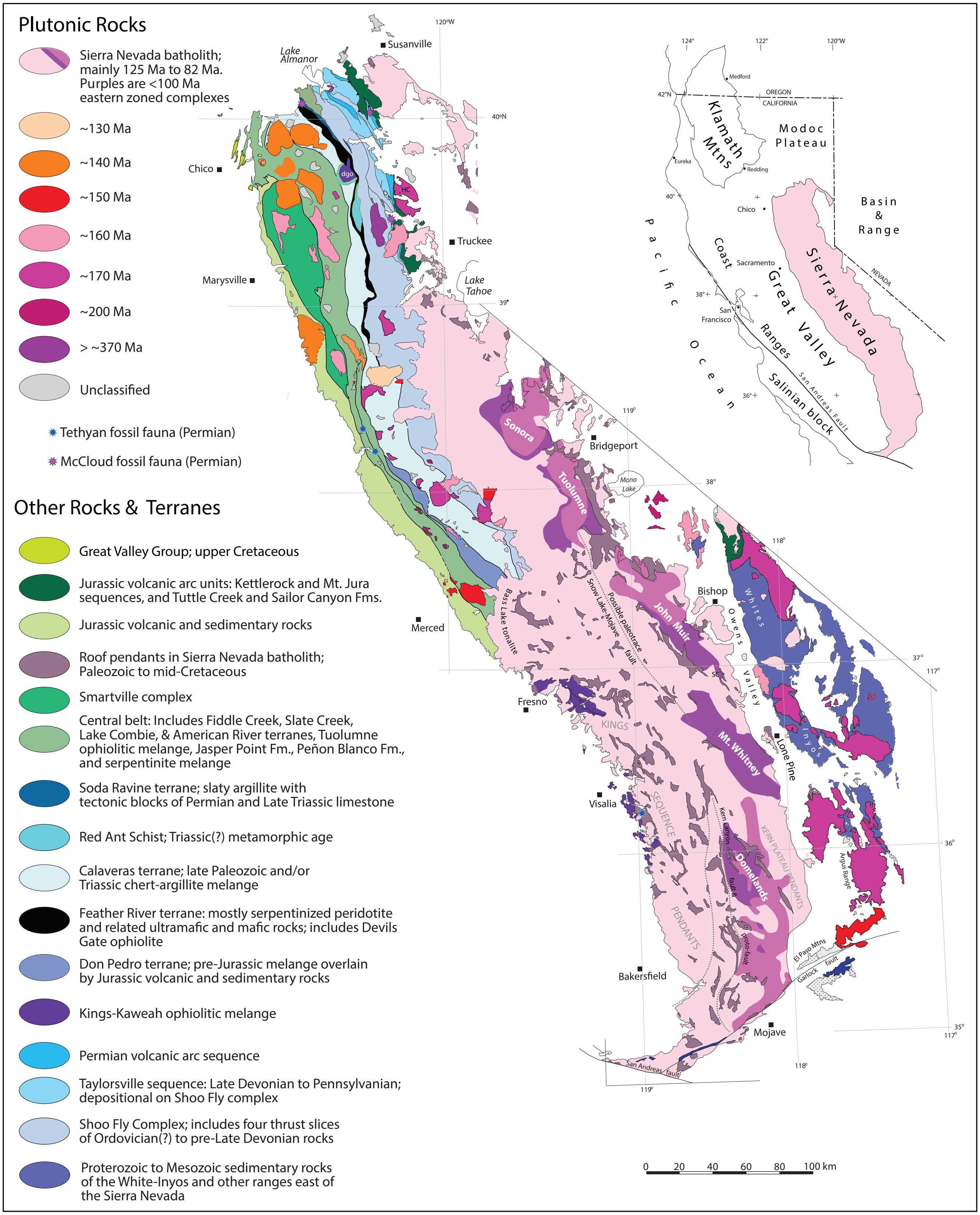


Figure 16. Generalized geological map of the Sierra Nevada batholith and environs, showing the basement terranes and the distribution of major plutonic complexes of the post–100 Ma Sierran Crest magmatic suite (modified from Irwin and Wooden, 2001; with additional data from Dunne et al., 1978; Saleeby et al., 1978; Bateman, 1992; Saleeby and Busby-Spera, 1993). Distribution of several post–100 Ma plutonic complexes was modified from Van Buer and Miller (2010).

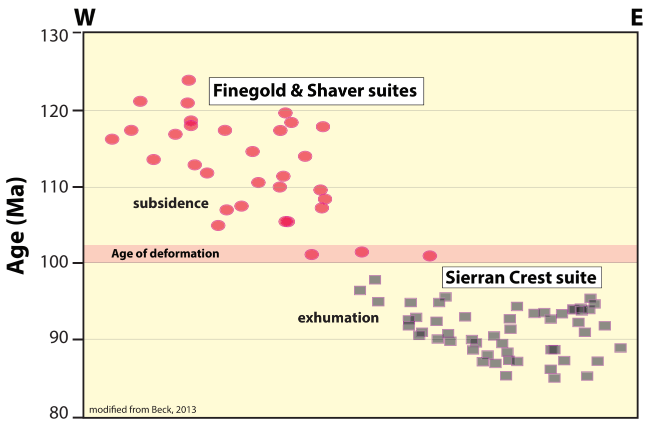


Figure 17. W-E section vs. age (Ma) showing U-Pb zircon ages for plutons in the central Sierra Nevada (from Beck, 2013) and their temporal relation to the ca. 100 Ma deformational event. Note the similar ages and relations as those in the Peninsular Ranges batholith in Figure 4.

the Spring Mountains, contains evidence for exhumation during the Late Jurassic (Giallorenzo, 2013), which perhaps reflects the Nevadan event; however, recent zircon (U-Th)/He thermochronology from the thrust sheet, where exposed in the Nopah Range (Fig. 27) to the southwest, shows that exhumation started at 100 Ma (Giallorenzo, 2013).

In the southern Spring Mountains, nonmarine sedimentary and volcaniclastic rocks of the Lavinia Wash sequence (Fig. 27), interpreted as synorogenic deposits by Carr (1980), lie structurally below the Contact thrust plate. A rhyolitic boulder in conglomerate of the Lavinia Wash sequence was dated as 98.0 Ma, and plagioclase within an ignimbrite in the sequence yielded a $^{40}Ar/^{39}Ar$ age of 99.0 ± 0.4 Ma (Fleck and Carr, 1990).

Just over the state line in the Mezcal Range of southeastern California, a sequence of 100.5 ± 2 Ma volcanic and volcaniclastic rocks known as the Delfonte volcanics (Fig. 27) was detached, folded, and transported eastward on thrust faults (Fleck et al., 1994; Walker et al., 1995) prior to the emplacement of high-Sr and high-Ba plutons of the Teutonia batholith (Fig. 22), which range in age from 97 to 92 Ma (Beckerman et al., 1982; Miller et al., 1996; DeWitt et al., 1984) and are likely part of the Oregonian slab failure suite.

To the southeast across the Ivanpah Valley, the New York Mountains contain highly strained metavolcanic rocks ranging in age from 98.4 to 97.6 Ma, whereas associated metasedimentary rocks of Sagamore Canyon (Fig. 27) have maximum depositional ages of 98 Ma (Wells, 2016). Thrust faults cut the volcanic rocks and are cut by ca. 90 Ma monzogranite of the Teutonia batholith (Wells, 2016).

In southern Nevada, northeast of Las Vegas (Fig. 27), the Upper Albian to Cenomanian Willow Tank Formation and Baseline Conglomerate, interpreted as synorogenic foreland deposits resting unconformably on Middle Jurassic Aztec sandstone in the Valley of Fire region, were dated as 98–96 Ma (Fleck, 1970; Bohannon, 1983; Bonde, 2008; Pape et al., 2011). More recent studies of detrital zircons from these and other local formations—as well as zircons from plutons and volcanic rocks—bracket deformation to be 102–96 Ma (Troyer et al., 2006; Bonde et al., 2012; Wells, 2016).

Farther north (Fig. 22), thrusts within the Garden Valley thrust system (Bartley and Gleason, 1990), part of the Central Nevada thrust system (Speed et al., 1988), are cut by the ca. 98 Ma Lincoln stock and the ca. 86 Ma Troy granite (Taylor et al., 2000). These data, plus the others cited above, are consistent with a major fold-and-thrust belt active at ca. 100 Ma.

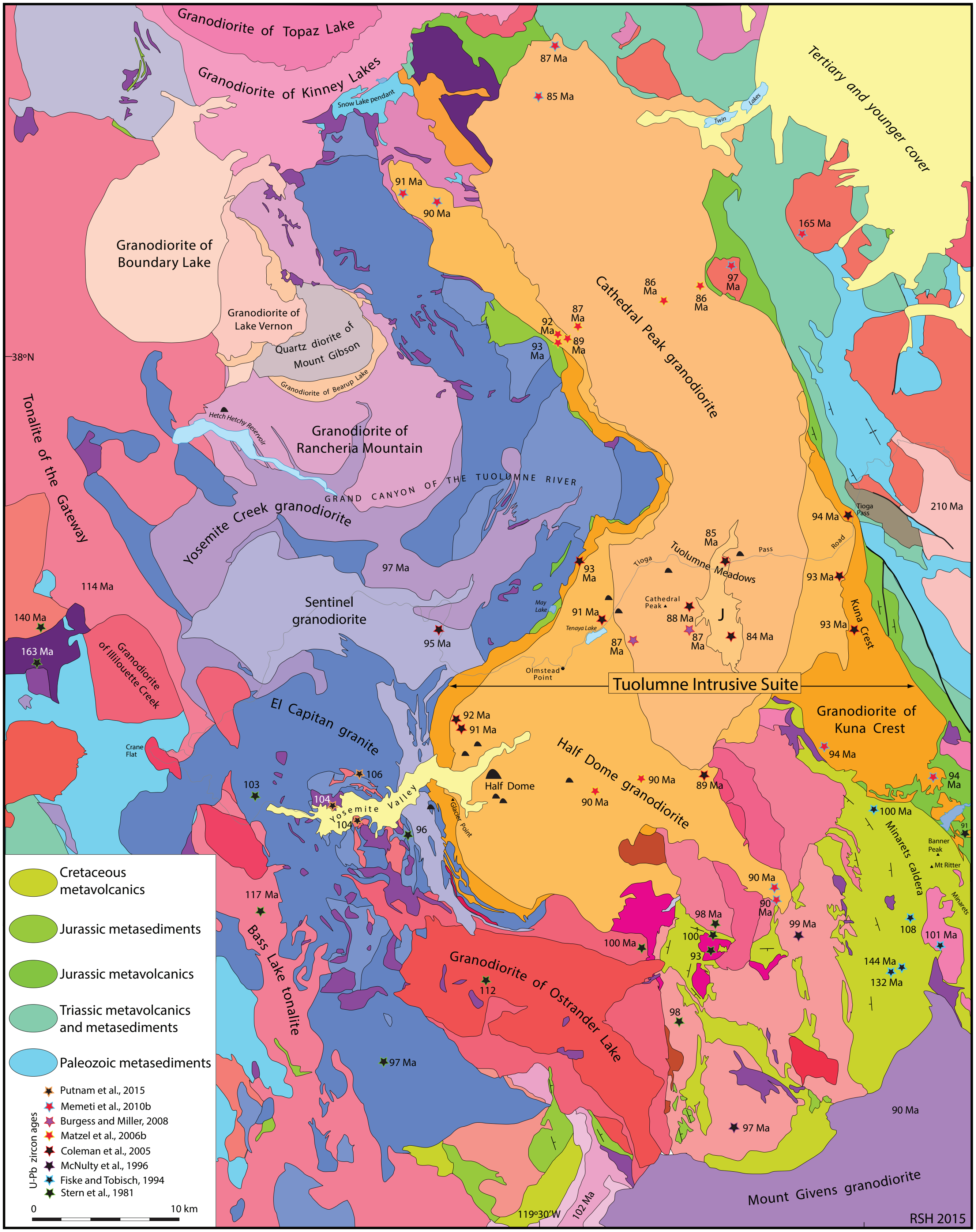
Overall, the structural and sedimentological data from the Great Basin area support the concept that a ca. 100 Ma suture lies to the east of the Sierra Nevada and Mojave Province, but its precise location is unknown. Because the basin developed from rifting, the rocks on both sides of any proposed suture were more or less the same: Truncations of older facies patterns are key features in locating the suture. We noted (Hildebrand and Whalen, 2014b) that rocks of Sonora, considered correlative with those in the Death Valley region and so commonly used as evidence for the inferred Mojave-Sonora megashear (Stewart, 2005), can be more easily explained by oblique opening and/or closing of the Bisbee-Arperos seaway.

Figure 22 shows one possible location for the suture through central and northern Nevada. It trends north-northwestward along the well-known southwesterly truncation of Lower Paleozoic platform, slope, and basinal facies of the Antler platform (Crafford, 2007, 2008) and then along the Golconda thrust, or other thrusts in the immediate area, such as the Clear Creek system, both of which are constrained to be younger than Late Triassic (Silberling, 1975; Dunston et al., 2001). This location places rocks of the polydeformed Nolan and Golconda domains of Crafford (2008) against the Paleozoic Antler platform (Hildebrand, 2014) and its structurally overlying Roberts Mountain allochthon. Other solutions are possible.

IDAHO-MONTANA

To the north, the next major batholithic package occurs within the Helena salient of Idaho and Montana (Fig. 28). There, the widespread plutonic rocks have long been considered to represent the products of arc magmatism (Hamilton, 1969a, 1969b; Dickinson, 1970; Burchfiel and Davis, 1972, 1975; Hyndman, 1983, 1984; Bateman, 1988; Hyndman et al., 1988; Foster and Fanning, 1997; Dickinson, 2004; Ducea and Barton, 2007; Lee et al., 2007). However, we note that the plutonic rocks (Figs. 28 and

Figure 18. Geology of the Tuolumne intrusive suite and vicinity, showing major mapped units and U-Pb zircon ages. Geology is modified from Huber et al. (1989) and Peck (1980). U-Pb ages are from Putnam et al. (2015); Memeti et al. (2010b); Burgess and Miller (2008); Matzel et al. (2006b); Coleman et al. (2005); McNulty et al. (1996); Fiske and Tobisch (1994); and Stern et al. (1981). J—Johnson granite porphyry.



Granodiorite of Topaz Lake
Granodiorite of Kinney Lakes
Snow Lake pendant
Twin Lakes
Tertiary and younger cover
Granodiorite of Boundary Lake
Granodiorite of Lake Vernon
Quartz diorite of Mount Gibson
Granodiorite of Bearup Lake
38°N
Hetch Hetchy Reservoir
Granodiorite of Rancheria Mountain
GRAND CANYON OF THE TUOLUMNE RIVER
Cathedral Peak granodiorite
Tonalite of the Gateway
Yosemite Creek granodiorite
Sentinel granodiorite
Granodiorite of Illilouette Creek
Tioga Pass Road
Tuolumne Meadows
Cathedral Peak
May Lake
Tenaya Lake
Olmstead Point
Kuna Crest
Tuolumne Intrusive Suite
Granodiorite of Kuna Crest
El Capitan granite
Half Dome granodiorite
Half Dome
Yosemite Valley
Glacier Point
Crane Flat
Minarets caldera
Banner Peak
Mt Ritter
Minarets
Bass Lake tonalite
Granodiorite of Ostrander Lake
Mount Givens granodiorite
119°30'W
210 Ma
165 Ma
163 Ma
140 Ma
114 Ma
117 Ma
102 Ma
97 Ma
95 Ma
90 Ma
RSH 2015
Cretaceous metavolcanics
Jurassic metasediments
Jurassic metavolcanics
Triassic metavolcanics and metasediments
Paleozoic metasediments
U-Pb zircon ages
Putnam et al., 2015
Memeti et al., 2010b
Burgess and Miller, 2008
Matzel et al., 2006b
Coleman et al., 2005
McNulty et al., 1996
Fiske and Tobisch, 1994
Stern et al., 1981
0
10 km

Figure 18.

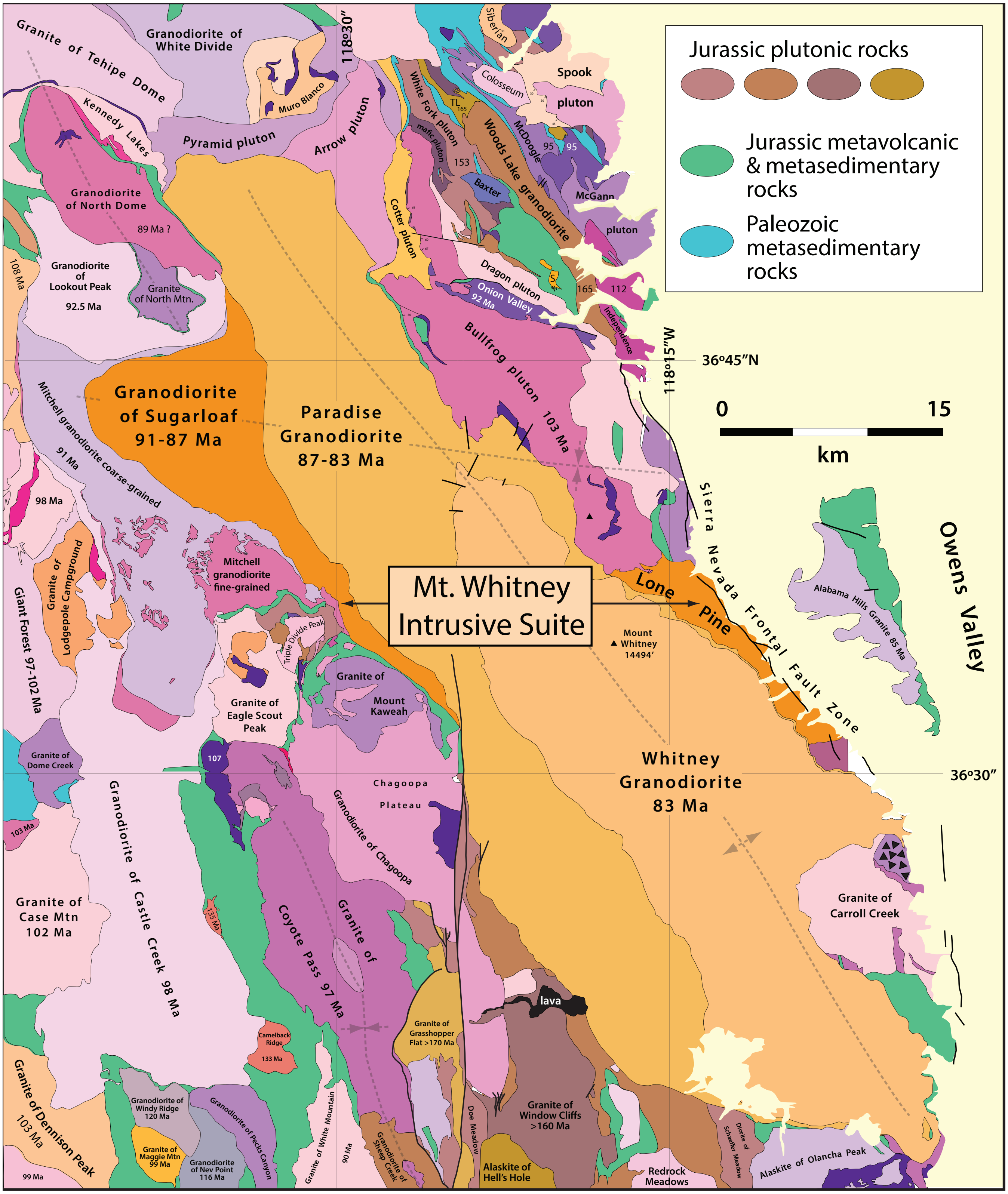


Figure 19. Geology and U-Pb ages of the Mount Whitney intrusive suite (Hirt, 2007), compiled from geological quadrangle maps by Moore (1963, 1978, 1981); Stone et al. (2000); du Bray and Moore (1985); and Moore and Sisson (1987). Purples, pinks, and oranges—Cretaceous plutons.

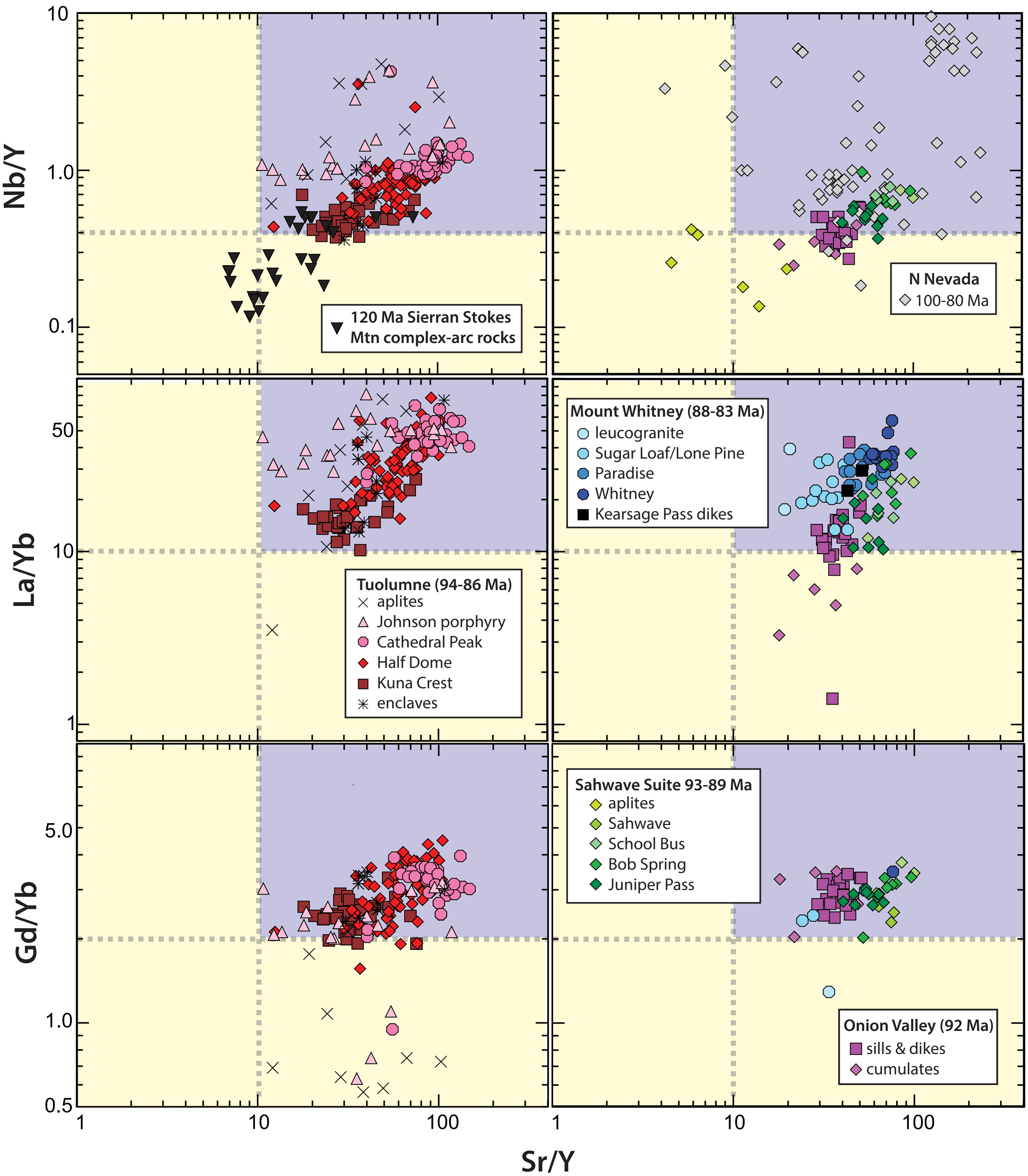


Figure 20. Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y discrimination diagrams for various Sierra Nevada and Nevadan plutonic suites, including precollisional 120 Ma Sierran Stokes Mountain complex arc rocks (Clemens-Knott, 1992), 100–84 Ma postcollisional Tuolumne intrusive suite (Memeti, 2009), Mount Whitney (Hirt, 2007), Sahwave intrusive suite of NW Nevada (Van Buer and Miller, 2010), and Onion Valley hornblende gabbro (Sisson et al., 1996) plus northern Nevada plutonic rocks (du Bray, 2007), illustrating that, except for aplites, late-stage leucogranites, and cumulates, the pre–100 Ma arc rocks of the Stokes Mountain complex and the remaining plutonic units (100–80 Ma) are well separated on our Nb/Y and Sr/Y discrimination diagrams.

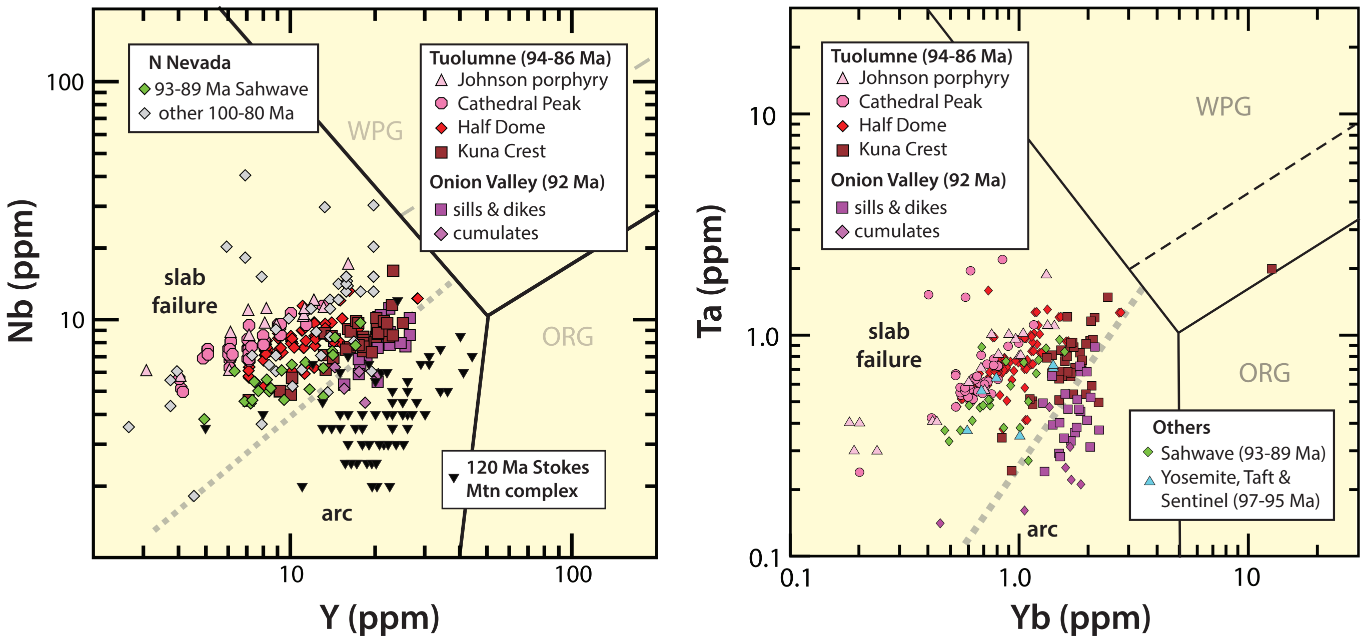


Figure 21. Nb vs. Y and Ta vs. Yb discrimination diagrams for various Sierra Nevada and NW Nevada plutonic suites. Data sources are as in Figure 20. WPG—within-plate granite; ORG—ocean-ridge granite.

29) were emplaced during three main stages, each temporally associated with a discrete period of deformation: (1) 98–87 Ma metaluminous plutons of the Idaho batholith emplaced during and after 100 Ma NE-vergent thrusting and metamorphism; (2) 83–53 Ma peraluminous and metaluminous magmatism of the Idaho, Boulder, and other batholiths emplaced during and after the Laramide event, which was another period of eastward thrusting; and (3) 51–44 Ma magmatism associated with extensional collapse of the Laramide hinterland (Robinson et al., 1968; Hyndman et al., 1988; House et al., 1997l Gaschnig et al., 2010). In this section, we describe and examine rocks of the first stage, which occur within and just west of the Atlanta lobe of the Idaho batholith (Fig. 29).

The western margin of the Idaho batholith more or less coincides with the trace of the Salmon River suture (Lund and Snee, 1988) and the overprinting and dextrally transpressive Western Idaho shear zone (McClelland et al., 2000; Giorgis et al., 2008; Fleck and Criss, 2004), which separate rocks of the Blue Mountains superterrane on the west from rocks of the Proterozoic Windermere and Belt supergroups to the east (Blake et al., 2009; Gray et al., 2012; LaMaskin et al., 2015). Overall, the geological relationships in the area are complex, metamorphic grades are high, and researchers argue about the significance and timing of various deformational events and their nomenclature. Nevertheless, just west of the main mass of the Idaho batholith, ca. 90 Ma foliated epidote-hornblende tonalitic, quartz dioritic, and granodioritic plutons, of which the Payette River tonalitic intrusion (Fig. 28) is the best studied example, are exposed (Lund and Snee, 1988; Manduca et al., 1993).

Along much of the western margin of the Atlanta lobe of the Idaho batholith, elongate plutons of hornblende-biotite tonalite dated at 87 Ma (Gaschnig et al., 2010) crop out, whereas within the main mass of the lobe and in the southeast corner of the Bitterroot lobe (Fig. 29), plutonic remnants of hornblende-biotite tonalite and porphyritic and nonporphyritic granodiorite range in age from 98 to 87 Ma (Lewis et al., 1987; Gaschnig et al., 2010). Just north of the Atlanta lobe, within and adjacent to the Coolwater culmination (Fig. 28), Lund et al. (2008) dated several bodies of orthogneiss to be 94–86 Ma. Over most of the area, the 98–87 Ma plutons and their wall rocks were deformed during the Laramide event and were intruded by younger plutons, some of which are themselves gneissic (Lund et al., 2008). The younger deformation created complexities that can lead to ambiguous or conflicting relations.

A possible mid-Cretaceous foredeep succession extends in broken fashion from the Flint Creek basin southward to the Snake River Plain (Fig. 28). There, a 3–4 km sequence of conglomerate, shale, porcellanite, and limestone, which ranges in age from Cenomanian to Santonian, is unique and does not have correlative rocks to the east in the Sevier-Laramide foredeep of the Western Interior Basin (Wallace et al., 1990).

We plotted trace-element data collected from the 98–87 Ma plutons and gneisses on our discrimination plots (Fig. 30), and they plot with other slab failure rocks from the Sierra and Peninsular Ranges and not in fields characterized by arc plutons. Thus, their age, geochemistry, and syn- to posttectonic emplacement suggest to us that the plutons were generated by Oregonian slab failure.

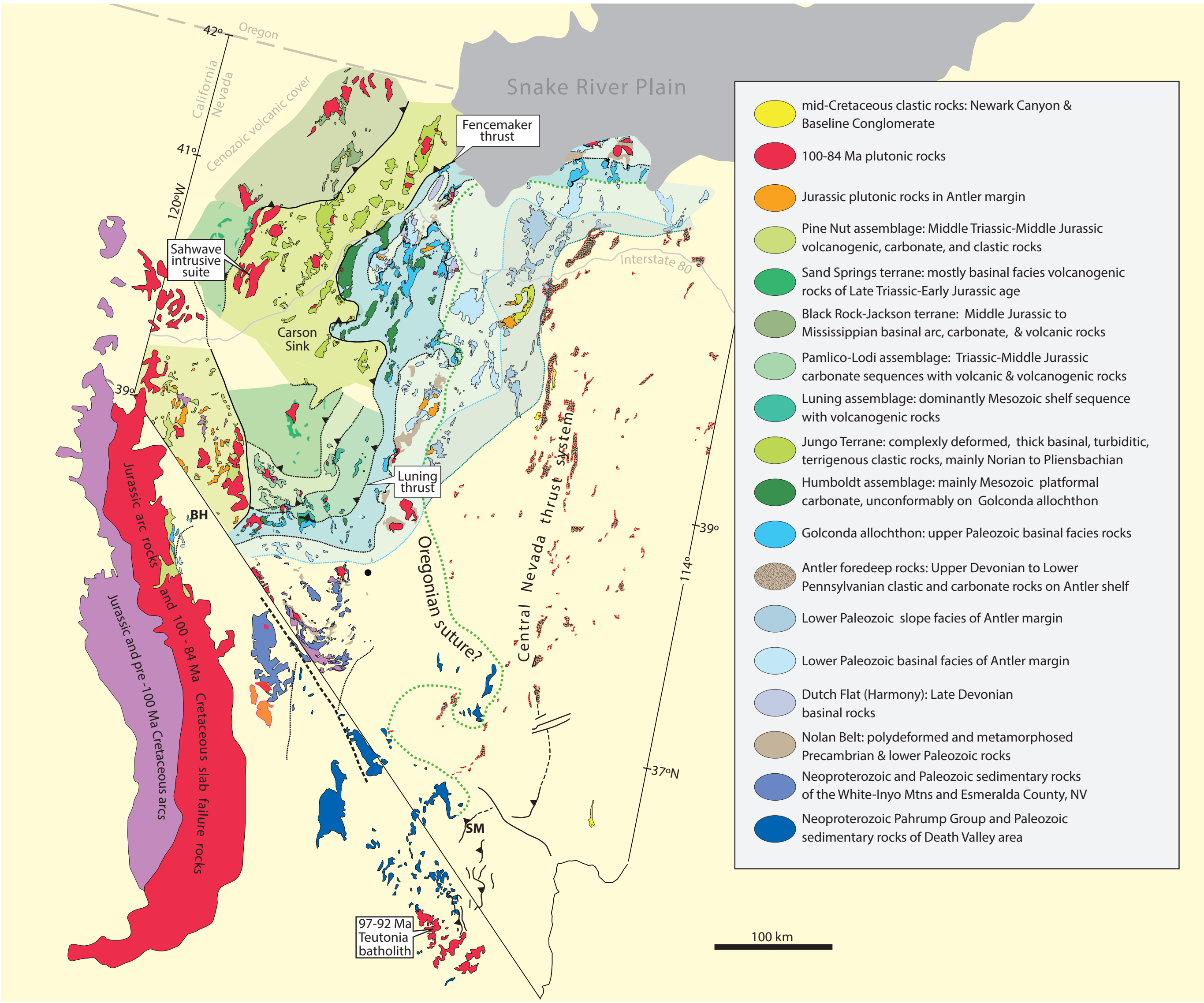


Figure 22. Geological map of NE California and Nevada showing the Sierra Nevada batholith, Cretaceous plutonic rocks of Nevada, various tectonic terranes of Nevada after Crafford (2007, 2008), and possible location of Oregonian suture. BH—Bodie Hills; SM—Spring Mountains; NV—Nevada.

CASCADES–COAST PLUTONIC COMPLEX

North of the Lewis and Clark line, there are rocks of the Coast-Cascade orogen (Gibson and Monger, 2014), which is the westernmost of the dual structural-metamorphic welts (Fig. 31) within the Canadian Cordillera (Monger et al., 1982). The orogen contains abundant plutonic rocks, which are collectively labeled as the Coast plutonic complex (Roddick, 1983; Mahoney et al., 2009) or the Coast Ranges batholith (Gehrels et al., 2009). Rocks of the U.S. Cascades and Coast plutonic complex of British Columbia were formerly continuous but are now separated by over 100 km along the dextral Fraser River–Straight Creek faults (Fig. 31) and other northerly and northwesterly trending faults (Umhoefer and Miller, 1996). Throughout the orogen, many plutons have been reasonably well dated (van der Heyden, 1989; Friedman and Armstrong, 1995; Gehrels et al., 2009; Miller et al., 2009), but compared to the Peninsular Ranges, Sierra, and Idaho batholiths, modern geochemical and isotopic data are sparse (Mahoney et al., 2009; Girardi et al., 2012).

Overall, the geology of the orogen is complex, but based on ages of structural features, plutonism, and metamorphism, it appears to contain the products of at least two Cretaceous collisional episodes: the ca. 100 Ma Oregonian event and the ca. 82 Ma Laramide event, and their precollisional rocks. The orogen is bisected by a steep to vertical fault known as the Coast shear zone (Crawford and Crawford, 1991). Although a few workers

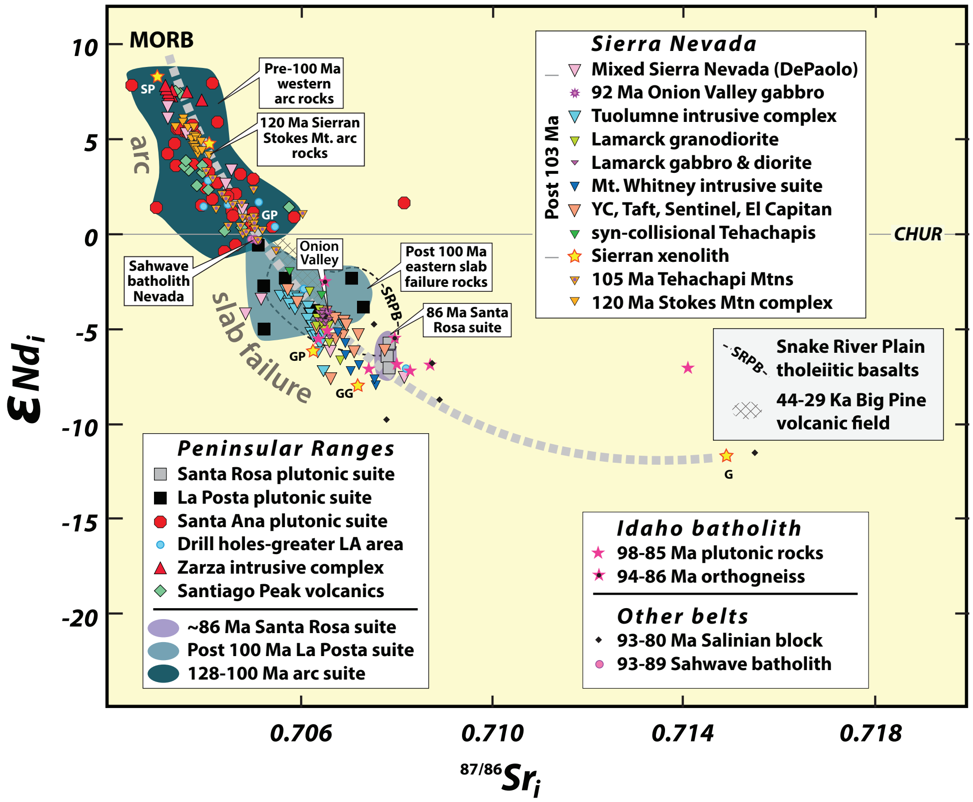


Figure 23. ε_{Nd} vs. $Sr_{initial}$ plot of various arc and slab failure plutonic and volcanic suites of the Peninsular Ranges and Sierra Nevada plus Idaho batholith and other belts and suites compared to some Cenozoic basalts of western North America, illustrating the isotopic differences between arc suites and slab failure suites and the isotopic similarities of the Oregonian slab failure suites with basalts from the Snake River Plain (SRPB) and Big Pine volcanic field. Peninsular Ranges samples from drill holes in greater Los Angeles (LA) area are from Premo et al. (2014); Zarza intrusive complex data, a Santa Ana arc pluton in Baja, are from Johnson et al. (2002); Santiago Peak volcanics are from Herzig and Kimbrough (2014); 120 Ma Stokes Mountain plutons of western Sierran arc are from Clemens Knott (1992); post–103 Ma plutons of eastern Sierra Nevada are from DePaolo (1981); Tuolumne intrusive suite is from Memeti (2009); 94 Ma Lamarck granodiorite is from Coleman et al. (1992); 88–83 Ma Mount Whitney intrusive suite is from Hirt (2007); 103–95 Ma Yosemite Creek (YC), Taft, Sentinel, and Taft data are from Ratajeski et al. (2001); Onion Valley, a 92 Ma hornblende gabbroic sill complex, is from Sisson et al. (1996); 93–89 Ma Sahwave intrusive suite, a post–100 Ma zoned complex located in western Nevada, with similar composition, zoning, and petrography to Sierran Crest plutons, is from Van Buer and Miller (2010); Tehachapi data are from Coleman and Glazner (1998); Salinian block is from Chapman et al. (2014); Snake River Plain is from Menzies et al. (1984); and Big Pine volcanic field is from Blondes et al. (2008). Sierran xenoliths: SP—spinel peridotite; GP—garnet peridotite; GG—garnet granulite; G—granulite. MORB—mid-ocean-ridge basalt; CHUR—chondritic uniform reservoir.

(Cowan et al., 1997; Hollister and Andronicos, 1997) have suggested that there was major dextral shear on the zone, most workers have argued for east-side-up reverse faulting (Brew and Ford, 1978; Klepeis et al., 1998; Rusmore et al., 2000; Saleeby, 2000; Crawford et al., 2005). Plutons east of the shear zone were emplaced into what is known as the Central gneiss complex and are much younger, temporally related to the Laramide and younger events, so they will be discussed in a subsequent section.

West of the shear zone, slivers and slices of Proterozoic to Jurassic rocks crop out and are apportioned to various terranes, such as Wrangellia, Taku, and Alexander (Rubin and Saleeby, 1991: Saleeby, 2000; Crawford et al., 2000). The Jurassic rocks are similar to those of the Sierra Nevada and Peninsular Ranges (Hildebrand, 2013; Hildebrand and Whalen, 2014b) in that rocks of the Bowen Island (Friedman et al., 1990) and Harrison Lake groups (Fig. 32), generally interpreted to represent part of the 202–165 Ma Bonanza arc complex of Vancouver and Queen Charlotte Islands (DeBari et al., 1999; Isachsen, 1987; Canil et al., 2010), were deformed prior to deposition of Callovian sedimentary rocks (Mahoney et al., 1995) and intrusion by abundant 155–145 Ma postcollisional plutons (Friedman and Armstrong, 1995). The Jurassic assemblages were overlain by Oxfordian to Albian metasedimentary and metavolcanic rocks, and intruded by 130–100 Ma plutonic rocks—all ascribed variously along strike to the Gravina-Gambier arc (Berg et al., 1972; Rubin and Saleeby, 1991, 1992; Haeussler, 1992; Journeay and Friedman, 1993; Crawford et al., 2000; Butler et al., 2002; Manuszak et al., 2007; Ricketts, 2008). Although geochemical analyses from the Gravina-Gambier rocks are relatively dated and the rocks altered, a few trace elements are plotted on Figure 33, where they generally plot within the arc fields.

The rocks of the Gravina-Gambier arc and their basement were deformed and metamorphosed during the mid-Cretaceous at ca. 100 Ma (Haeussler, 1992; Rubin et al., 1990; Rubin and Saleeby, 1991; Journeay and Friedman, 1993; Lynch, 1992, 1995; Crawford et al., 2000; McClelland and Mattinson, 2000; Crawford et al., 2005; Mahoney et al., 2009; Gehrels et al., 2009). Deformation was accompanied in part by the emplacement of a suite of deep-seated, syn- to posttectonic 100–85 Ma tonalitic, quartz dioritic, and granodioritic plutons commonly containing magmatic epidote and titanite (Crawford and Hollister, 1982; Zen, 1985; Zen and Hammarstrom, 1984a, 1984b; Crawford et al., 1987; Himmelberg et al., 2004; Gehrels et al., 2009). Selected trace-element data from these plutons are

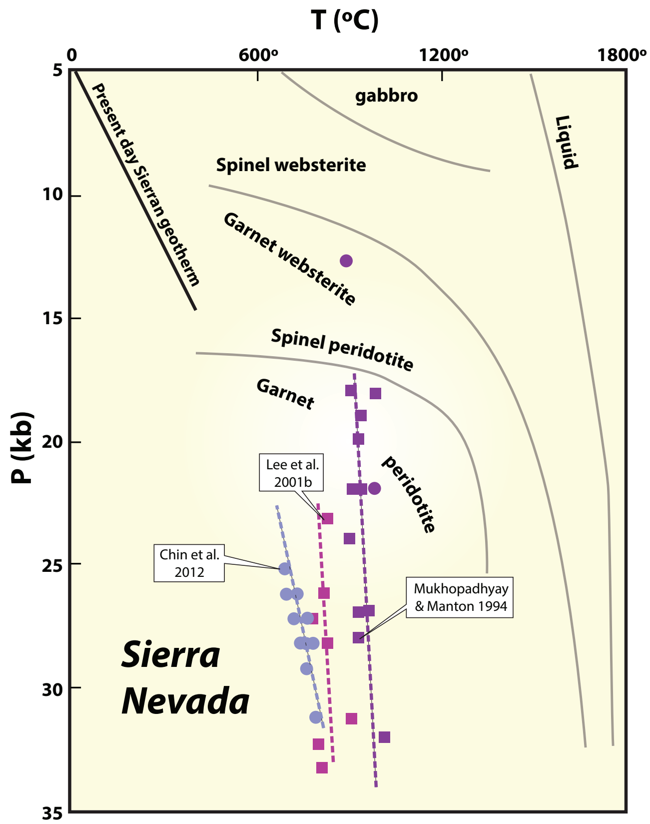


Figure 24. Pressure-temperature (*P-T*) grid illustrating the final equilibrium conditions calculated for various groups of mantle xenoliths from the Sierra Nevada. Although the arrays are artifacts of the different reaction kinetics of the geothermometers and geobarometers, the points represent minimum *P* values (Chin et al., 2012). These high pressures (~3 GPa) support the model presented here in that the La Posta and Sierran Crest magmatic suites resulted from melting at great depth caused by slab failure, upwelling of asthenosphere, and melting within the garnet stability field.

plotted on Figure 34. They plot mostly within the slab failure fields, as do rocks from the Sierra and Peninsular Ranges. Thus, the ages and syn- to posttectonic setting, along with the geochemistry, suggest a slab failure origin for these post-Oregonian plutons.

Within the High Cascades of Washington, the oldest exposed rocks are schists, amphibolite, metachert, marble, metapelite, and ultramafic rocks of Paleozoic–Mesozoic age that were deformed and metamorphosed at ca. 100 Ma (Misch, 1966; Mattinson, 1972; McGroder, 1991; Miller et al., 2009). They were intruded by many plutons in the age range 96–89 Ma (Fig. 35) as well as younger bodies of Laramide age, many now gneissic.

The most extensively exposed component of the region, the Skagit gneiss, forms a NNW-trending band, 20–30 km wide, for over 100 km through the High Cascades (Fig. 35), where it comprises mostly biotite tonalitic to trondhjemitic orthogneiss with lesser amounts of biotite granodiorite cutting older amphibolites and schists. The oldest orthogneiss body is 89 Ma, but the largest volume of orthogneiss was intruded between 76 and 59 Ma, with another pulse at 50–45 Ma (Miller et al., 2009). Thus, the bulk of this magmatism was Laramide and will be discussed in the appropriate section.

The Mount Stuart batholith is one of the largest intrusive complexes in the North Cascades, and it is composed mostly of ilmenite-bearing tonalite and granodiorite dated by U-Pb methods on zircon to be 96–91 Ma (Walker and Brown, 1991; Matzel et al., 2006a; Matzel, 2004; Stowell et al., 2011). The batholith postdates the ca. 100 Ma 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., 2006a). Other similar age intrusive complexes of the area, such as 92–87 Ma Black Peak, the 92–90.5 Ma Seven-Fingered Jack, and the 92–89.5 Ma Tenpeak intrusive complexes, were intruded at 1–3 kbar, 6–7 kbar, and 7–10 kbar, and have $\delta^{18}O_{zircon}$ of 6%–7%, 7%–8%, and 7.7%–9%, respectively (Shea, 2014). Modern chemical analyses with extensive trace-element data sets are scarce, but we plotted data from two post-Oregonian plutonic complexes (Fig. 36), and the rocks

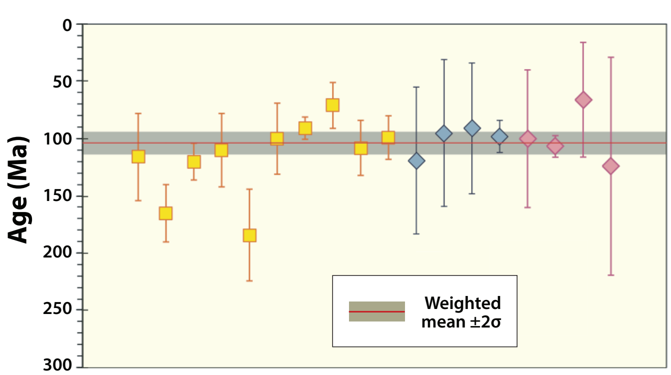


Figure 25. U-Pb age (Ma) plot for metamorphic rims on zircon cores in garnet-bearing, granulite-facies sedimentary quartzite xenoliths from the Cenozoic volcanic rocks in the Sierra Nevada from Chin et al. (2013). The different colored symbols represent different xenoliths. Hf isotopic ratios from the zircons are similar to those from Proterozoic basement of the eastern Sierra Nevada. The data suggest that Archean and Proterozoic metasedimentary rocks of Nevada were subducted westward beneath the Sierra Nevada and metamorphosed at ca. 100 Ma because the cores are Precambrian, rims are ca. 100 Ma, and Precambrian sedimentary rocks are unknown from regions west of the Sierra Nevada.

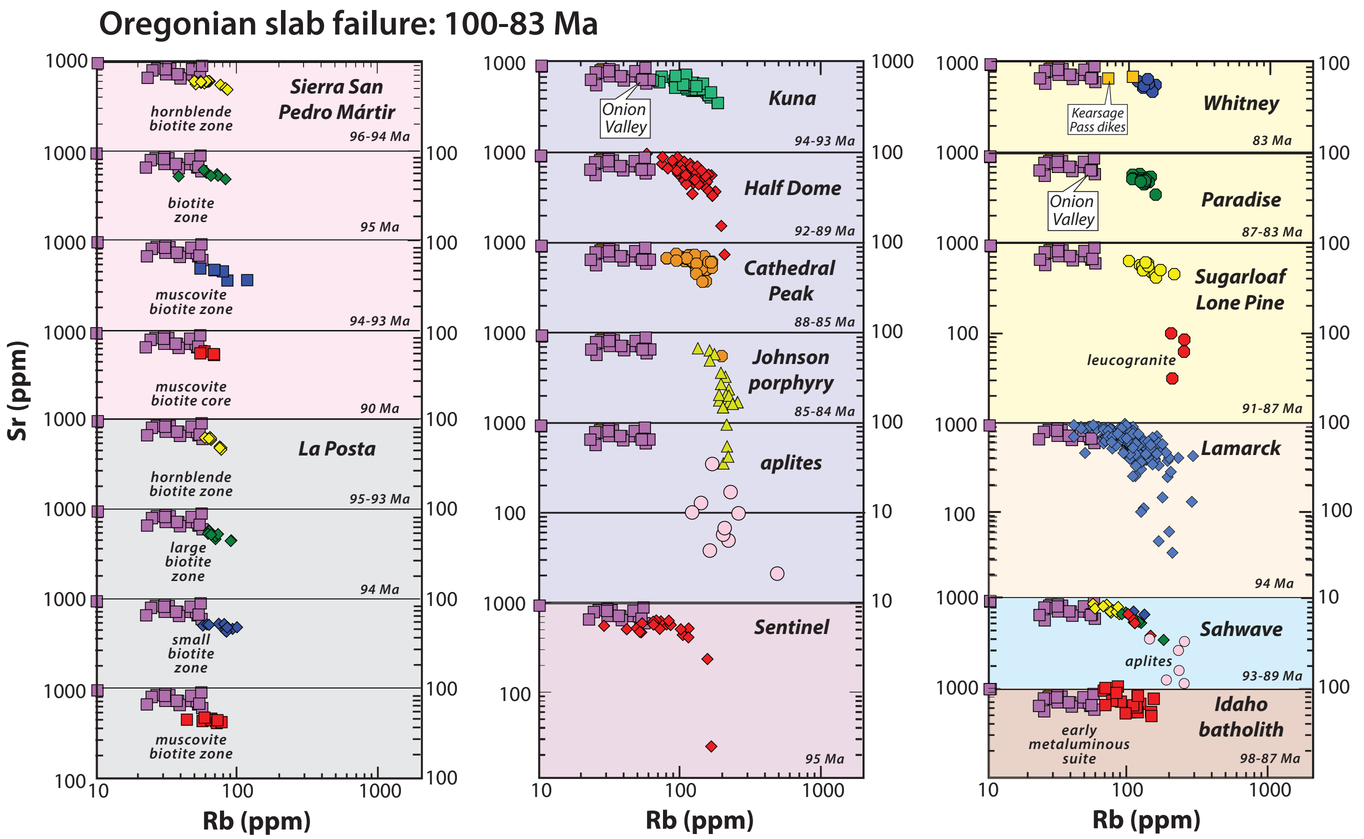


Figure 26. Rb vs. Sr plots for Oregonian (100–83 Ma) slab failure plutonic units comprising the Sierra San Pedro Mártir, Tuolumne, Mount Whitney, and Sahwave plutonic suites plus the Lamarck and Idaho batholiths, illustrating the similarity of individual plutons within each suite and also the remarkable similarity of plutons between suites. On each plot, samples from the relatively unevolved Onion Valley hornblende gabbro (Sisson et al., 1996) are shown for reference. While the plots show depletion of Sr with increasing Rb in individual plutons, the starting composition of each pluton is about the same, which rules out any sort of fractionation or mixing scheme to produce the different plutons within each plutonic complex. Data sources are as in Figure 20, except for Lamarck from Frost and Mahood (1987); Idaho batholith data are from Gaschnig (2015, personal commun.).

have trace-element ratios more consistent with slab failure than with arc magmatism.

SALINIA

The Salinian block (Fig. 37) is a narrow composite terrane, 40–70 km wide and more than 500 km long, that occupies much of California's Coast Ranges (Ross, 1978). Rocks of the Salinian block include amphibolite-granulite-facies gneiss and schist lacking the pure quartzite and carbonate rocks characteristic of the North American margin (Ross, 1977), and these are cut by 100–82 Ma plutons ranging in composition from gabbro to granodiorite, although intermediate compositions dominate volumetrically (Mattinson, 1978, 1990; Kistler and Champion, 2001; Kidder et al., 2003; Chapman et al., 2014). The Salinian block was considered to be an out-of-place "orphaned" arc block, because it is missing its accretionary prism and forearc on the west and is bounded by the San Andreas fault on the east (Page, 1970, 1982).

Several workers have considered the Salinian block to represent part of the California arc that was originally located between the Sierra Nevada and the Peninsular Ranges batholith (Ducea, 2001; Barbeau et al., 2005; Chapman et al., 2014). This idea was based largely on restorations of >300 km slip across the San Andreas fault—and thus Salinia may have been located south of the Sierra Nevada just prior to the Neogene slip on the fault—but its location during the Cretaceous is unclear and hinted at only by controversial paleomagnetic results, which suggest a more southerly provenance. Paleomagnetic data from Upper Cretaceous and Paleocene sedimentary rocks of the Salinian block suggest that they were deposited 2800–2100 ± 500 km south of their present location (Champion et al., 1984), although there is some controversy about the data based on study of different rocks (Whidden et al., 1998). Nevertheless, paleontological data suggest that the faunal assemblage of Salinia is a reasonable match with those in the Peninsular Ranges of southern California (Elder and Saul, 1993), and they are considered to be far-traveled rocks (Hagstrum et al., 1985). Given that the entire Cordillera west of the Cordilleran

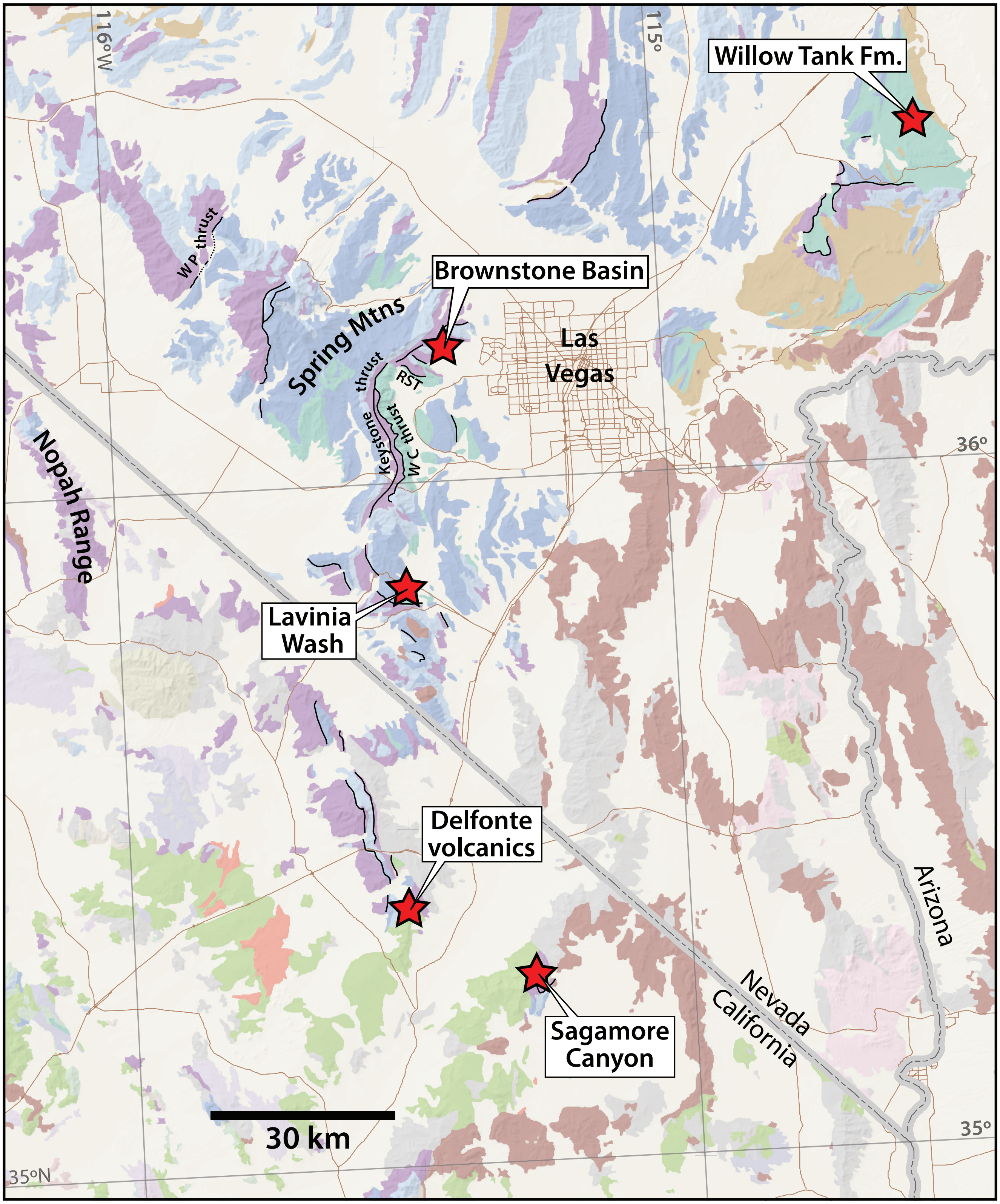


Figure 27. Location map from Wells (2016) on a geological base from Sue Beard (U.S. Geological Survey) showing the location of sites near Las Vegas, Nevada, with evidence for 100 Ma thrusting. WC—Wilson Cliffs; RST—Red Springs thrust; WP—Wheeler Pass.

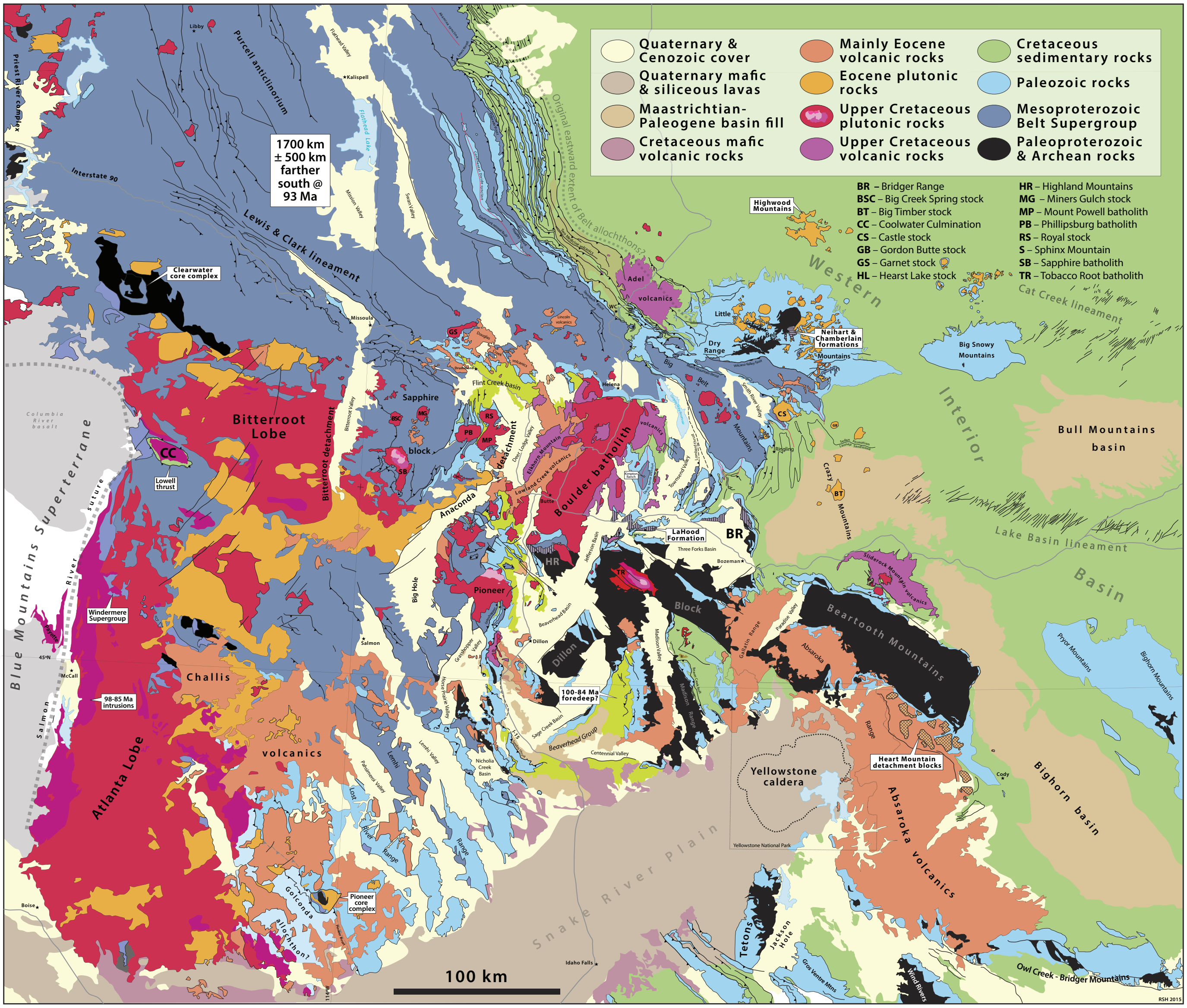


Figure 28. Regional geological sketch map of the Helena salient, north-central U.S. Cordillera showing distribution of plutonic rocks and illustrating features discussed in text. This map was compiled from the *Geologic Map of Montana* (Vuke et al., 2007), the *Geologic Map of Idaho* (Lewis et al., 2012), and the *Geologic Map of Wyoming* (Love and Christianson, 1985), with additions from Gaschnig et al. (2010); Lewis et al. (2007); Lund et al. (2008); and other sources cited in text. See insert for larger version of this figure.

fold-and-thrust belt migrated northward 1300 km after 70 Ma (Hildebrand, 2015), the Salinian block has apparently migrated northward a minimum of 1600 km since the latest Cretaceous.

Although modern geochemical data are scarce, Chapman et al. (2014) provided a few geochemical analyses from a variety of plutons and their wall rocks, but, unfortunately, they did not publish analyses for Nb and Ta. We plotted the remaining key trace elements and Sr/Nd isotopes from obvious metaplutonic bodies (Figs. 23 and 38), and they are consistent with other Oregonian slab failure suites, with Sm/Yb ratios >3, which are typical for slab failure rocks derived largely from a garnet-bearing source.

No matter where it was located, the Salinian block appears to represent only half of a typical mid-Cretaceous Cordilleran batholith, because it contains the high-grade exhumed sector (3–7.5 kbar) intruded by 100–84 Ma plutons, but it contains no lower-grade pre–100 Ma volcanics and associated arc plutons typical of other Cretaceous batholithic terranes. This led Hildebrand and Whalen (2014a) to suggest that it once belonged with the Coastal batholith of Peru, which some recognized to be another "orphaned" block (Loewy et al., 2004). There, 7–9 km of relatively low-grade arc rocks, known as the Casma Group, were deformed at 100 Ma; however—while the batholith contains

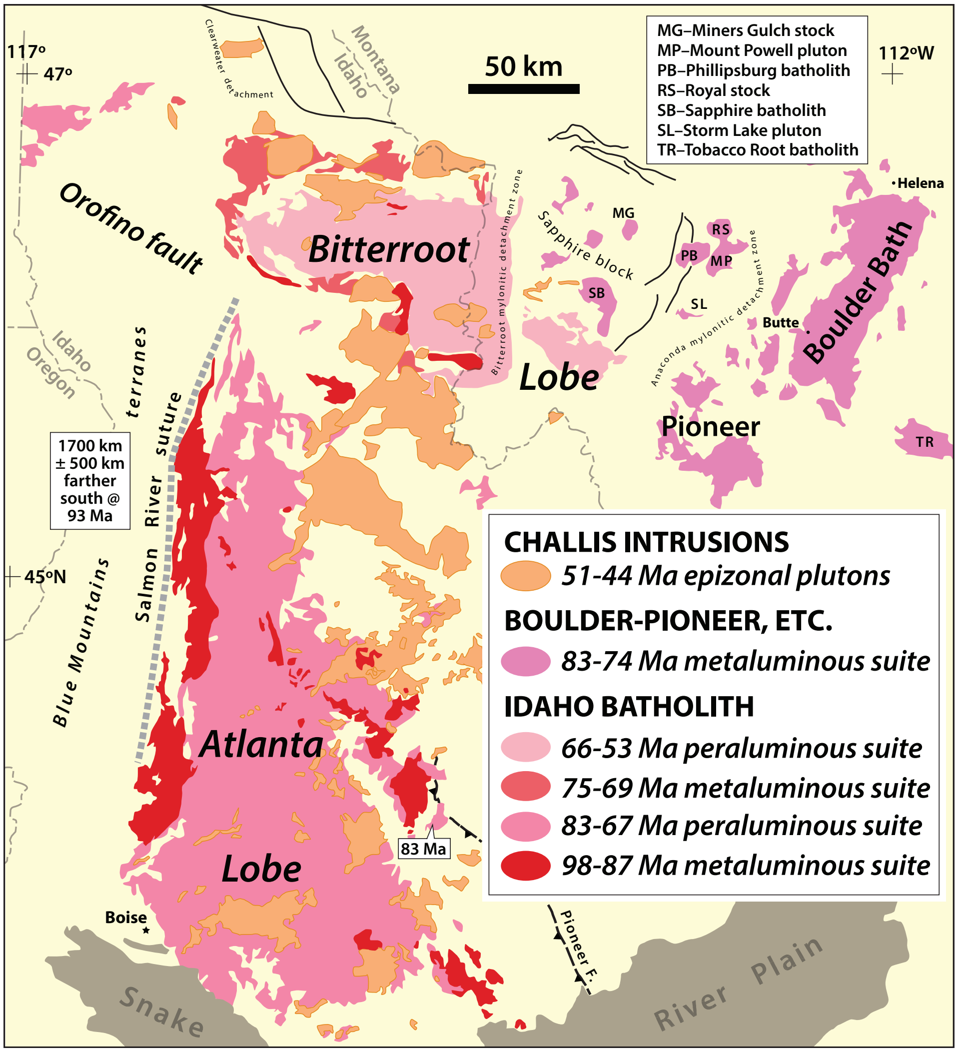


Figure 29. Sketch map showing the locations and ages of principal plutonic rocks of the Helena salient (modified from Gaschnig et al., 2010; and others cited in text).

some slab failure plutons—it is missing its exhumed metamorphic hinterland and main mass of 100–85 Ma slab failure plutons. Thus, during the Cretaceous, given the suggestion of large-scale meridional migration from paleomagnetic studies, instead of lying between the Sierra Nevada and Peninsular Ranges batholiths, the Salinia block was more likely located well to the south, and was possibly joined with the Arequipa block and its cover of Casma volcanics, which, on the basis of magmatic ages and deformation, both probably formed part of the Oregonian belt, although additional work is required to document and confirm the precise relations.

MID-CRETACEOUS OREGONIAN EVENT

A century ago, Blackwelder (1914) used the term Oregonian to refer to the mid-Cretaceous deformational event in western North America. This event, or orogeny, took place after a rifting and spreading event that occurred following the Late Jurassic–Neocomian Nevadan orogeny and associated postcollisional magmatism (e.g., Allen and Barnes, 2006; Barnes et al., 2006; Arth et al., 1989a, 1989b; Roeske et al., 1995, 2015). Following the rifting, continued spreading led to the formation of an ocean (Bisbee-Arperos seaway), which was of indeterminate width, but contained a locally well-preserved, west-facing passive-margin terrace along its eastern side (Dickinson and Lawton, 2001a, 2001b; Hildebrand and Whalen, 2014b). As we have seen, fragments of the basin are widespread up and down the Cordillera and are known in Mexico from Zihuatanejo to Sonora up to southern Arizona (Hildebrand and Whalen, 2014b; Peryam et al., 2012; Lauierre et al., 1992a, 1992b; Monod et al., 1994; Centeno-García et al., 2008; González-León et al., 2008; Martini et al., 2012); the Sierra Nevada of California (Nokleberg, 1981; Memeti et al., 2010a); the Coastal plutonic complex of British Columbia and Alaska (Ricketts, 2008; Berg et al., 1972; Rubin and Saleeby, 1991, 1992; Lynch, 1991, 1992; Monger, 1991; Haeussler, 1992; Arthur et al., 1993; Ridgway et al., 2002; Kalbas et al., 2007); the Yukon Territory of Canada north of the Tintina fault to the Mackenzie River delta (Dixon, 1992), and possibly

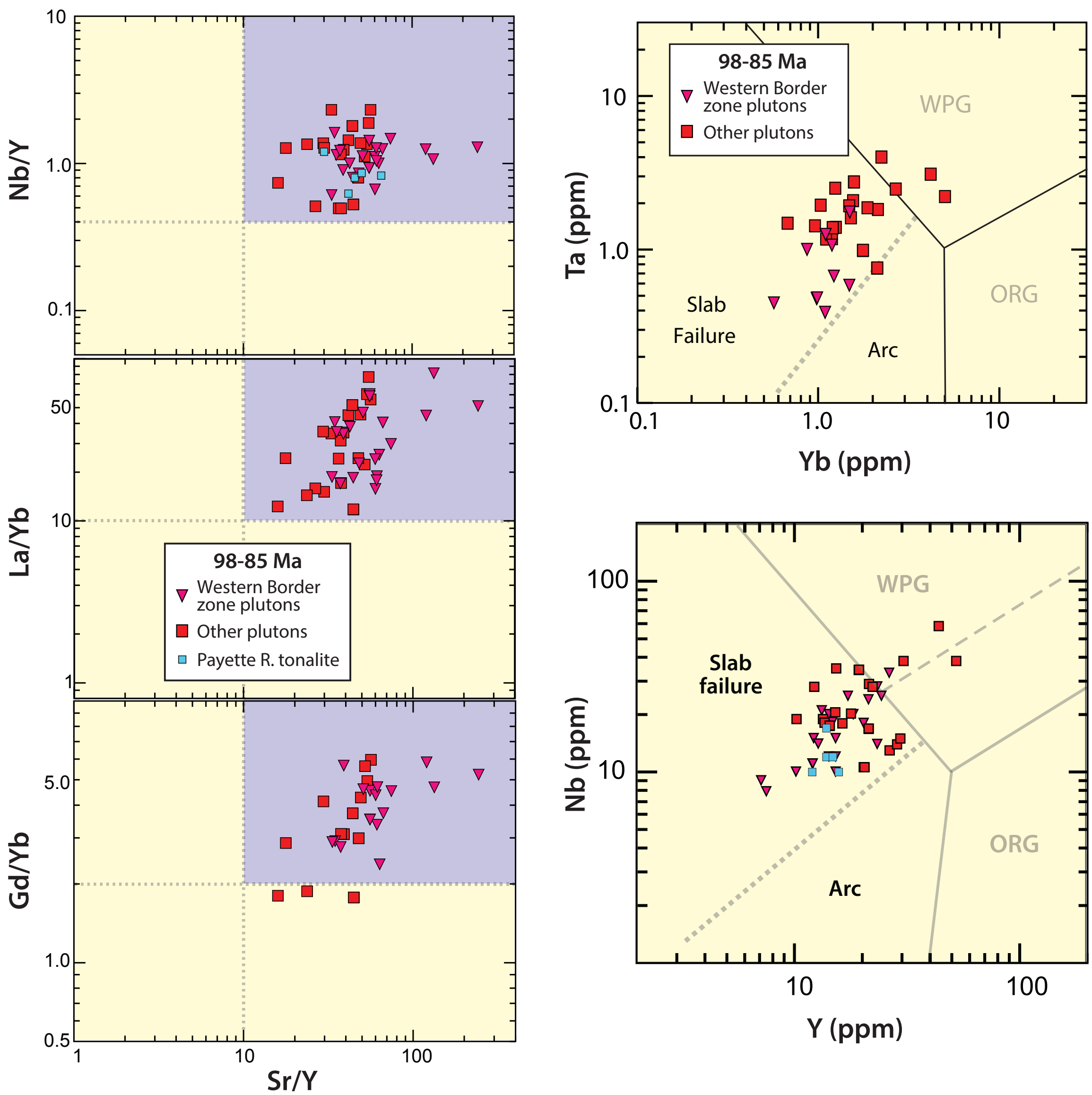


Figure 30. Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y (as in Fig. 8) and Ta vs. Yb and Nb vs. Y discrimination diagrams for 98–85 Ma Oregonian plutons in the Helena salient. Based on these data and the postkinematic nature of plutonism, we interpret the bodies to be slab failure features, not arc features. Data are from Manduca et al. (1993); Gaschnig (2015, personal commun.). WPG—within-plate granite; ORG—ocean-ridge granite.

the Canada Basin, which opened at the same time (Grantz et al., 2011; Helwig et al., 2011).

The evidence in western Mexico suggests that the Alisitos–Santiago Peak arc collided with a west-facing passive margin at ca. 100 Ma. The polarity of the subduction was westward, such that the western edge of the passive margin was partially subducted beneath the arc. The platform terrace within the seaway was buried by westerly derived orogenic debris (Mack, 1987).

A feature that supports the idea that the basins collectively formed parts of a more extensive linear basin is the observation that they all closed—as documented by deformation and juxtaposition of basin fill and arc rocks—near the Albian-Cenomanian boundary at ca. 100 Ma, except for the Canada Basin, which stopped spreading at about that time (Hildebrand and Whalen, 2014b; Lynch, 1992; Journeay and Friedman, 1993; Crawford et al., 2000; Haeussler, 1992; Manuszak et al., 2007; Gehrels et al., 2009; Grantz et al., 2011). As stated, the evidence in western Mexico suggests that the Alisitos–Santiago Peak arc collided with a west-facing passive margin at ca. 100 Ma (Rangin, 1986; Hildebrand and Whalen, 2014b). To the north, an easterly vergent thrust belt occurs east of the Sierra Nevada, and thrusts within it are ca. 100 Ma (Wells, 2016), as discussed earlier, but the suture remains elusive, perhaps due to extensive Basin and Range tectonism. In the Coast plutonic complex of British Columbia,

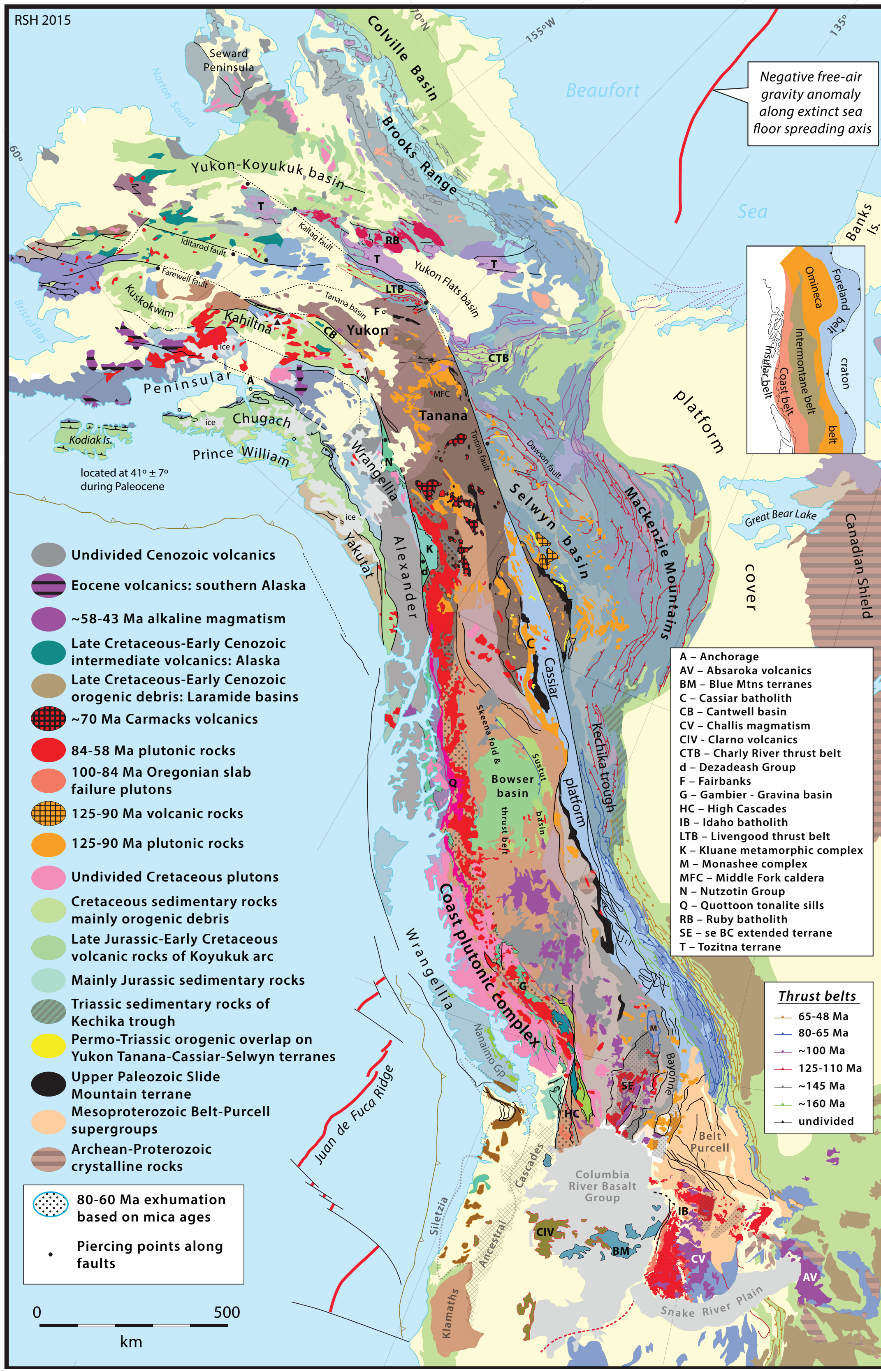


Figure 31. Geological sketch map of the northern North American Cordillera illustrating features discussed in text, tectonic terranes, plutonic belts, and approximate ages of fold-and-thrust belts, based on Wheeler and McFeely (1991); Reed et al. (2004); Pană and van der Pluijm (2015); Dover (1992, 1994); Hults et al. (2013); and Norris (1984). BC—British Columbia. See insert for larger version of this figure.

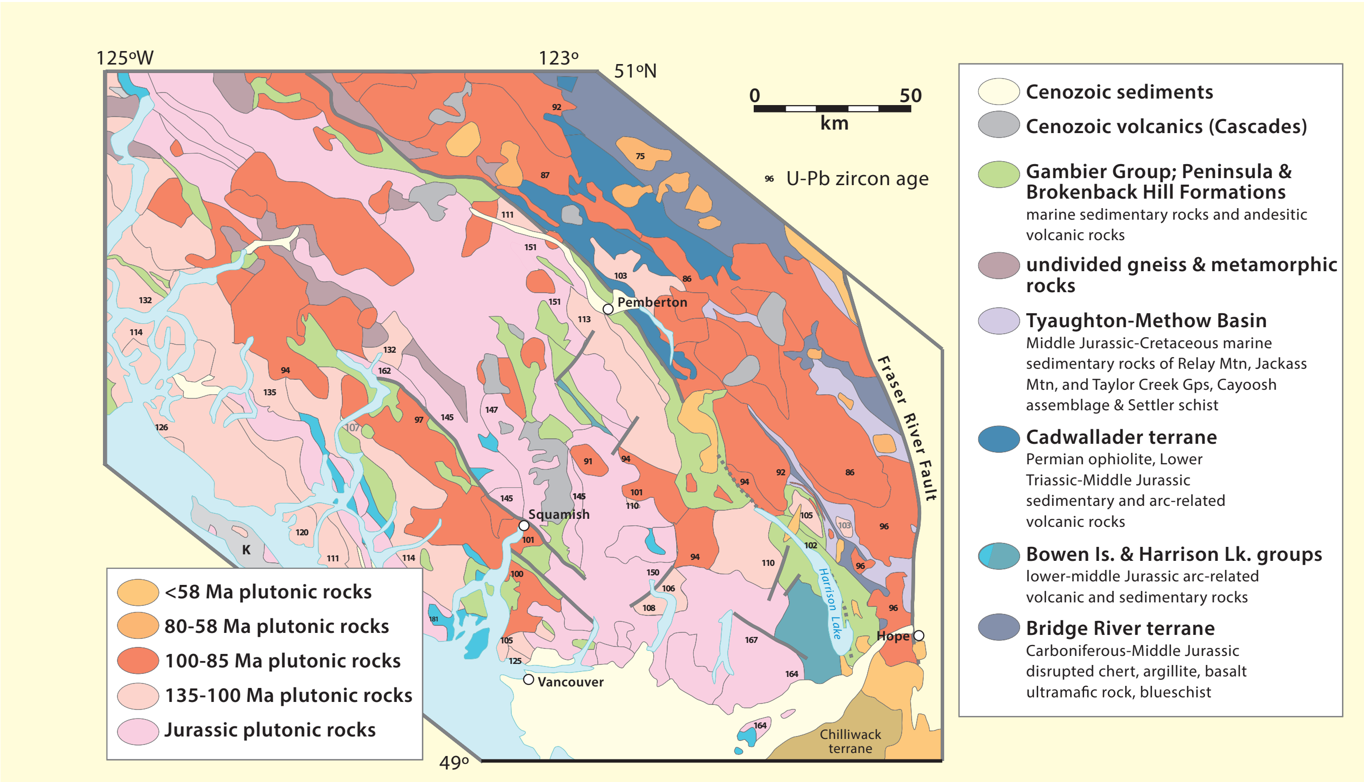


Figure 32. Geological sketch map of the southern portion of the Coast plutonic complex, British Columbia, modified from Friedman and Armstrong (1995), with additional ages from Gibson and Monger (2014).

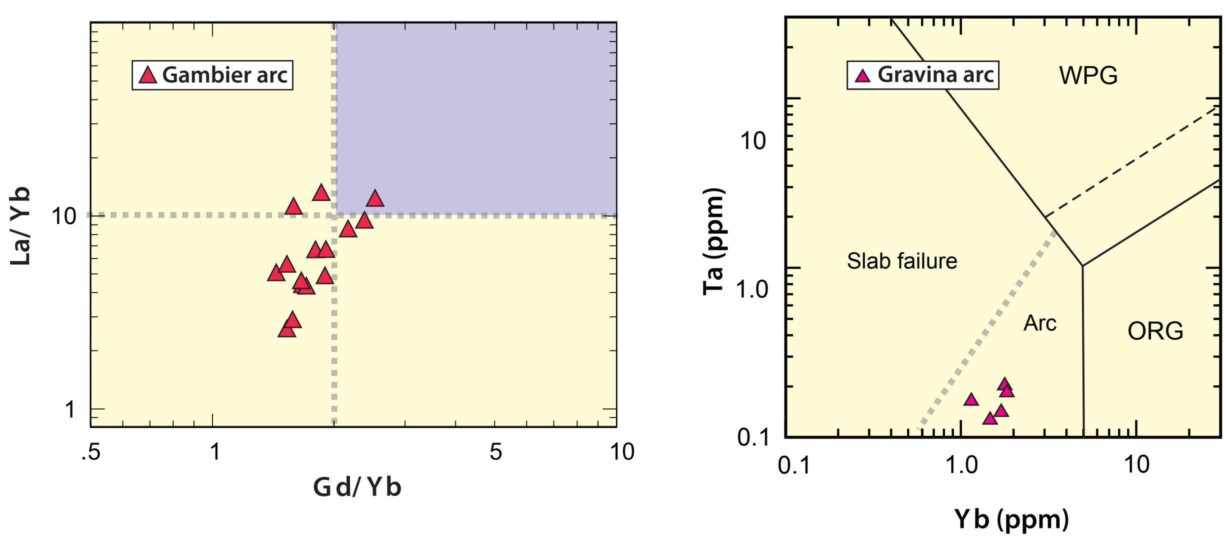


Figure 33. La/Yb vs. Gd/Yb and Ta vs. Yb discrimination diagrams for volcanic rocks of Gambier Group, generally interpreted to represent remnants of a volcanic arc (Lynch, 1995), and the coeval Gravina arc (Rubin and Saleeby, 1991). Trace elements required for other discrimination plots were unavailable. WPG—within-plate granite; ORG—ocean-ridge granite.

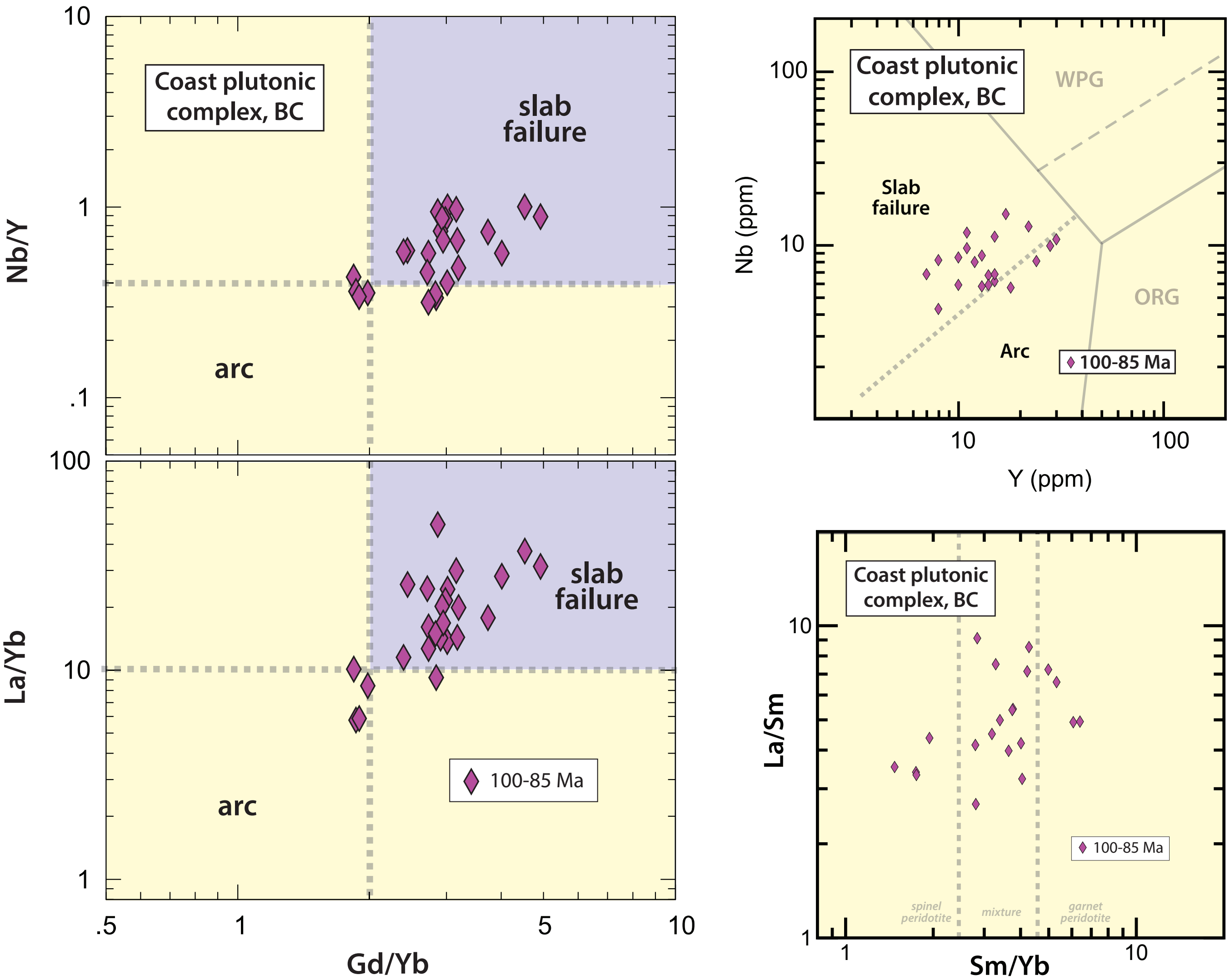


Figure 34. Nb/Y and La/Yb vs. Gd/Yb discrimination diagrams, Nb vs. Y discrimination diagram, La/Sm vs. Sm/Yb plot for 100–85 Ma Oregonian rocks of the Coast plutonic complex, British Columbia (BC). Data are from Crawford et al. (2005); Mahoney et al. (2009); Girardi et al. (2012). Labeled subdivisions on Sm/Yb vs. La/Sm plot indicate possible source rocks from work of Putirka (1999). WPG—within-plate granite; ORG—ocean-ridge granite.

Albian rocks of the Gambier-Gravina arc were deformed (Lynch, 1992; Arthur et al., 1993) prior to the emplacement of the 92 Ma Dickson-McClure batholith (Parrish, 1992).

Early work on the structure of the southern Coast plutonic complex suggested that ~100 Ma faults were low-angle thrust faults that placed high-grade metamorphic rocks westward over lower-grade rocks (Journeay and Friedman, 1993), which led to the general notion that subduction beneath the belt was eastward. However, more recent work suggests that the faults are high-angle reverse faults with over 10 km of reverse displacement (Brown et al., 2000; Gibson and Monger, 2014) and that the higher-grade rocks were placed westward atop the western plutonic terrane, which consists dominantly of precollisional 125–100 Ma arc plutons (Gehrels et al., 2009) and sub-amphibolite-grade arc volcanic and associated epiclastic rocks (Lynch, 1992; Arthur et al., 1993). These findings suggest that the arc was on the western block prior to collision at 100 Ma, and that the upper plate was more extensive than previously thought. If so, then the reverse faults are back thrusts, and, therefore, the collisional suture related to basinal closure, if preserved, should lie to the east, not the west. Similar relations exist along the western side of the Peninsular Ranges batholith, where amphibolite-grade upper-plate rocks were thrust westward over much lower-grade volcanic and sedimentary rocks along steep easterly dipping faults such as the Main Martir and Rosarito thrusts (Schmidt and Paterson, 2002; Schmidt et al., 2014; Hildebrand and Whalen, 2014b). These

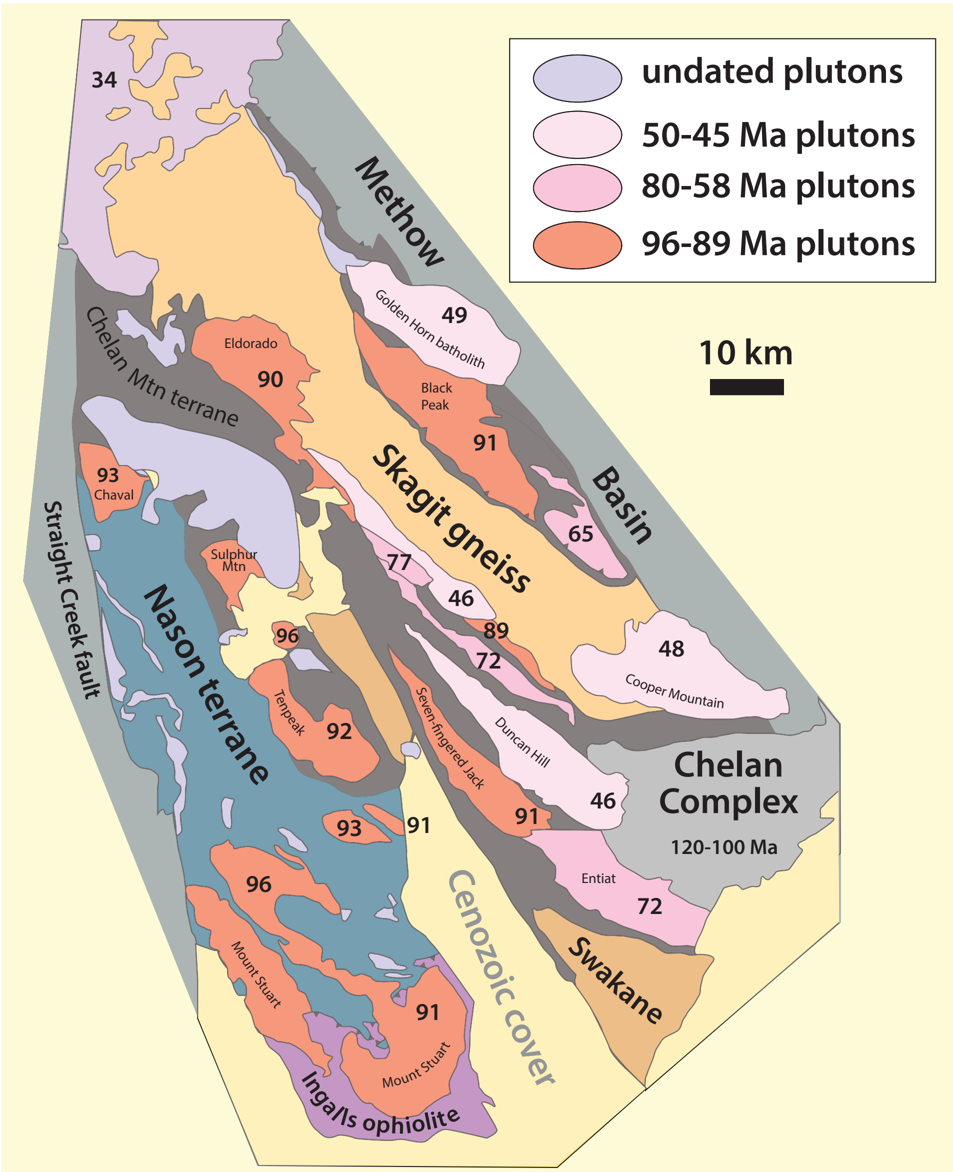


Figure 35. Geological sketch map of North Cascades core showing ages of plutons (modified from Miller et al., 2009). Note that there are three groups with ages similar to other suites within the Cordillera. Oregonian plutons are 96–89 Ma; Laramide plutons are 80–58 Ma; and plutons emplaced during extensional collapse are 50–45 Ma. The 120–100 Ma Chelan complex likely represents the roots of the pre–100 Ma magmatic arc.

huge reverse faults are a form of rétrocharriage, or back thrusts (Roeder, 1973) and apparently form where there is strong coupling between the descending lower and overriding upper plate, as observed in the recent models by Vogt et al. (2017).

The nature of the Oregonian event helps to resolve the confusion in the Canadian Cordillera regarding the timing of collision between the Insular and Intermontane superterranes, which has variously been ascribed to have occurred during the Jurassic (Anderson, 1976; Tipper, 1984; van der Heyden, 1989), Early Cretaceous (Kleinspehn, 1985; Armstrong, 1988), or mid-Cretaceous (Tennyson and Cole, 1978; Monger et al., 1982; Crawford et al., 1987; Thorkelson and Smith, 1989). Whereas in our model, the Oregonian event was caused by closure of a marginal basin that had opened after the Late Jurassic–latest Jurassic–Neocomian Nevadan event, it is possible that the last two models could both be correct. Just as occurred much farther south, the two superterranes might have collided during the latest Jurassic–Early Cretaceous, possibly during the Nevadan event, and then been torn apart during development of the Bisbee-Arperos seaway at ca. 135 Ma, only to be reconnected at ca. 100 Ma during westerly subduction and closure of the seaway.

CORDILLERAN RIBBON CONTINENT AND WESTERLY SUBDUCTION

In our model, the Bisbee-Arperos seaway formed within what is known as the Cordilleran ribbon continent (Fig. 39), but that part of the ribbon continent east of the seaway collided, during the Sevier event, with the western margin of North America before complete closure of the basin some 10–15 m.y. after

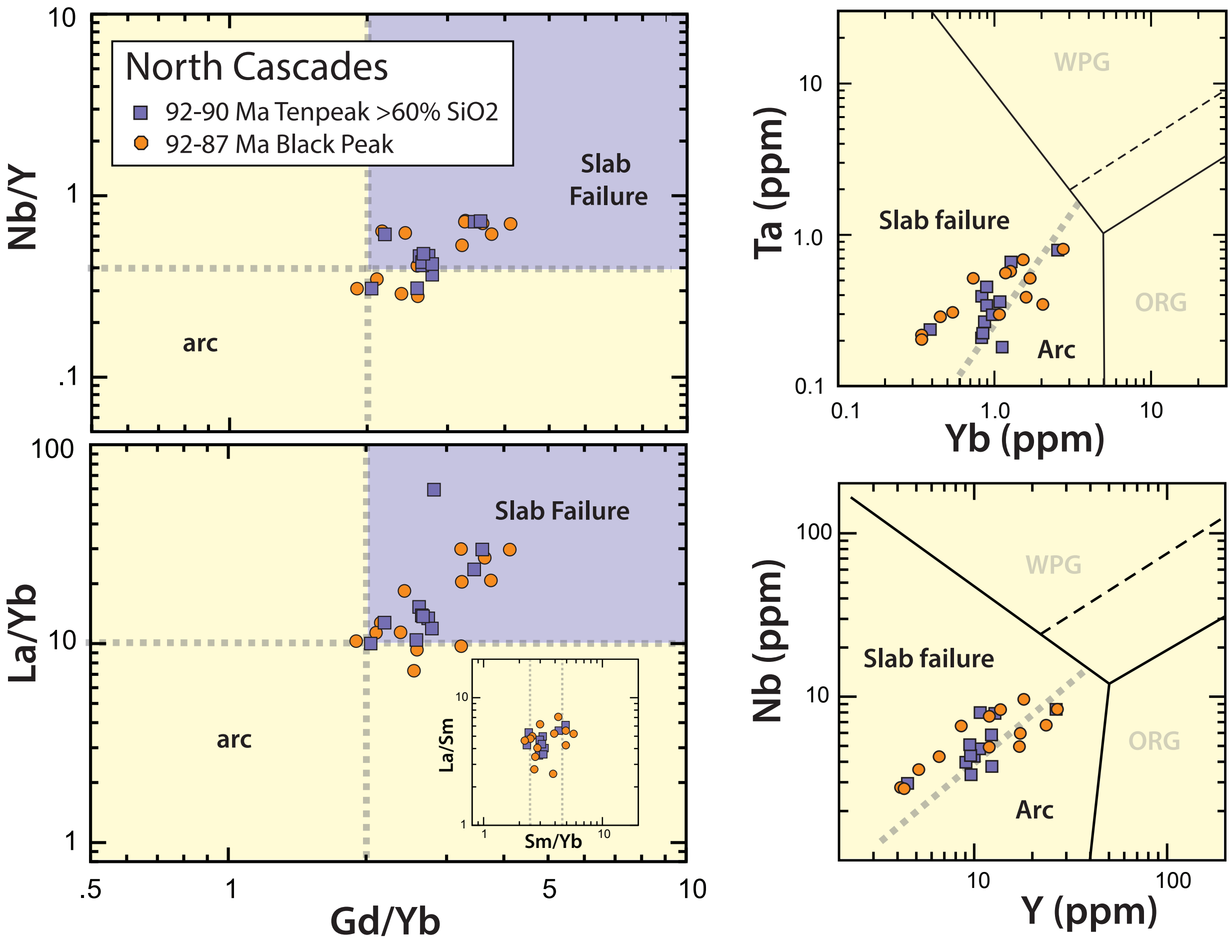


Figure 36. Nb/Y and La/Yb vs. Gd/Yb and Ta vs. Yb and Nb vs. Y discrimination diagrams for the Northern Cascades Tenpeak and Black Peak plutonic suites. Data are from Shea (2014). WPG—within-plate granite; ORG—ocean-ridge granite.

opening. The Sevier event is the oldest Mesozoic deformational thickening event known to affect the Cordilleran, or Laurentian, platform terrace and is well documented by easterly vergent folds and thrust faults, as well as a flexural foredeep involving the continental terrace in the western United States (Armstrong, 1968; DeCelles, 2004; DeCelles and Coogan, 2006; Yonkee and Weil, 2011). For those who have not read our previous contributions, and to better place Cordilleran plutonism into an overall tectonic framework, we briefly introduce the Cordilleran ribbon continent and explore some of its structural elements, before returning to our examination of Cordilleran batholiths.

The Cordilleran ribbon continent was a linear amalgamation of rocks assembled largely during the Mesozoic and comprising nearly the entire spectrum of geological settings (Johnston, 2008; Hildebrand, 2013, 2014). The ribbon was hypothesized to have formed above a long-lived zone of mantle downwelling located between the Panthalassic and proto-Pacific oceanic tracts (Hildebrand and Whalen, 2014a) and then to have collided with western North America above a westerly dipping subduction zone during westward migration of North America as the Atlantic Ocean opened (Moores, 1970); Johnston and Borel, 2007; Johnston, 2008; Hildebrand, 2009).

The detailed geological evidence for the existence of the Cordilleran ribbon continent, and a westerly subduction regime between it and North America, was presented in several recent papers (Johnston, 2008; Hildebrand, 2009, 2013), but it includes (1) the existence of two distinct and different-age passive-margin terraces, the Lower Cambrian Antler and Middle Cambrian Laurentian platforms, each of which had different histories; (2) a lack of latest Neoproterozoic–Early Cambrian rift basins and related volcanic-evaporitic sequences on the western Laurentian margin; (3) the presence of persistent mafic magmatism throughout the Paleozoic on the Antler platform; and (4) a conspicuous lack of deformation and orogenic sedimentation on the North American platform from the Cambrian to the Cretaceous, despite major crustal thickening, many kilometers of exhumation, and related sedimentation located just to the west during the Jurassic and late Paleozoic. Additionally, high-resolution mantle tomography beneath North America (Grand, 1994; Grand et al., 1997) revealed a huge, steeply inclined slab wall that extends down into

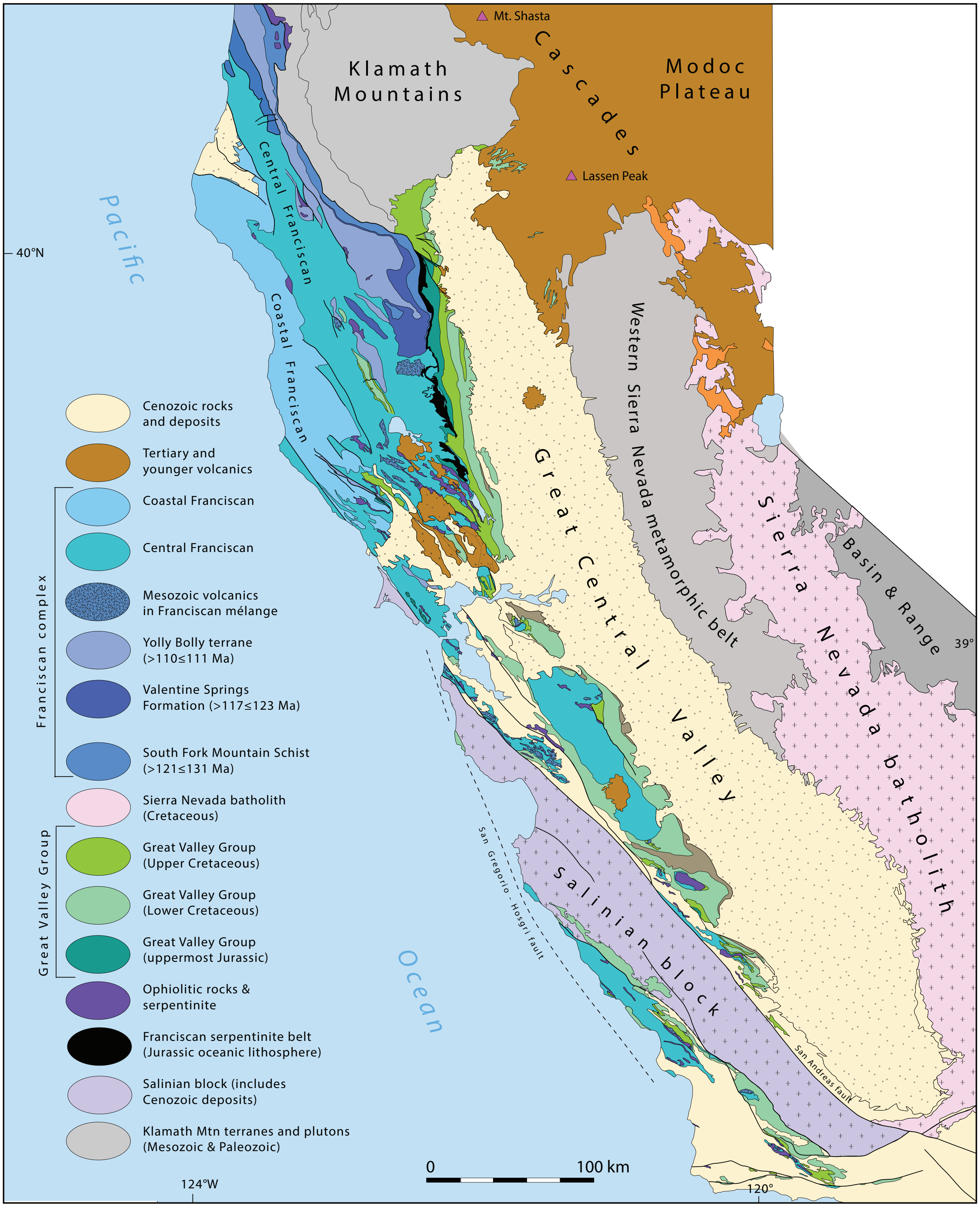


Figure 37. Geological sketch map showing the location of the Salinian block and its relation to nearby geologic elements within northern California, from Hildebrand (2013), but modified from maps by Jennings (1977) and Dumitru et al. (2010).

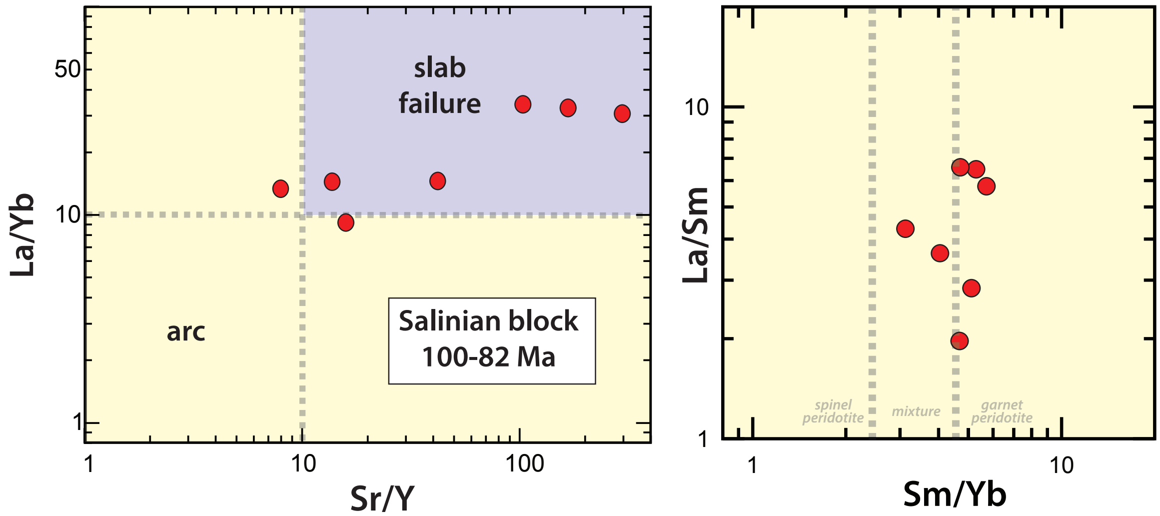


Figure 38. La/Yb vs. Sr/Y and La/Sm vs. Sm/Yb discrimination diagrams for 100–85 Ma plutonic rocks of the Salinian block (data from Chapman et al., 2014).

the mantle for over 1000 km, marks the site of long-lived subduction, and provides independent support for the westward-dipping subduction model (Sigloch and Mihalynuk, 2013, 2017).

The standard alternative to the ribbon continent model involves long-lived easterly subduction of oceanic lithosphere beneath the Sierra Nevada, which most workers inferred to have been constructed atop North American cratonic crust, and development of the Cordilleran fold-and-thrust belt and related flexural foredeep in a back-arc setting (Dewey and Bird, 1970; Burchfiel and Davis, 1975; Dickinson, 2004). A recent, and widely accepted, variant of the standard back-arc model involves cyclic magmatism generated by westerly underthrusting of lower continental lithosphere in a back-arc setting and melting of that lithosphere by asthenospheric magma (DeCelles et al., 2009). This model supposes that all of the Triassic–Jurassic magmatism is related to the same subduction, which, given the number of different collisions (Hildebrand, 2013), is unlikely. Second, because the back-arc model calls for 300–400 km of thin-skinned shortening east of the Sierra Nevada (DeCelles, 2004; DeCelles et al., 2009), some 300–400 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) have suggested that this 300–400 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. 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 the model; but even were it possible, the model suffers from other problems. For example, as mixtures of upper and middle continental crust are similar in bulk composition to the Sierran batholith (Ducea, 2002), nearly an equal volume of cratonic crust must be melted to create the same mass with the bulk composition of the Sierra Nevada, so there would be little restite remaining to sink into the mantle.

Cordilleran Fold-and-Thrust Belt

We now outline several geologic events that overlap in part within the Cordillera fold-and-thrust belt, which extends from northern Canada to southern Mexico (Fig. 40). Earlier, we showed that in just a small segment, located near Las Vegas, Oregonian and younger thrusts overlap spatially. The relations are also complex elsewhere, with several ages of thrusts occurring in discrete domains—possibly megaphacoids juxtaposed on strike-slip faults—within the fold-and-thrust belt (e.g., Pană and van der Pluijm, 2015). Deformation within the belt is both older and younger than that of the 100 Ma Oregonian event.

The older event is generally referred to as the Sevier, whereas the younger is known as the Laramide. Both events had a thin-skinned component, whereas the Laramide had both thin- and thick-skinned components of slightly different ages. Most workers have interpreted both events to form a deformational continuum and to have occurred within a retro-arc setting (Burchfiel and Davis, 1975; Yonkee and Weil, 2015; DeCelles and Currie, 1996; DeCelles, 2004; DeCelles and Coogan, 2006; DeCelles et al., 2009; Fuentes et al., 2010), but Hildebrand (2013, 2014) argued that they represent two discrete collisional events between ribbon-like terranes and North America. Because the relationships are contentious, we start with a historical overview and then go on to explore the relations within the fold-and-thrust belt prior to examining plutonism with each belt. Those readers not concerned with the details of the two deformational events can simply jump ahead to the appropriate sections where the plutons related to each belt are described and interpreted.

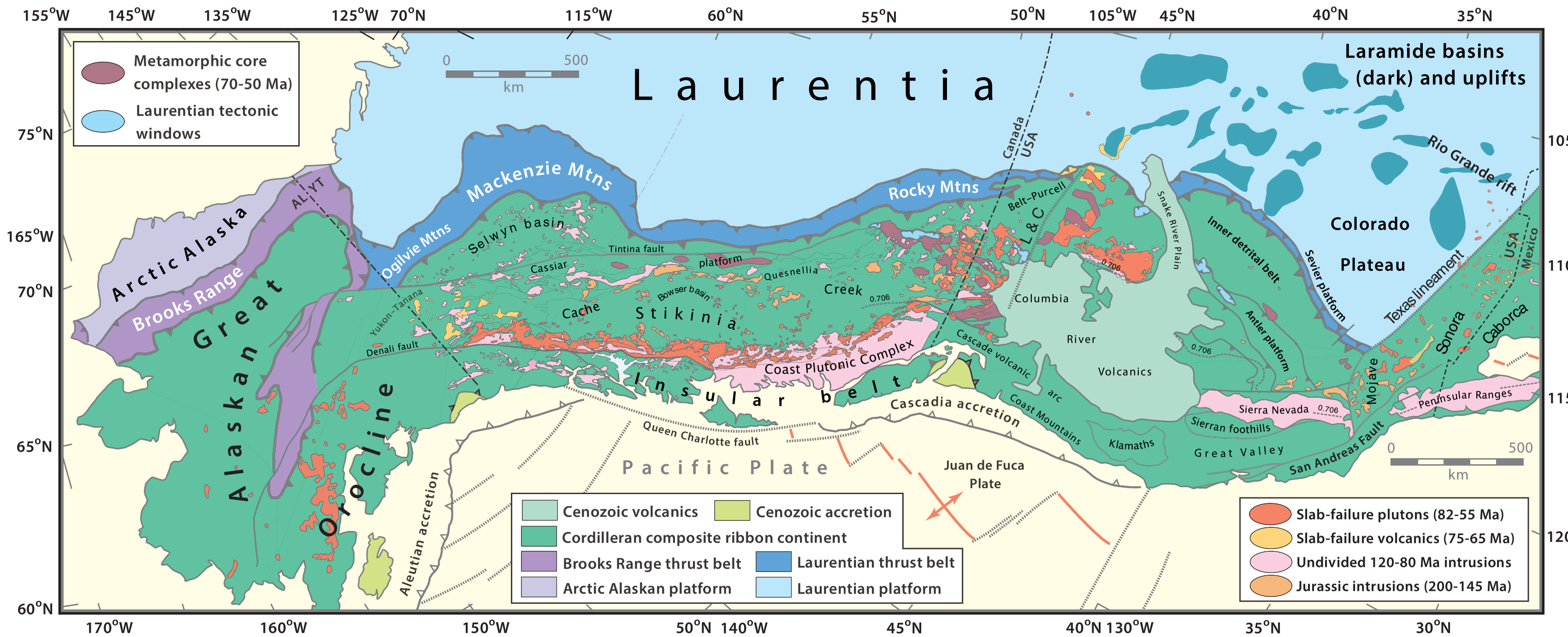


Figure 39. Generalized sketch map illustrating the hypothesized extent of the Cordilleran ribbon continent in central and northern North America, modified from Hildebrand (2014). AL—Alaska; YT—Yukon Territory; L & C line—Lewis and Clark lineament.

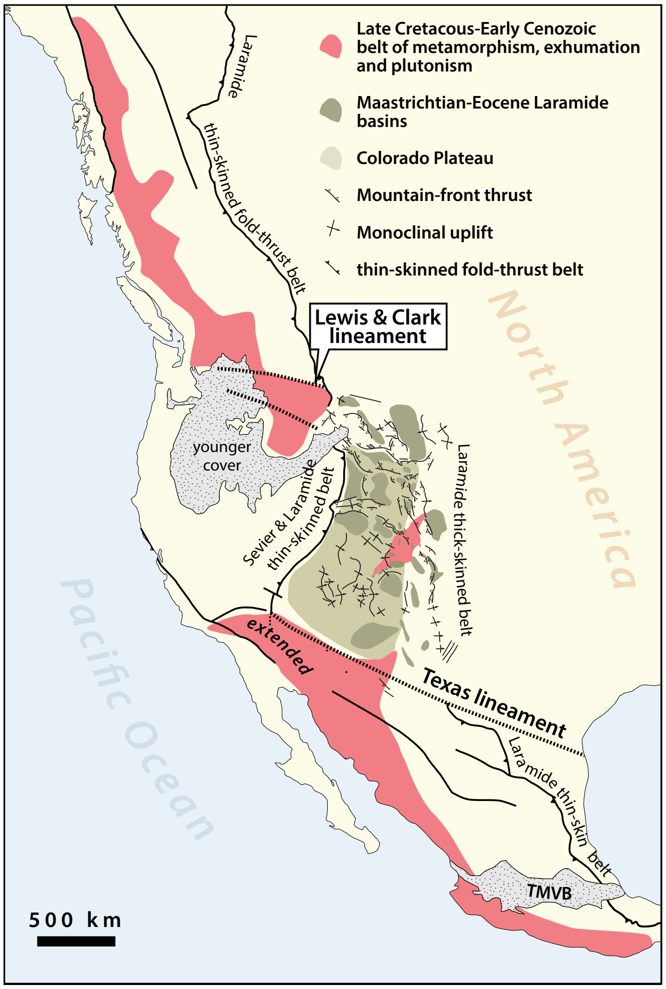


Figure 40. Sketch map showing location of thin-skinned and thick-skinned thrust belts. Thick-skinned folds and faults of Rocky Mountain foreland are from D.M. Miller et al. (1992). The two lineaments that Hildebrand (2015) used as piercing points are shown in their present positions. Note that when they are restored together, the two belts of Late Cretaceous–early Cenozoic magmatism form a continuous belt. TMVB—Trans-Mexican volcanic belt.

No doubt early trappers, such as John Coulter and Ezekial Williams, both of whom found and traversed the central Rockies (Goetzmann, 1966), wondered how the mountains and their adjacent "wintering-holes" formed, but the subject was not addressed in the literature until examination of the west by U.S. government scientists after the Civil War. In his classic report on the geology of the 40th parallel, Clarence King (1878) named the uppermost Cretaceous sedimentary rocks the Laramie and recognized that their deposition was followed by (p. 357) "an energetic orographic disturbance, which closed the Mesozoic age" (p. 357). To the north, the pioneering Canadian geologist, G.M. Dawson, separated the Laramie, which he found to be Paleocene, from the Cretaceous and noted that a belt of "flexed and disturbed" rocks extended the length of the Canadian Rockies (Dawson, 1885, 1886). In a perceptive analysis, James Dana (1890) argued that the Rocky Mountain deformational axis split into two parts along an eastward bend in Montana and Wyoming but that (p. 187) "the great bend of the protaxis is passed by the Paleozoic and Cretaceous formations without essential change of characteristics either in kinds of rocks, or in their disturbed condition, or in time of disturbance" (p. 187). He also noted that the Colorado Front Ranges, and related mountains of the eastern Rockies, formed later and within the area of Mesozoic subsidence, whereas the western mountains formed outside it to the west. Subsequently, in his hugely influential *Manual of Geology*, Dana (1896), noted that the (p. 874) "close of Mesozoic time was marked by the making of the greatest of North American Mountain systems" (p. 874), which he called the Laramide system. Thus, by the close of the nineteenth century, geologists understood in a general way that there were two distinct regions characterized by deformations different in age and structural style.

SEVIER FOLD-AND-THRUST BELT

Nearly 100 yr later, Armstrong (1968) followed up the nineteenth-century geological concepts with a landmark paper in which he clearly and unambiguously separated miogeosynclinal sedimentary rocks and "eastward overturning and thrusting" in the thin-skinned belt from younger thick-skinned deformation of the carbonate-shelf–dominated Rocky Mountain foreland (p. 434). He coined the term "Sevier orogenic belt" for the belt of thin-skinned folds and thrusts based in part on previous work by Harris (1959), who argued that orogenic deposits of the foreland were derived from a broad arch-like uplift—not thrusts—located farther west.

While this simple model has much to commend it, researchers have commonly considered all northerly trending, easterly vergent thrust faults of the region to have formed during the Sevier orogeny, despite the understanding that thin-skinned deformation in the fold-and-thrust belt in both Canada and Mexico is the same age as the younger thick-skinned Laramide deformation of the Rocky Mountain foreland (Armstrong, 1974). Also, over the years, geologists have tended to ascribe various deformational features to either the Laramide or Sevier orogeny based almost entirely on tectonic style, as opposed to age. For example, scientists working in the Great Basin region to the west of the Sevier fold-and-thrust belt recognized that there are many thin-skinned thrust faults, both older and younger, than those in the narrow thrust belt defined by Armstrong (1968), which led some workers to hypothesize a west to east progression of thrusting across Nevada from the ca. 160 Ma Luning-Fencemaker thrust belt to the Sevier of central Utah and western Wyoming (DeCelles, 2004; DeCelles and Coogan, 2006; Yonkee and Weil, 2015). There are several problems with this concept. First, the Luning-Fencemaker fold-and-thrust belt (Wyld et al., 2003; Wyld and Wright, 2009) appears to mark the collision zone between the latest Triassic–Early Jurassic Black Rock arc terrane of northwestern Nevada (Quinn et al., 1997; Wyld, 2000) and a west-facing carbonate-dominated passive margin (Oldow, 1984; Hildebrand, 2013) developed atop Triassic Koipato volcanics and their subjacent Golconda basement. Second, thrusts within the Garden Valley thrust system (Bartley and Gleason, 1990), part of the Central Nevada thrust system (Speed et al., 1988), located midway between the Luning-Fencemaker and Sevier systems, are cut by Cenomanian to Coniacian plutons (Taylor et al., 2000) and so may be more closely related to the ca. 100 Ma Oregonian event, as suggested by Wells (2016).

Within the more narrowly defined fold-and-thrust belt, the age of Sevier thrusting is reasonably well constrained by a number of detailed sedimentological and paleontological studies. The earliest record of sedimentation that can be tied to disturbances anywhere along the entire western margin of North America occurred in the Aptian at ca. 124–115 Ma, when gravels and conglomerates, holding a wide variety of sedimentary clasts such as chert, quartzite, limestone, and siltstone, were eroded from the Rocky Mountain platform and dispersed eastward to form a thin veneer over a regional unconformity and a calcrete-silcrete paleosol complex on the Upper Jurassic Morrison Formation and its equivalents (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 Cadomin, Kootenai, Lakota, Cloverly, Ephraim, Buckhorn, Pryor, and so on, and they are extensive (Heller et al., 2003), occurring up and down the continent, eastward to the Black Hills of South Dakota, and as far south as Silver City, New Mexico, where the gravels appear above a basal mudstone in the lower part of the Beartooth quartzite (Hildebrand et al., 2008). The gravels are overlain by marine mudstones and siltstones of Albian and Cenomanian age, which mark the first sedimentary rocks of the Western Interior Basin (Kauffman, 1977). Modern analytical techniques better constrain the initiation of Sevier thrusting to be 124–120 Ma, using dated ash beds in the basal foredeep of Utah (Greenhalgh and Britt, 2007), detrital zircons in the foredeep (Britt et al., 2007), a 119.4 ± 2.6 Ma U-Pb date of uraniferous carbonate from the forebulge, and a good match between $\delta^{13}C_{org}$ excursions in early terrestrial foredeep sedimentary rocks and well-dated Albian features of the global carbon isotope chemostratigraphic record (Ludvigson et al., 2010).

Major thrusting of the Sevier event stopped before or during the mid-Albian, as documented by alluvial-fan and fluvial sedimentary rocks of the Canyon Range wedge-top basin, which unconformably overstepped the Canyon Range thrust (Lawton et al., 2007). Farther north, within the Wyoming salient, conglomerates of the 120–110 Ma Gannett Group recorded major slip on huge thrust sheets (DeCelles et al., 1993; DeCelles and Cavazza, 1999; Yonkee and Weil, 2011). There, zircon fission-track and $^{40}Ar/^{39}Ar$ ages of muscovite, plus westward-coarsening conglomerates of the Bear River and Aspen formations in the salient, indicate that the thrust sheets were uplifted and exhumed mainly during the Cenomanian–Turonian (Yonkee and Weil, 2011, 2015). Hildebrand (2013) suggested that thrusting stopped and exhumation began when the westward-dipping subducting slab attached to North America failed.

A mainly early to mid-Campanian thin-skinned fold-and-thrust belt lies farther east (DeCelles 1994; Yonkee and Weil, 2011). In western Wyoming and north-central Utah, these are thrusts such as the Crawford, Absaroka, Darby, and Hogsback, which all root in a detachment within the Cambrian, whereas the older thrusts farther west detached deep within the Precambrian section. Farther south, thrusts such as the Gunnison of central Utah and the Keystone thrust of southern Nevada are likely of the same general age. This belt of thrusts is separated from the western belt in northern Utah and southern Idaho by the Wasatch anticlinorium, a regional culmination cored by allochthonous Paleoproterozoic crystalline basement (Yonkee et al., 1997, 2000; Yonkee and Weil, 2011; Shervais et al., 2013). On the basis of its much younger age, the eastern fold-and-thrust belt is interpreted to have formed during the younger Laramide event. This interpretation—different from that made by most workers, who consider that the Sevier event includes both an older Aptian to mid-Albian phase in the west and a younger phase of Late Cretaceous–Paleocene shortening to the east (DeCelles, 2004; Yonkee and Weil, 2011)—is based upon the cessation of thrusting in the west before 105 Ma, the 100–90 Ma exhumation of the western (Sevier) thrust belt, and the well-defined 15–20 m.y. gap between thrusting in the western and eastern thrust belts.

Deformation and sedimentation related to the Sevier event were apparently confined to the region between the Lewis and Clark lineament and the southern Colorado Plateau: They are unknown in platformal rocks of western North America to the north in Canada or south in Mexico. This suggests that plate geometries were complex, possibly involving a promontory on the ribbon continent, and that oceanic crust remained between North America and the ribbon continent both to the north and south. Nevertheless, following the Sevier event, at least parts of the ribbon continent were attached to North America and were therefore neoautochthonous.

SEVIER HINTERLAND

In the Great Basin region to the west of the fold-and-thrust belt, neither deformation nor metamorphism in the 125–105 Ma range has been reported; instead, compressional deformation occurred in the Jurassic (Elko) and Upper Cretaceous (Laramide), as described by Camilleri et al. (1997). Within the Canadian sector of the orogen, north of the Lewis and Clark line, the reverse situation exists. There, deformation on the North American platform started during the Upper Cretaceous, as documented by the presence of Santonian–Campanian marine shales in the footwall syncline of the Borgeau thrust, which in the southern Canadian Rockies is the westernmost major thrust fault to affect the platform terrace (Price, 2013); however, just to the west, where Paleozoic sedimentary rocks sit atop the metasedimentary rocks of the Mesoproterozoic Purcell Group, thrusts were shown to predate

108 Ma, because they are cut by syn- to postkinematic plutons of the Bayonne suite (Logan, 2002a; Larson et al., 2006). Likewise, within the Monashee Mountains of southern British Columbia (Fig. 31), rocks of the Selkirk allochthon contain three periods of kyanite growth: 162–143 Ma, 130–88 Ma, and 82–60 Ma (Gervais and Hynes, 2013), which are reasonably interpreted to correspond to metamorphism associated with the Elko, Sevier, and Laramide deformational/thickening events. Within the southern Omineca belt (Fig. 31), an undeformed 111 Ma pluton of the Bayonne suite cuts deformational fabrics (Webster et al., 2017). On the basis of the age of the youngest deformed rocks, thrusting within the Selwyn Basin (Fig. 31) to the north occurred after the latest Jurassic and, based on $^{40}Ar/^{39}Ar$ in muscovite from greenschist-grade rocks, had ceased by ca. 104 Ma (Mair et al., 2006).

Using mismatches in geology and robust paleomagnetism of Cordilleran rocks, Hildebrand (2013, 2014, 2015) argued that the western parts of the Sevier belt are located within the Canadian Cordillera and that Cretaceous plutons within the eastward-younging Omineca crystalline belt of Monger et al. (1982) represent rocks emplaced into the upper-plate hinterland during and after the collision of the Cordilleran ribbon continent with North America. A factor implicit in this model is the juxtaposition of two different passive-margin platforms in western North America: (1) the North American platform, formed in the Cambrian and characterized by a Middle Cambrian carbonate reef complex; and (2) a more westerly carbonate platform—known as the Antler platform in the United States and the Cassiar platform in Canada—that formed during the Neoproterozoic and that contains a conspicuous Early Cambrian carbonate bank dominated by *Archeocyathids* and oolite shoals (Johnston, 2008; Hildebrand, 2009, 2013). The western platform, which sits atop 8–12 km of coarse Late Precambrian clastic sedimentary rocks, was involved in several orogenic events: (1) initially during the latest Devonian–early Mississippian, when rocks of the Roberts Mountain allochthon were thrust upon it (Roberts et al., 1958; E.L. Miller et al., 1992); (2) between 260 and 253 Ma, when the Slide Mountain ocean was subducted during the collision of the Yukon-Tanana terrane (Murphy et al., 2006; Beranek and Mortensen, 2011); (3) at 250 Ma, when the Golconda allochthon was emplaced on the western margin of the Roberts Mountain allochthon (Silberling and Roberts, 1962); and (4) at both 187–185 Ma and 173 Ma, when the Kootenay terrane, which contains the Cassiar platform, or equivalents, in southern Canada, collided with Quesnellia to the west (Nixon et al., 1993) and then with rocks of the Belt-Purcell-Windermere block to the east, respectively (Murphy et al., 1995; Colpron et al., 1996). In comparison, the North American platform remained undeformed until the ca. 122 Ma Sevier event.

After the Sevier event, rocks of the Canadian Cordillera, formerly part of the Cordilleran ribbon continent, were located much farther south and formed neoautochthonous North American crust (Hildebrand, 2015). On the basis of temporal relations, the Bisbee-Arperos seaway remained open during the event and closed several million years later at ca. 100 Ma. We now look at syn- to postdeformational plutonic rocks of the proposed Sevier hinterland belt preserved in the Omineca crystalline belt of the Canadian Cordillera.

Plutons of the Sevier Event

A >1000-km-long band of 120–100 Ma, syn- to posttectonic plutons occurs within the Omineca crystalline belt (Fig. 31), the eastern of the two structural-metamorphic welts within the Canadian Cordillera (Monger et al., 1982; Logan, 2002a, 2002b). In the north (Fig. 31), the magmatism gets younger eastward, and plutons in the 100–90 Ma range intrude metasedimentary rocks within the Selwyn Basin (Hart et al., 2004a, 2004b; Johnston, 2008; Rasmussen, 2013). The mismatched deformation ages, paleomagnetic data, and piercing points led Hildebrand (2013, 2014, 2015) to argue that the Omineca belt represents the hinterland to the ~122 Ma Sevier event of the western United States, and that after 70 Ma, it was transported northward, along with most of the Canadian Cordillera, on faults located within the Cordilleran fold-and-thrust belt. Thus, because the plutons within the Omineca crystalline belt and Selwyn Basin are syn- to postdeformational with respect to the Sevier event, many might be slab failure rocks.

Many of the plutons and their wall rocks host economic and subeconomic mineral deposits (Fig. 41), and so they have been studied in some detail (Newberry et al., 1990, 1996; Driver et al., 2000; Brandon and Lambert, 1993, 1994; Hart et al., 2004a, 2004b; Logan, 2002a, 2002b; Mortensen et al., 2007; Morris and Creaser, 2008; Mair et al., 2011; Rasmussen, 2013). In many cases, U-Pb zircon ages, trace-element concentrations, and radiogenic isotope ratios are available, although many plutons remain unstudied.

Plutons of the northern Omineca belt and Selwyn Basin are divided into several suites based on geography, age, and composition (Fig. 41). Similar plutons occur in Alaska near Fairbanks and together with their Canadian counterparts form what is known as the Tintina Gold Belt (Smith, 2000; Gough and Day, 2007; Goldfarb et al., 2000, 2007). The following descriptions are summarized from the previously cited publications, but mainly those of Rasmussen (2013) and Hart et al. (2004a, 2004b).

The oldest suite of plutonic rocks is the 115–104 Ma high-K calc-alkaline Cassiar suite, which mainly intrudes rocks of Cassiar platform (Fig. 41). They range from muscovite-biotite– to biotite- and hornblende-bearing plutons (Rasmussen, 2013). K-feldspar megacrystic biotite ± muscovite monzogranite, ranging in age from 114 to 111 Ma, and with associated late-stage dikes containing tourmaline and beryl, constitute the Anvil intrusive suite. The 106–100 Ma Hyland suite intrudes rocks of both the western and eastern Selwyn Basin and constitutes biotite granodiorite and monzogranite. Tay River plutons intrude rocks of the Selwyn Basin and are biotite-hornblende ± clinopyroxene granite and granodiorite ranging in age from 99 to 96 Ma, which overlap with nine 98–97 Ma caldera-forming eruptions of ignimbrite collectively grouped with mafic and siliceous lavas as the South Fork volcanics.

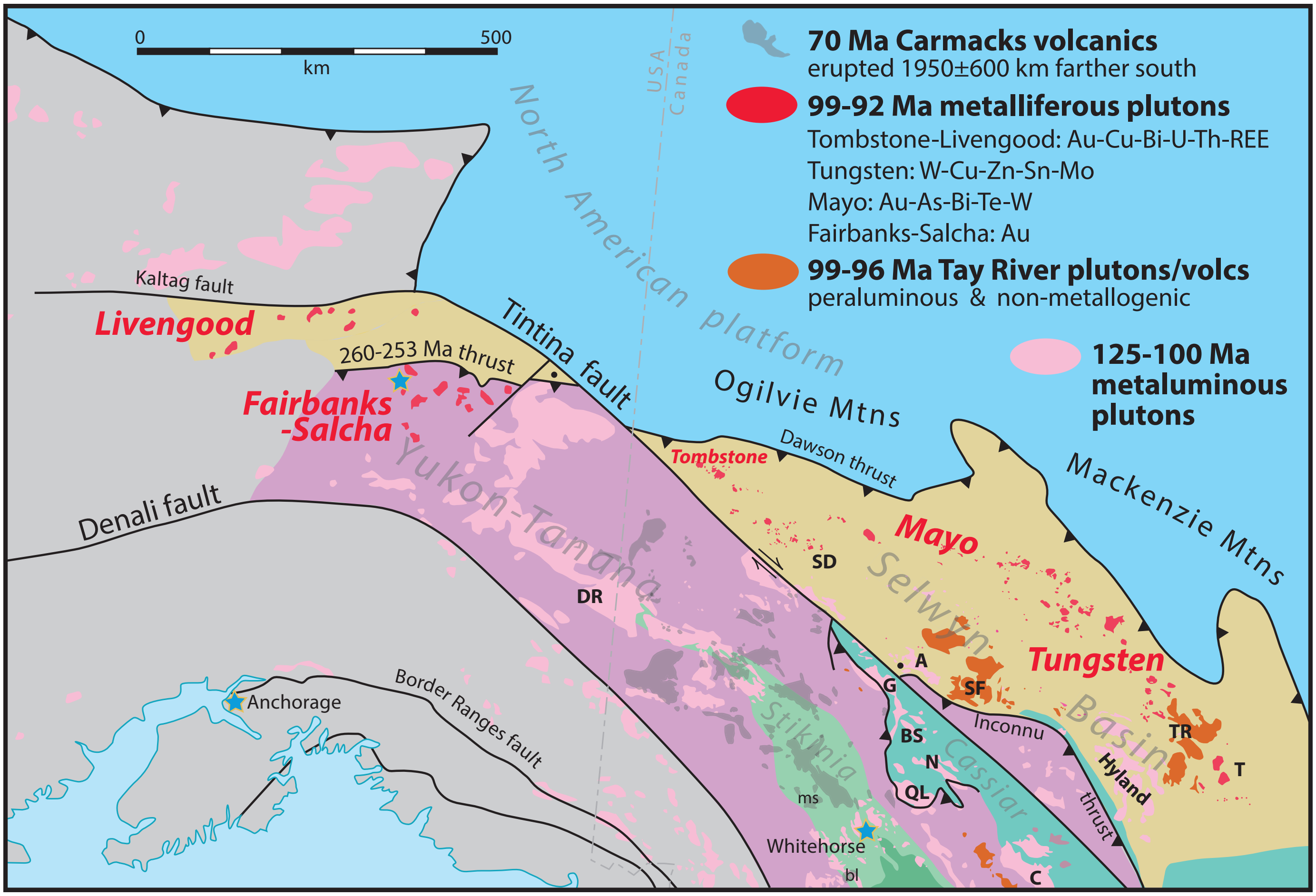


Figure 41. Geological sketch map showing the distribution of post-Sevier metalliferous slab failure plutons now located within northern Canada and Alaska (Hart et al., 2004a, 2004b; Rasmussen, 2013) interpreted to result from failure of the west-dipping North American oceanic plate (Hildebrand, 2013, 2014). Note the abundant remnants of 70 Ma Carmacks volcanic rocks, which in multiple robust studies yielded paleolatitudes of present-day San Francisco (Johnston et al., 1996; Enkin et al., 2006a). As rocks of the Selwyn Basin, Cassiar platform, and Yukon-Tanana terrane were joined together at 260–253 Ma (Beranek and Mortensen, 2011), they are all interpreted to have been located within the Great Basin–Sonora segments of the orogen at 70 Ma and migrated to their present northerly latitude mostly during the early Cenozoic (Hildebrand, 2014, 2015). Dots adjacent to the Tintina fault are possible prefault piercing points (Gabrielse et al., 2006). Other Late Cretaceous–Paleocene magmatic complexes commonly have associated mineralization and, along with the Carmacks, are probably slab failure rocks related to the accretion of Wrangellia to the western margin of the neoautochthonous Intermontane superterrane during the Laramide event. DR—Dawson Range, SD—Scheelite Dome, A—Anvil, SF—South Fork volcanics, TR—Tay River, T—Tungsten, G—Glenlyon pluton, BS—Big Salmon pluton, N—Nisultin pluton, QL—Quiet Lake pluton, C—Cassiar batholith; Eocene calderas: bl—Bennett Lake; ms—Mount Skukum. REE—rare earth elements.

Three suites intrude the Selwyn Basin farther to the east. The 98–95 Ma Mayo suite consists of subalkaline, metaluminous to weakly peraluminous, titanite-bearing, hornblende-biotite granodiorite and monzogranite. Plutons of the suite have associated Au plus As, Bi, Te, and W mineralization. Plutons of the 94–91 Ma Tombstone suite are commonly zoned composite bodies of alkalic, magnetite- and titanite-bearing biotite monzogranite, syenite, and peraluminous leucogranite with sparse bodies of granite, granodiorite, hornblende diorite, pyroxenite, and gabbro. The Mayo and Tombstone suites have equivalent Au-bearing rocks in northeastern Alaska (Fig. 41), where they are known as the Fairbanks-Salcha and Livengood suites, respectively (Anderson, 1987; Newberry et al., 1990, 1996; Reifenstuhl et al., 1997a, 1997b). Last, the 98–95 Ma plutons of the Tungsten suite are weakly porphyritic, ilmenite-bearing, peraluminous muscovite-biotite granite to granodiorite containing abundant xenocrystic zircons (Hart et al., 2004a, 2004b). They have associated tungsten skarns and Cu, Zn, Sn, and Mo anomalies. Overall depth of emplacement for plutons of the suite is 1–2.5 kbar (Hart et al., 2004a, 2004b).

In the southern Omineca belt, 115–90 Ma, syn- to post-tectonic plutons are collectively known as the Bayonne suite (Archibald et al., 1984; Logan, 2002a). Plutons of the suite (Fig. 31) consist mostly of slightly peraluminous, subalkaline, hornblende-biotite granodiorite and strongly fractionated two-mica granites, aplites, and pegmatites (Brandon and Lambert, 1993, 1994). Several of the plutons have associated Au-W-Bi-As quartz-vein–type deposits similar to deposits of the Tintina Gold Belt (Logan, 2002b).

We plotted trace-element concentrations and isotopic data for a large number of plutons emplaced within the Omineca belt on our standard discrimination plots (Figs. 42–45). As expected for plutons emplaced into an orogenic hinterland during and immediately after a collisional event, the trace elements and isotopic ratios are more typical of slab failure magmatism than arc magmatism.

There is a much broader spread of isotopic values for this suite than for the Oregonian slab failure suite, but each group in the Sevier suite has members that plot in the field of Oregonian plutons (Fig. 45) and then spread to form an array of more-enriched values. When examined in detail, several of the groups have large ranges in isotopic values at the same SiO_2 content (Fig. 46). For example, individual bodies of the

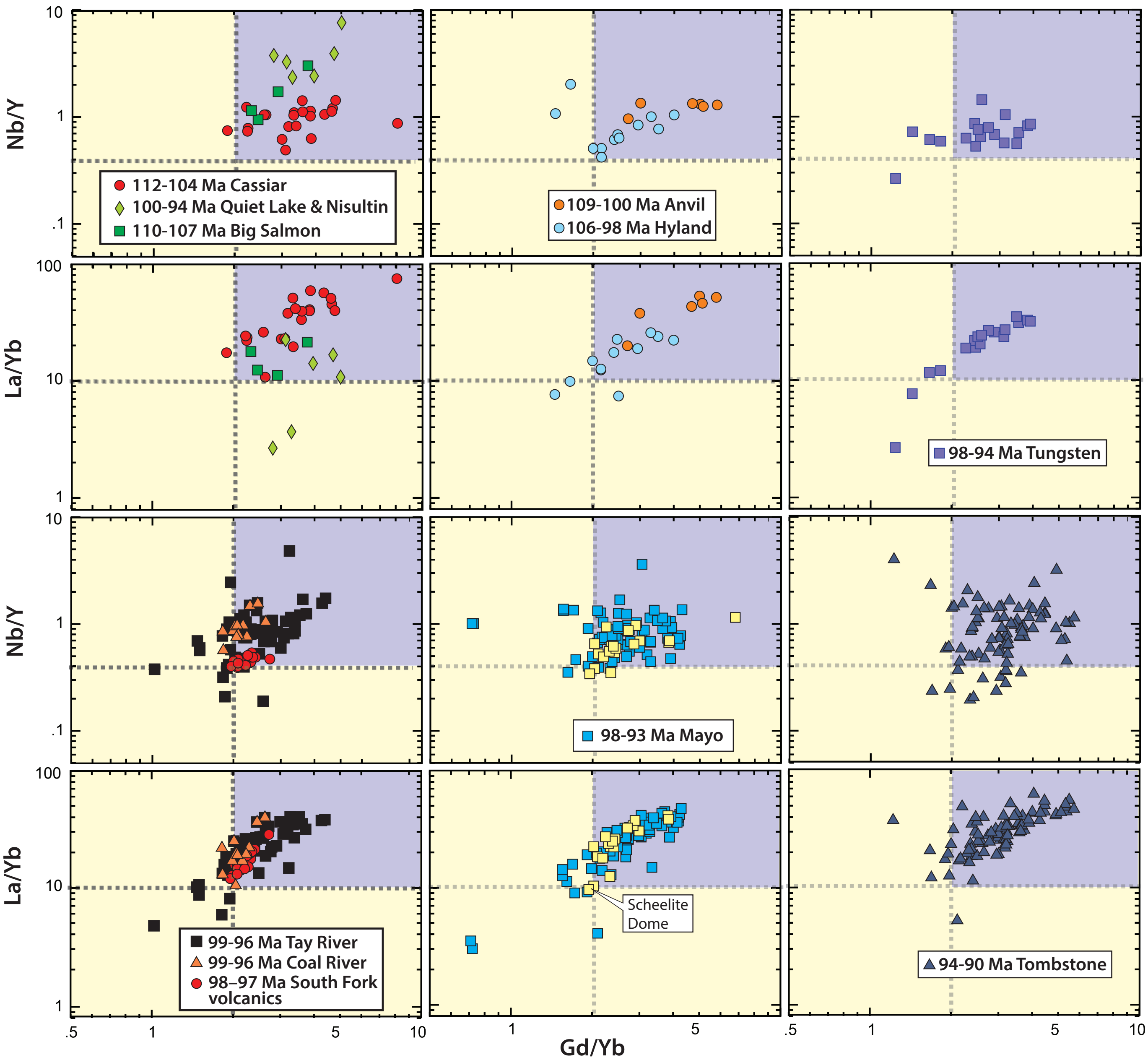
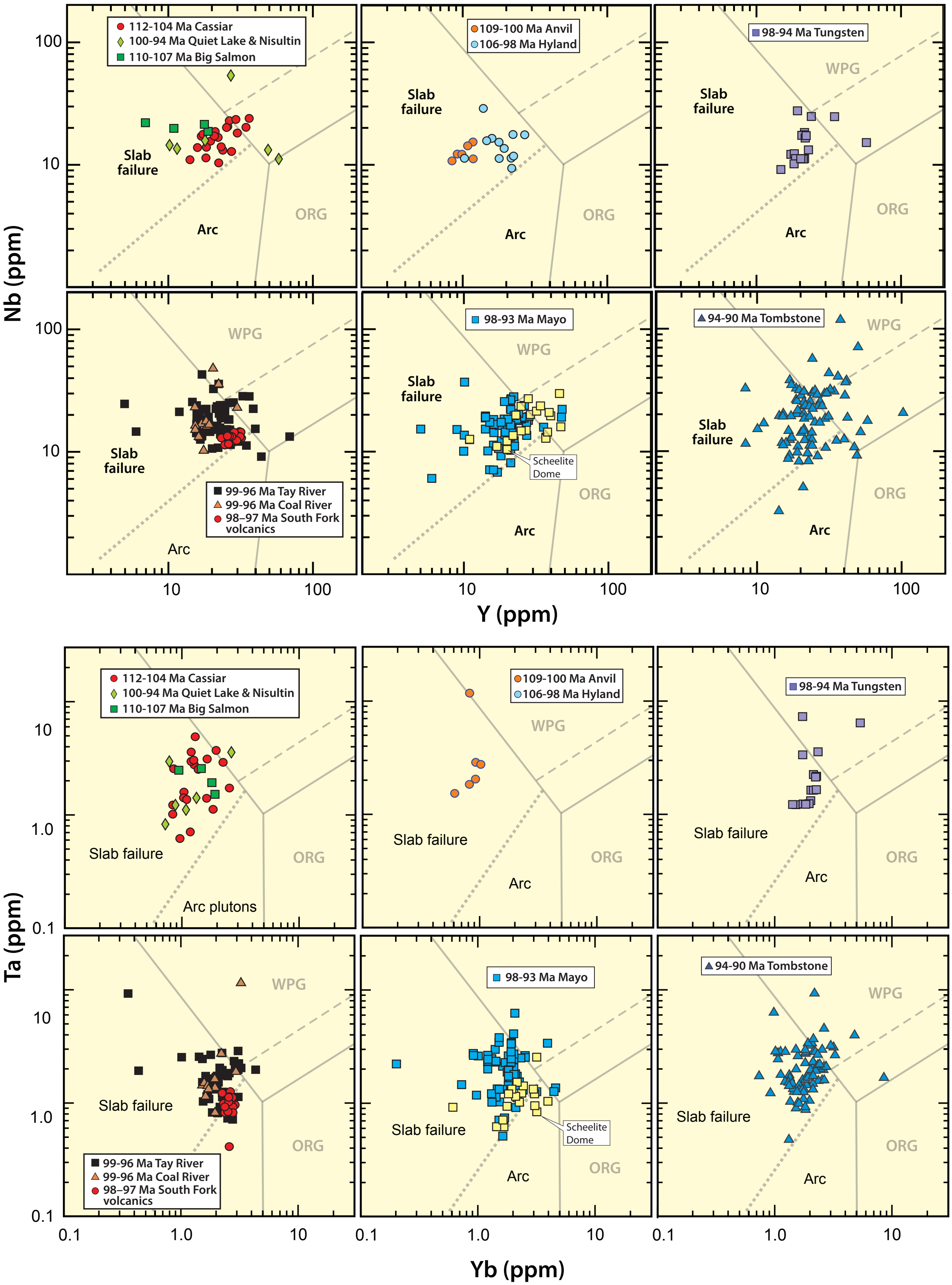


Figure 42. Syn- to posttectonic magmatic suites within the northern Omineca belt and Selwyn Basin plotted on our Nb/Y and La/Yb vs. Gd/Yb discrimination plots. All of the suites appear to be of slab failure origin. See Figure 41 for locations of plutons and suites. Data are from Rasmussen (2013), Pigage et al. (2014), Driver et al. (2000), and Morris and Creaser (2008).



112-104 Ma Cassiar
100-94 Ma Quiet Lake & Nisultin
110-107 Ma Big Salmon
109-100 Ma Anvil
106-98 Ma Hyland
98-94 Ma Tungsten
99-96 Ma Tay River
99-96 Ma Coal River
98–97 Ma South Fork volcanics
98-93 Ma Mayo
94-90 Ma Tombstone
Scheelite Dome
Slab failure
Arc
Arc plutons
WPG
ORG
Nb (ppm)
Y (ppm)
Ta (ppm)
Yb (ppm)

Figure 43.

Figure 43. Samples from syn- to posttectonic plutons within the northern Omineca belt and Selwyn Basin plotted on our Nb vs. Y and Ta vs. Yb discrimination diagrams, illustrating their slab failure affinity. Scheelite Dome is a mineralized plutonic member (intrusion-related gold) of the Mayo suite (data from Mair et al., 2011). Data are from Rasmussen (2013), Pigage et al. (2014), Driver et al. (2000), and Morris and Creaser (2008). WPG—within-plate granite; ORG—ocean-ridge granite.

Tombstone group range over 12 epsilon Nd units at 66%–68% SiO_2 (Fig. 46).

The data from several plutons of the Bayonne suite in the southern Omineca belt show considerable scatter on the diagrams, especially with respect to La/Yb ratios, so we looked into the possibility of alteration in a few examples by color coding for SiO_2 content (Fig. 47). As there is little to no change in Yb contents, the strong correlation between increasing SiO_2 and decreasing La/Yb is readily interpreted as the result of removal of La by a fluid component during alteration.

We also plotted isotopes of Nd and Sr against SiO_2 for various members of the Sevier suite and found that the higher-SiO_2 samples also showed extreme variability in isotopic values, as well as correlating with La/Yb (Fig. 48). These variations are difficult to derive from any sort of fractionation scheme and instead are best explained by high-level fluid-rock interaction with Proterozoic and early Paleozoic wall rocks. Whether the interaction involved hydrous crustal melts (Reiners et al., 1996) or simply aqueous fluids is unimportant here, because the rocks are not representative of the original magmatic composition. Therefore, this demonstrates that epizonal plutonic rocks with greater than ~70% SiO_2 are generally unreliable predictors of source region and tectonic setting.

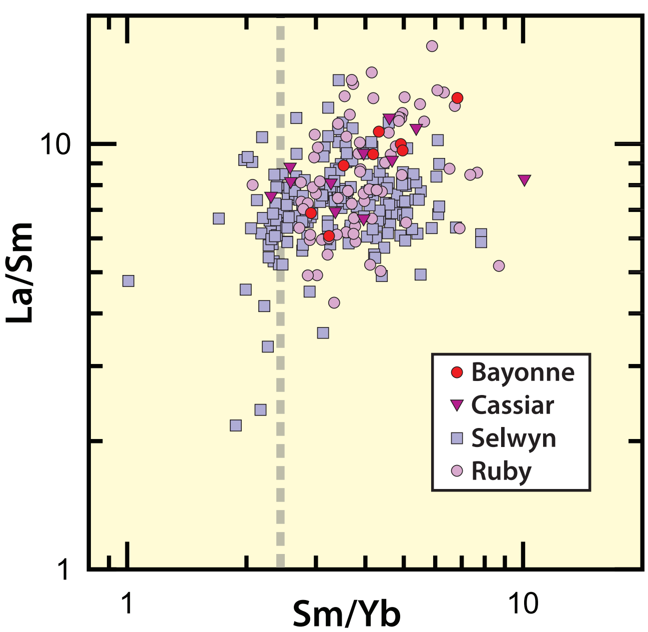


Figure 44. Rocks of the Bayonne, Cassiar, Selwyn, and Ruby plutonic suites with <70% SiO_2 plotted on our La/Sm vs. Sm/Yb discrimination plot. The dashed line at Sm/Yb = 2.5 represents our dividing line between dominantly arc rocks with low Sm/Yb and slab failure rocks with ratios higher than 2.5 (see Fig. 10).

LARAMIDE EVENT

The Laramide event is a Late Cretaceous–early Cenozoic deformational event that appears to have affected rocks from Tierra del Fuego to Alaska (Hildebrand and Whalen, 2014a). Deformation was both thin- and thick-skinned deformation, and thrust faults are dominantly easterly vergent. Here, we focus on the North American sector, which not only includes thrust and strike-slip faults within both the Cordilleran fold-and-thrust belt and Rocky Mountain foreland, but also an exhumed metamorphic hinterland riddled with largely postdeformational plutons. We first address the traditional Laramide thick-skinned deformation within the Rocky Mountain foreland, as most geologists consider this the essence of the Laramide event, or orogeny.

Thick-Skinned Deformation

The Rocky Mountain foreland of west-central North America contains a number of spectacular mountain ranges and adjacent syntectonic basins filled by alluvial and lacustrine deposits. The structure is variable, ranging from huge crystalline massifs hundreds of kilometers long in the Rocky Mountains to enormous monoclinal flexures on the Colorado Plateau. The uplifts trend north-south, northwest-southeast, or east-west. The deformation is thick-skinned, as it involved cratonic basement. Overall, the deformation is commonly referred to as "Laramide style," as opposed to deformation on thin-skinned thrusts, which do not generally involve crystalline basement.

The deformation occurred mainly during the Maastrichtian–Eocene (Dickinson et al., 1988; Lawton, 2008) and is distinctly younger than the majority of Laramide thrusts in the Cordilleran fold-and-thrust belt directly to the west, where leading early Campanian thrusts were beveled and buried by huge west-derived alluvial fans by the mid-Campanian (DeCelles and Cavazza, 1999). However, Laramide deformation in the Rocky Mountain foreland occurred at the same time as some of the thin-skinned thrusting to the north in Canada and to the south in Mexico (Armstrong, 1974), a feature recognized for over 100 yr (Blackwelder, 1914).

In spite of the early advances, without understanding the driving mechanisms, the nature and origin of the Laramide thick-skinned deformation have remained enigmatic and, therefore, factious. Early controversy revolved around whether the basement-cored uplifts were the products of vertical or horizontal tectonics (Thom, 1923; Chamberlin, 1945; Eardley, 1962; Lovering, 1929; Woodward, 1976). More recently, after collection and analysis of drilling and seismic data, the pendulum swung toward

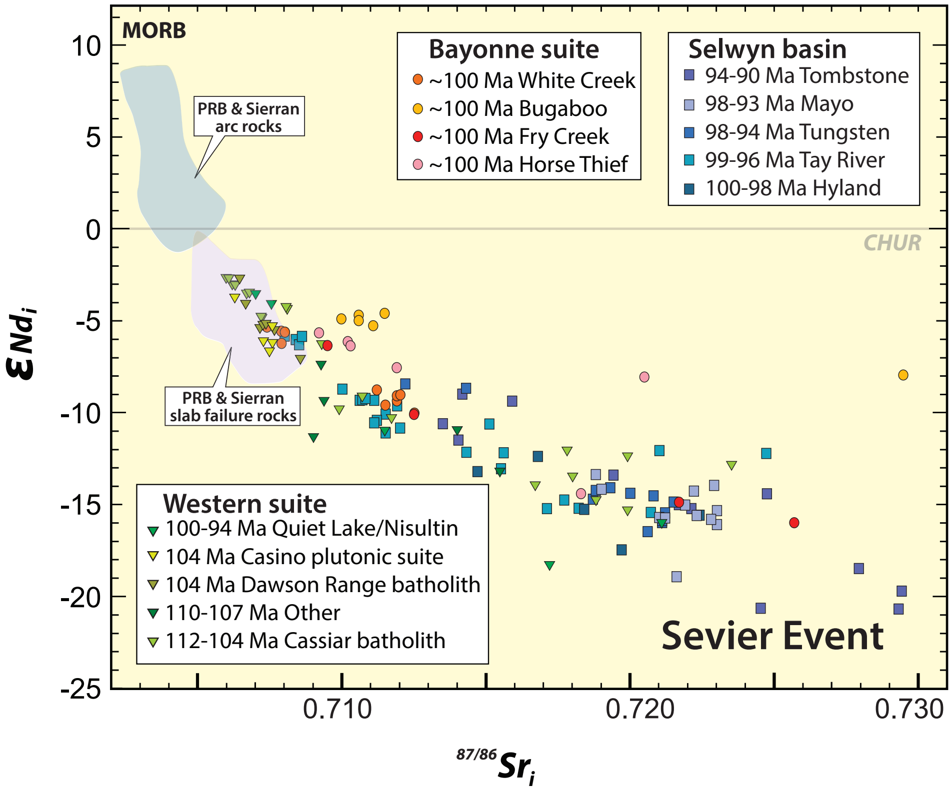


Figure 45. ε_{Nd} vs. $Sr_{initial}$ plot for 112 to 90 Ma plutonic rocks of likely slab failure plutons within the Omineca belt and Selwyn Basin. Included are rocks of the Bayonne suite from the southern Omineca belt and suites of the northwestern Omineca belt plus the various Selwyn Basin intrusive suites. Fields for samples from the Peninsular Ranges batholith (PRB) and Sierran arc and slab failure suites are shown for comparison (see Fig. 23). Note that each suite has members that plot within or near the field of <100 Ma slab failure plutons in the Peninsular Ranges (PRB) and Sierran batholiths, yet they have huge spreads in Nd-Sr space for other members of the suites. Data sources are as in previous figures. See Figure 41 for locations of suites. MORB—mid-ocean-ridge basalt; CHUR—chondritic uniform reservoir.

horizontal compression as the main driving force (Berg, 1962; Gries, 1983; Sprague, 1983; Smithson et al., 1979).

With the development of plate tectonics in the late 1960s (Frankel, 2012), scientists soon developed models utilizing plate interactions. Lipman et al. (1971) used the age, composition, and distribution of Cenozoic volcanic rocks to suggest shallow subduction beneath the western United States. They were followed by Lowell (1974) and Coney and Reynolds (1977), who summarized and interpreted radiometric data from volcanic rocks of California and southern Arizona, which suggested to them that arc magmatism swept inboard 1000 km at 80 Ma and then back again at ca. 45 Ma due to progressive slab shallowing followed by steepening.

Even though others (Glazner and Supplee, 1982) presented contrary data sets that challenged the shallow subduction model, nearly all subsequent workers developed models in which the Laramide event was connected to shallowly dipping subduction beneath North America (Dickinson and Snyder, 1978; Bird, 1988; Hamilton, 1988; Dumitru et al., 1991; Grove et al., 2003; Jacobson et al., 2011; Jones et al., 2011; Chapin, 2012; Fan and Carrapa, 2014), but 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 suggested its continued presence (Farmer et al., 1989; Livaccari and Perry, 1993; Lee et al., 2000, 2001a). This led Saleeby (2003) to develop a segmented model in which only a narrow region of the subducting plate was shallow.

In a variant of the flat-slab model, some scientists hypothesized that the Laramide deformation was 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 variant (Liu et al., 2010) involves slightly different-aged collisions of hypothetical conjugates to the Hess and Shatsky Rises, now located in the NW Pacific region (Sager, 2005).

While the complete subduction of the hypothesized collisional plateaux precludes any direct tests, Hildebrand (2013, p. 99) pointed out that: (1) there are several such collisions taking place today and that, on the basis of studies by others (Miura et al., 2004; Phinney et al., 2004; von Huene and Ranero, 2009), subduction of plateaux is unlikely to produce deformation akin to that observed in the Rocky Mountain foreland; and (2) recent studies of flat subduction and magmatism beneath Mexico and in southern Alaska indicate that flat slabs are able to transport volatiles well inboard of the normal arc-trench distance and still produce copious quantities of arc magmatism (Blatter et al., 2007; Neal et al., 2001; Richter et al., 1995, 2006; Trop et al., 2012), so that shallow subduction may not be a viable explanation for lack of magmatism in the region of thick-skinned deformation.

The intense focus by the geological community on the thick-skinned deformation and its relations to flat-slab subduction generally ignored several important regional observations: (1) thick-skinned deformation of the Rocky Mountain foreland was approximately coeval with the thin-skinned deformation

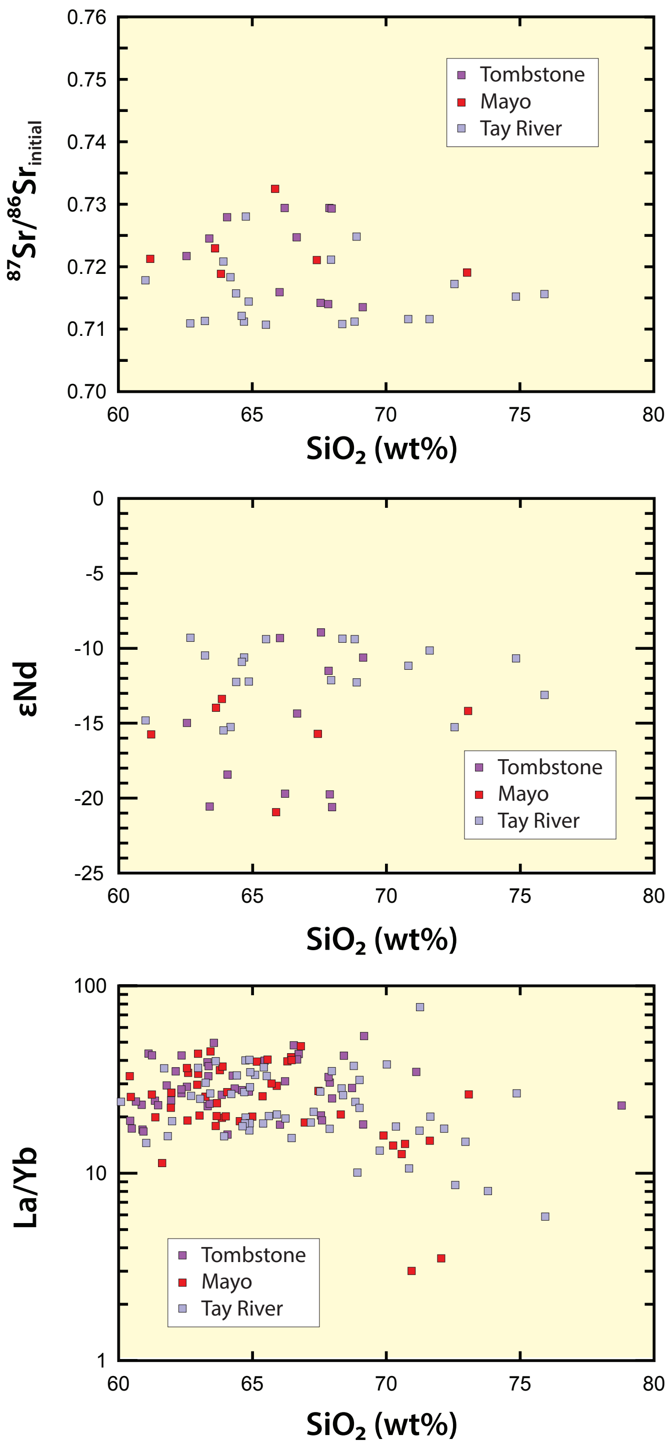


Figure 46. Rocks of the Tombstone, Mayo, and Tay River intrusive suites of the Selwyn Basin area shown on $^{87}Sr/^{86}Sr_{initial}$, ε_{Nd}, and La/Yb vs. SiO_2 plots. The plots show that extensive variation in ε_{Nd} and $Sr_{initial}$ does not correlate with silica, and that La/Yb, a key index of residual garnet in slab failure magmas, exhibits scatter at >70% SiO_2, which we interpret to be due to fractionation and/or fluid-related alteration (see Fig. 47).

north of the Lewis and Clark lineament within the Canadian Cordillera, and south of the Texas lineament within the Mexican sector of the orogen (Armstrong, 1974); (2) the direction of compression changed progressively with time from nearly E-W to N-S (Gries, 1983); (3) the large-scale arrangement of basins and uplifts along the east side of the Colorado Plateau in Colorado and New Mexico is en echelon, which Chapin and Cather (1983) suggested provides evidence for right-lateral slip, during which the plateau moved as many as 120 km north-northeasterly; (4) the Laramide foredeep migrated northward from the Campanian to Paleocene (Catuneanu et al., 2000; Roberts and Kirschbaum, 1995); (5) a linear band of intense Campanian deformation, exhumation, and plutonism occurs farther west, both north of the Lewis and Clark lineament and south of the Texas lineament, in what we believe represents a dismembered orogenic hinterland; and (6) an early to mid-Campanian (82–75 Ma) thin-skinned thrust belt is located east of the older Sevier fold-and-thrust belt. It is worth reiterating that thin-skin thrusting in this belt was more or less coeval with deformation, metamorphism, and plutonism within the proposed orogenic hinterland and stopped by the mid-Campanian, when the leading edge of the thrust belt was eroded and, along with rocks of the adjacent foredeep, buried by conglomerate and gravels of fluvial megafans derived from the more interior portions of the thrust wedge (DeCelles and Cavazza, 1999; DeCelles, 2004; Liu et al., 2005).

Hildebrand (2015) proposed a simple model to explain all the recognized features of the Laramide deformation. He matched two large-scale piercing points, the Texas lineament and the Lewis and Clark transverse zone (Figs. 49 and 50)—which both end at the Cordilleran fold-and-thrust belt but from opposite directions—to document 1300 km of northward migration of the Cordillera from 70 to ca. 50 Ma, consistent with robust paleomagnetic studies that indicated the bulk of the older geological elements within the Cordillera were dismembered and transported meridionally over 1000 km along the orogen (Beck and Noson, 1972; Beck, 1991, 1992; Irving, 1979, 1985; Enkin, 2006; Enkin et al., 2006a, 2006b; Housen and Dorsey, 2005; Kent and Irving, 2010; Wynne et al., 1995, 1998). By restoring the orogen to its pre–strike-slip configuration, he showed, not only that most of the present-day Canadian Cordillera originated in the Great Basin sector of the orogen prior to 70 Ma, but also that once meridional migration is restored, a metamorphic-plutonic hinterland of Campanian age, with its postcollisional exhumation and coeval magmatism, was continuous from Mexico to Alaska. Given that the Late Cretaceous magmatic gap in the region of Laramide thick-skinned deformation was perhaps the principal reason to ascribe the thick-skinned deformation to flat-slab subduction (Dickinson and Snyder, 1978; Humphreys, 2009), the reconstruction of the Campanian Laramide hinterland over thousands of kilometers along strike obviates the need for east-dipping flat-slab subduction. The temporal overlap of the thick-skinned deformation, which formed mainly during the Maastrichtian and Paleocene

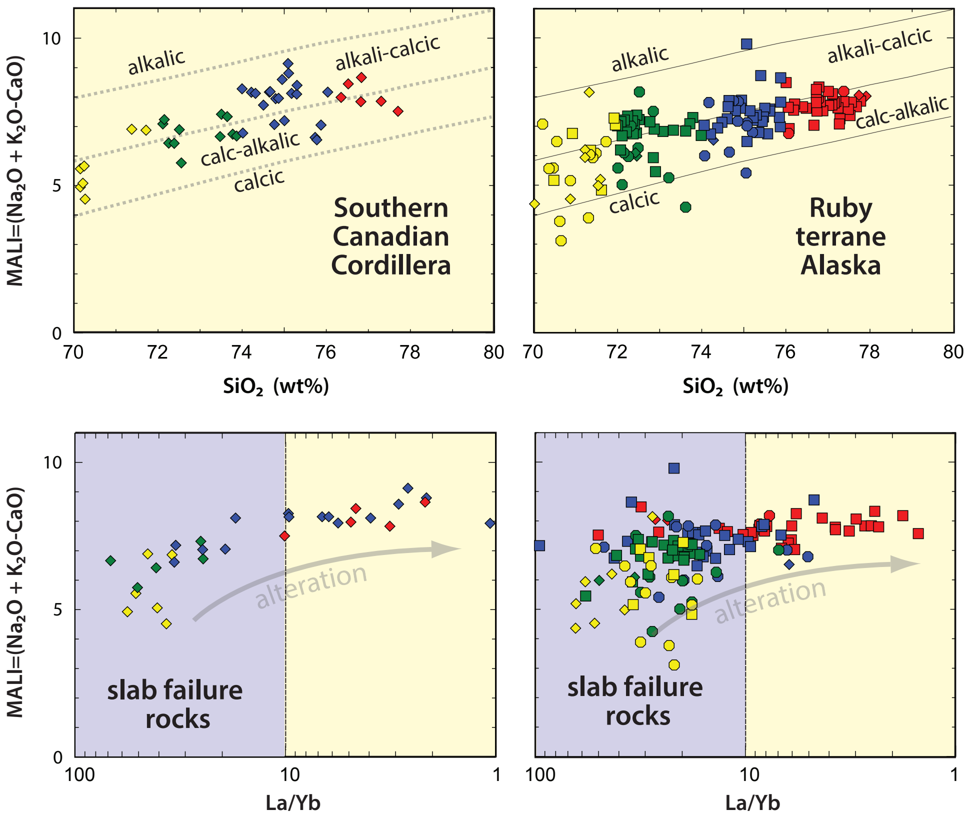


Figure 47. Plutonic rocks with SiO_2 >70 wt% from the Bayonne suite, southern Omineca belt (Brandon and Lambert, 1993, 1994) and Ruby terrane, Alaska (Arth et al., 1989a, 1989b), both likely slab failure–related plutonic suites, plotted on $Na_2O + K_2O$-CaO (or MALI) vs. SiO_2, and MALI vs. La/Yb discrimination diagrams. We subdivided samples using silica bin intervals of 2% and colored each interval with a different color. In both cases, rocks with higher silica contents show decreasing La/Yb—a feature more likely to have resulted from alteration by a fluid component rather than fractionation. MALI diagram is from Frost et al. (2001).

(Dickinson et al., 1988), and the major northerly, or dextral, migration of the adjacent Cordilleran block (Hildebrand, 2015) suggests that the thick-skinned deformation was generated by transpression during meridional migration. The observation that the Campanian hinterland belt was separated and transported northward during the period of meridional migration also demonstrates that the accompanying main phase of thick-skinned deformation is younger than the deformation, metamorphism, and plutonism of the hinterland belt.

The meridional migration model presented here and in Hildebrand (2015) encompasses a wide variety of observations—the dual metamorphic-plutonic hinterland belts in the Canadian Cordillera, the northwardly migrating Laramide foredeep, the lack of a Sevier-age metamorphic-plutonic welt in the Great Basin, the robust paleomagnetic record of northward migration, the termination of both the Lewis and Clark and Texas lineaments at the Cordilleran fold-and-thrust belt, the lack of contrast between the thick- and thin-skinned belts outside the central sector of the orogen, and the 14° clockwise rotation of the Colorado Plateau and its northward migration, as well as the changing directions of shortening within the Rocky Mountain foreland. The model brings all the observations into consilience and solves inconsistencies, such as the general lack of 80–60 Ma magmatism in the region of the Rocky Mountain foreland.

Laramide Orogenic Hinterland

Our proposed Laramide hinterland belt of dominantly Campanian age extends from Alaska to southern Mexico (Fig. 49), and possibly beyond (Hildebrand and Whalen, 2014a), as originally suggested by Blackwelder (1914), but here we are concerned only with the North American portion. Due to varying intensities of younger deformation, especially extension in the Basin and Range, Mojave, and Colorado River extensional corridor, as well as dismemberment by strike-slip faults, the belt as a whole is perhaps best delineated on maps by plutons ranging in age from 82 to 58 Ma (Figs. 31 and 50; for a similar approach, see also figure 1 *of* Armstrong, 1974), and by overlapping ages of deformation, metamorphism, and exhumation.

In addition to the magmatism, there is a broad swath of Cretaceous deformation, metamorphism, and exhumation that extends northward from Sonora westward across the Mojave Desert into the Transverse Ranges. 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—with similar cover lithologies and stratigraphy to those of the North American platform (Stone et al., 1983) that were metamorphosed to amphibolite grade—locally thinned to 1% of their original thickness and recumbently folded between

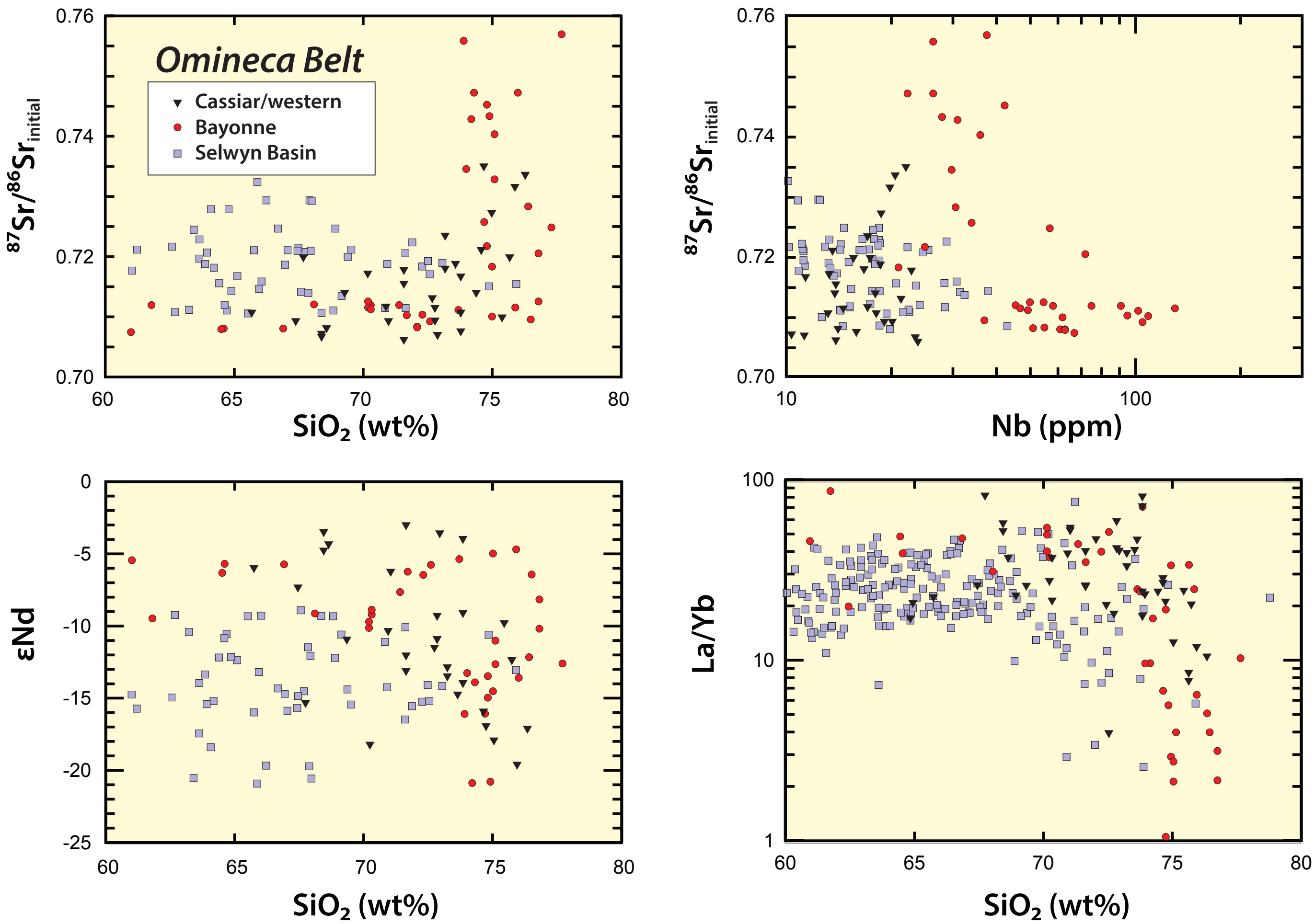


Figure 48. Samples from plutonic suites in the Omineca belt and Selwyn Basin plotted on $^{87/86}Sr_{initial}$ vs. SiO_2, $^{87/86}Sr_{initial}$ vs. Nb; ε_{Nd} vs. SiO_2, and La/Yb vs. SiO_2 plots. On the plots, samples from the Bayonne and western Cassiar suites with SiO_2 >73 wt% exhibit major dispersion in isotopic compositions and Nb contents, whereas lower-silica samples show less variation. In contrast, Selwyn Basin samples show significant variation in isotopes at a given silica value over the suite's full silica range, which could represent heterogeneities within the source region. All suites exhibit a significant decrease in La/Yb in samples with >70 wt% silica, consistent with alteration seen in the previous figures. Data are from Rasmussen (2013), Pigage et al. (2014), Driver et al. (2000), Morris and Creaser (2008), and Brandon and Lambert (1993, 1994).

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). Wells and Hoisch (2008) demonstrated that extensional collapse, exhumation, and magmatism within the Mojave sector occurred largely during the interval 75–67 Ma.

South of Tucson, Arizona, and extending westward to Ajo and Organ Pipe National Monument, there 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 74–64 Ma (D.M. Miller et al., 1992). The McCoy Mountains Formation, which outcrops in faulted ranges in southern California and Arizona, trends more or less E-W and contains 84 Ma zircons near the exposed top, which was overthrust by Jurassic rocks (Tosdal, 1990; Tosdal and Stone, 1994) and cut by a granitic pluton dated at 73.5 ± 1.3 Ma (Barth et al., 2004).

Farther to the west at the NW end of the San Bernardino Mountains, 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 Cretaceous migmatites and mylonitic thrust zones, including the Cucamonga and San Antonio slices in the San Gabriel Mountains, 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

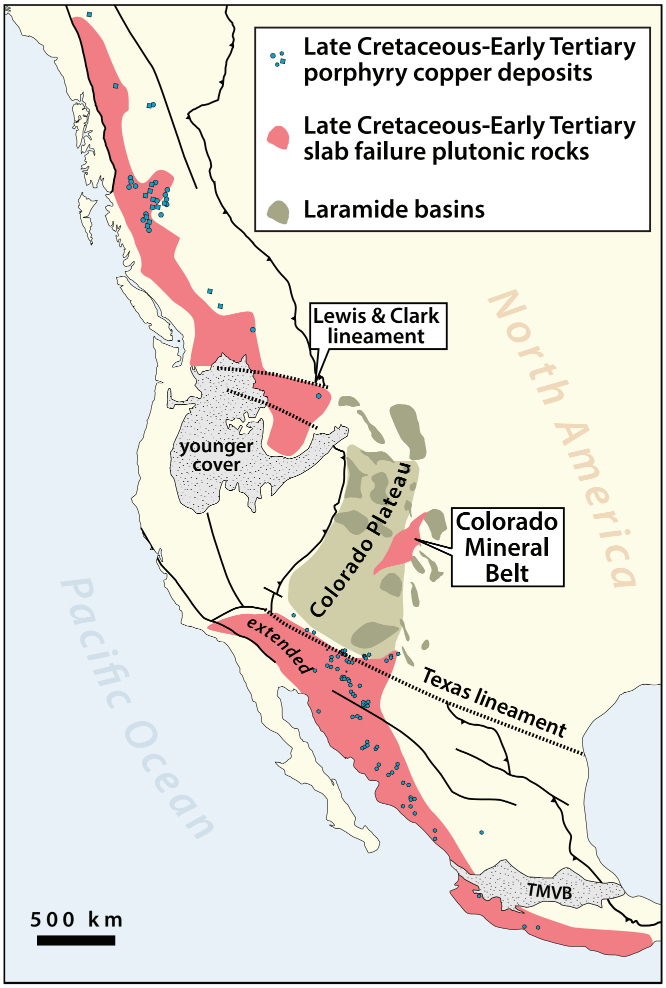


Figure 49. Geological sketch map showing the distribution of Late Cretaceous–early Cenozoic magmatism in western North America and associated porphyry copper deposits. Hildebrand (2015) noted the similar relations along the Lewis & Clark and Texas lineaments and used them as piercing points to resolve 1300 km of northward meridional migration in keeping with robust paleomagnetic data sets. By restoring the lineaments into a single through-going structure, the Laramide hinterland and its Late Cretaceous–early Cenozoic plutonic rocks are joined into a continuous belt. TMVB—Trans-Mexican volcanic belt.

regional K-Ar study of the eastern Transverse Ranges and southern Mojave Desert area showed that most rocks, including Precambrian gneisses, yielded biotite ages in the range 70–57 Ma (Miller and Morton, 1980), which suggest 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., 2003).

A broad, but crudely linear belt of 76–55 Ma plutons (Figs. 49 and 50) coincides with the belt of deformation, metamorphism, and exhumation. It extends N-S 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, 2017). Recent U-Pb zircon dating in northern Sonora revealed that the Tarahumara Formation (Fig. 50), a >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). The $^{40}Ar/^{39}Ar$ geochronometry indicates exhumation between ca. 68 and 50 Ma (Valencia-Moreno et al., 2006; González-León et al., 2017).

Large numbers of porphyry copper deposits occur in the Sonoran segment and, along with their epizonal plutonic sources, formed between ca. 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, but they are broadly constrained by geological relations to be Late Cretaceous–Paleogene. Within the southern Arizona sector, porphyry copper systems are better known, and most of the mineralized intrusions are 80–50 Ma in age (Lang and Titley, 1998). Late Cretaceous–Early Tertiary magmatism and porphyry mineralization also occurred 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).

Latest Cretaceous–Early Tertiary plutons coincide with the region of thrusting in the southwestern United States and are a characteristic of the Mojave and Colorado River desert region. A well-studied example within eastern California is the Old Woman–Piute Range batholith, which postdates peak metamorphism and deformation; consists of 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 ca. 70 Ma Chemehuevi Mountains plutonic suite, 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 and heavy REEs (Kapp et al., 2002). Within the Whipple Mountains metamorphic core complex, the 73 Ma Axtel quartz diorite has similar characteristics (Anderson and Cullers, 1990).

The ca. 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. They tend to be more quartz-rich than the older predeformational, 80 Ma plutons (John, 1981), and they have higher Sr/Y ratios (R. Economos, 2011, personal commun.).

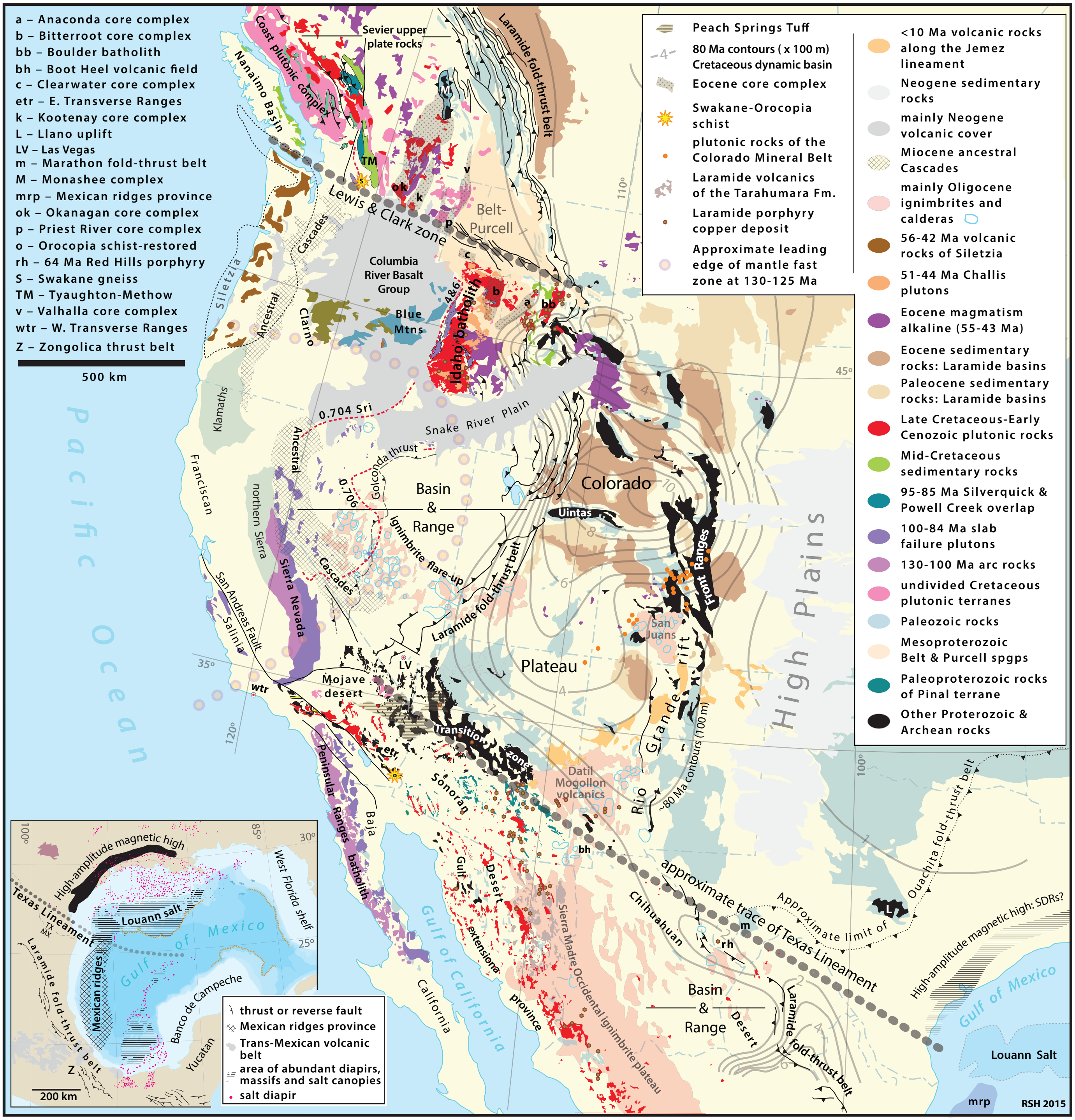


Figure 50. Geological sketch map of west-central North America showing the two transverse structural zones and some major geological units (from Hildebrand, 2015). Geology is from Reed et al. (2004), with 80 Ma contours of Cretaceous rocks from Roberts and Kirschbaum (1995). Great Basin calderas are from Henry and John (2013); Late Cretaceous–Paleocene $^{87/86}Sr_i$ isopleths are from Armstrong et al. (1977); spgps—supergroups. Inset: Gulf of Mexico region, as simplified from Reed et al. (2004), showing the sinistral offset of early rift-phase Callovian salt deposits (reflected in salt domes) and the southward truncation of the high-amplitude magnetic high, interpreted to represent basalts of a volcanic rifted margin from Mickus et al. (2009). Hildebrand (2015) used the salt deposits to suggest that the SW margin of North America formed during the Late Jurassic and had a west-northwest orientation. SDRs—seaward-dipping reflectors; 4&6—coincident 0.706 and 0.704 Sr_i isopleths. See insert for larger version of this figure.

Exposed within the eastern Transverse Ranges (Fig. 50), there is an oblique cross section representing as many as 22 km of paleodepth (Barth et al., 2008). 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 they are only weakly deformed (Needy et al., 2009).

We plotted geochemical data from several plutons, volcanic belts, and porphyry intrusions from Sonora and the U.S. Southwest (Figs. 51, 52, and 53) on our standard plots, and they form arrays more typical of slab failure rocks than arc assemblages.

Today, the hinterland belt ends rather abruptly in the Transverse Ranges (Powell, 1993), but rocks there and within the Mojave Desert must be restored southeastward and eastward to account for younger displacement on the San Andreas fault system and extension within the Mojave–Colorado River corridor–Basin and Range extensional provinces. Although the precise amount of extension across the region is unknown, it is substantial, likely exceeding 100% (see Frost and Martin, 1982), which would restore the western edge of the belt approximately to the California-Arizona state line, just east of the southward projected Cordilleran fold-and-thrust belt.

Thirteen hundred kilometers to the north, the Late Cretaceous–early Cenozoic metamorphic-plutonic hinterland occurs west of the fold-and-thrust belt within the Helena salient (Fig. 28). There, large batholithic masses like the Boulder and Idaho batholiths were emplaced during and after thrusting, as were much smaller complexes such as the Pioneer, Tobacco Root, Sapphire, Mount Powell, and Phillipsburg batholiths (Fig. 29).

This is an area critical to understanding the tectonic development of the Cordillera, and one often cited as proof that there was no Laramide collision. There, rocks were involved in both the Oregonian and Laramide events, and so relations are complex, confusing, and locally problematic. Additionally, there is general consensus among researchers that rocks of the Belt Supergroup were deposited upon North America and therefore are not exotic, whereas others (Johnston, 2008; Hildebrand, 2009, 2013) have argued that rocks of the Belt Supergroup formed part of the upper plate beneath which the leading edge of North America was pulled during the Laramide event. Thus, a more extensive discussion of the area should be beneficial to understanding the nature of the Laramide event, and so we have included a more detailed examination of the controversy in Appendix 1.

The Butte granite (Fig. 54), the largest pluton within the Boulder batholith (Hamilton and Myers, 1967), and host to rich porphyry copper deposits, was emplaced at about 77 Ma and exhumed from 6–9 km to ~1 km between 66 and 58 Ma (Houston and Dilles, 2013a). Just to the west, within the Bitterroot lobe of the Idaho batholith, plutonism preceded and overlapped with Paleogene tectonic collapse and exhumation (Foster and Fanning, 1997). Other bodies in the region, mostly laccoliths, occur in the Big Belt and Slide Rock mountains just to the east (Fig. 28).

There have been numerous studies of the plutonic rocks, but only a few have suitable modern analyses. We plotted the available geochemical data from the area on our standard plots (Figs. 51, 52, and 53), and they define arrays that sit almost exclusively within the slab failure fields. Their syn- to postdeformational emplacement and their geochemical characteristics suggest that the 82–53 Ma plutons of the Helena salient are slab failure rocks and not arc rocks.

To the west, the Cascades core contains several intrusions in the age range 80–58 Ma (Fig. 35), including many bodies of orthogneiss, the dominant rocks within the Skagit gneiss (Miller et al., 2009), but these are not delineated on the figure. Unfortunately, we were unable to access any modern analyses from those rocks. In some locations, Gordon (2009) found leucosomes in migmatite to be 71–47 Ma.

The Coast plutonic belt is rich in plutons in the age range 80–58 Ma, and on the basis of K/Ar and Ar ages, it was metamorphosed and exhumed during the Paleogene (Armstrong, 1988; Erdmer and Mortensen, 1993). Substantial plutonic complexes include the late synkinematic Ruby Range batholith, largely biotite-hornblende tonalite to granodiorite (Israel et al., 2011; Johnston and Canil, 2007), and the 1200-km-long Great Tonalite sill, a linear and steeply dipping composite body of mainly hornblende-biotite tonalite with U-Pb ages ranging from 68 to 55 Ma and with K-Ar biotite ages close to 50 Ma (Thomas and Sinha, 1999; Gehrels et al., 1991, 2009; Mahoney et al., 2009; Barker et al., 1986; Erdmer and Mortensen, 1993). Plutonic rocks of the Coast plutonic belt also plot mostly within the slab failure fields on our standard discrimination plots (Figs. 51 and 52), are syn- to postkinematic, and were mostly emplaced during exhumation, so we believe that they are products of slab failure magmatism.

Other magmatic rocks of the Canadian Cordillera that fall in the 80–53 Ma range are the dominantly intermediate 70 Ma Carmacks volcanics (Johnston et al., 1996), rocks of the ca. 70 Ma Middle Forks caldera (Bacon et al., 2014a), and mineralized ca. 76 Ma volcanic rocks of the Aishihik lake area, Yukon (Morris et al., 2014), all of which are located on the Yukon-Tanana terrane (Fig. 31).

The distribution of Late Cretaceous–early Cenozoic plutons and similar ages of exhumation indicate that the hinterland passes from south to north through Mexico and turns nearly east-west through Sonoran Arizona and the Mojave Desert to end abruptly in the Transverse Ranges of southern California (Figs. 44 and 50). It reappears in the Helena salient of Idaho-Montana and continues northward—somewhat dislocated by the Lewis and Clark line—through the Cascades core and the Coast plutonic complex of British Columbia–Yukon Territory to the Denali fault, where it is offset dextrally several hundred kilometers before it arcs across Alaska to the Alaskan Peninsula (Fig. 31). As discussed previously, Hildebrand (2015) used large-scale piercing points to palinspastically restore the two outcrop belts (Fig. 49) into one continuous strand.

The best-preserved examples of the thin-skinned fold-and-thrust belt directly related and adjacent to the orogenic hinterland

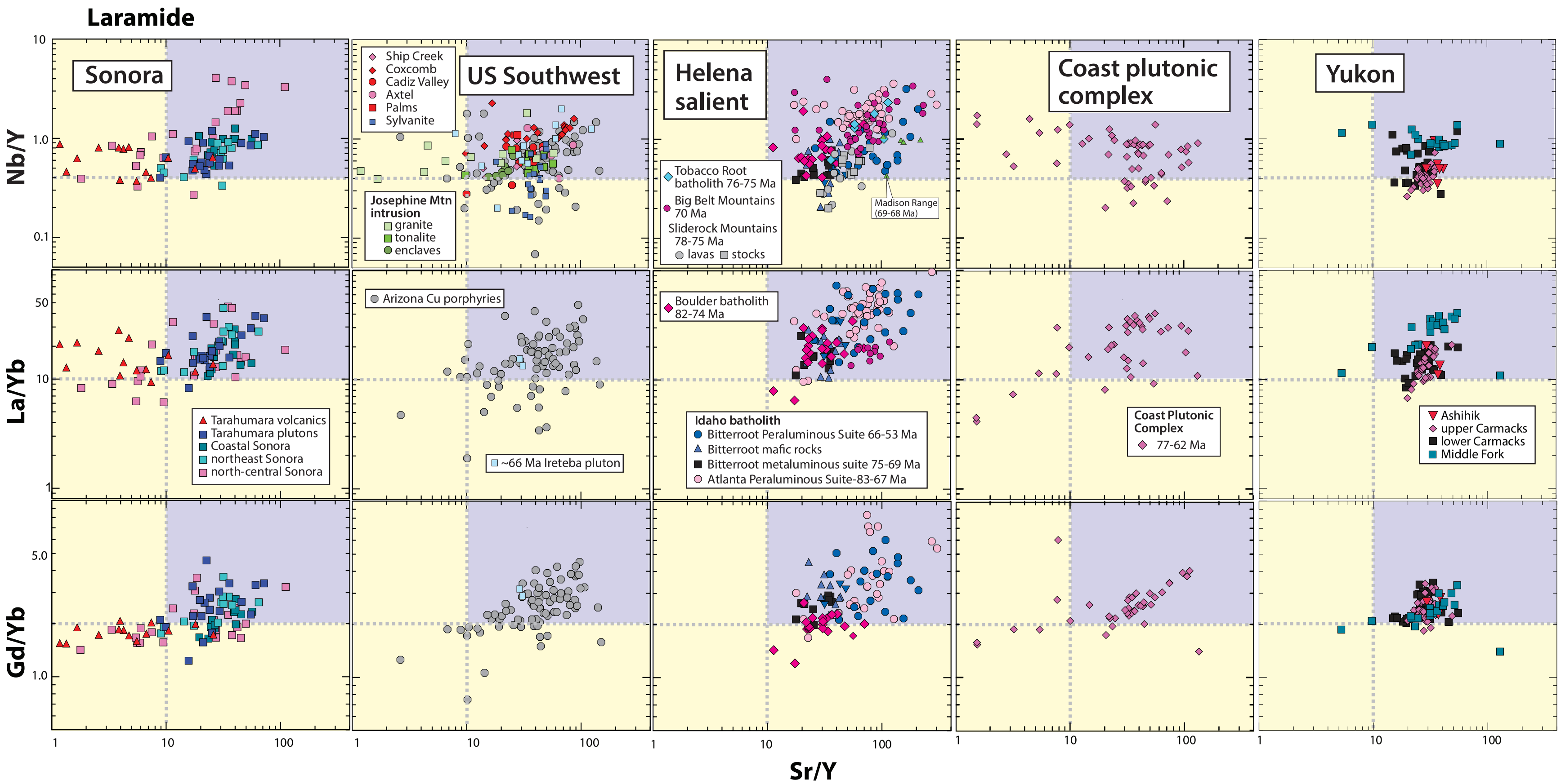


Figure 51. Nb/Y, La/Yb, and Gd/Yb vs. Sr/Y discrimination diagrams for syn- to posttectonic Laramide magmatic rocks of the Cordillera, illustrating that all suites plot in our slab failure fields on these diagrams. Note that we did not remove samples with >70% SiO_2 because we wanted to illustrate the effect of fractionation and/or alteration on the Sr/Y ratios. Those rocks trend horizontally to the left on the plots due to Sr loss in the most siliceous samples. Data sources: Tarahumara volcanics and plutonics from González-León et al. (2011, 2017); Coastal Sonora from Valencia-Moreno et al. (2003); northeast Sonora from González-León et al. (2000, 2017); north-central Sonora from Valencia-Moreno et al. (1999); Coast plutonic complex from Girardi et al. (2012), Mahoney et al. (2009), and Crawford et al. (2005); Ship Creek, Coxcomb, Cadiz Valley, and Axtel from Anderson (2015, personal commun.); Palms pluton from Roell (2009); Sylvanite pluton from Channell et al. (2000); Josephine Mountain (Mtn) pluton from Barth et al. (1995); Arizona Cu porphyries from Lang and Titley (1998); Ireteba pluton from D'Andrea Kapp et al. (2002); Aishihik from Morris et al. (2014); Carmacks from Francis (2015, personal commun.); Middle Fork from Bacon et al. (2014a); Tobacco Root batholith from Sarkar et al. (2009); Big Belt Mountains from du Bray and Snee (2002); Sliderock Mountains from du Bray and Harlan (1998); Boulder batholith from du Bray et al. (2012); Idaho batholith from Gaschnig (2015, personal commun.); and Madison Range from Tysdal et al. (1986).

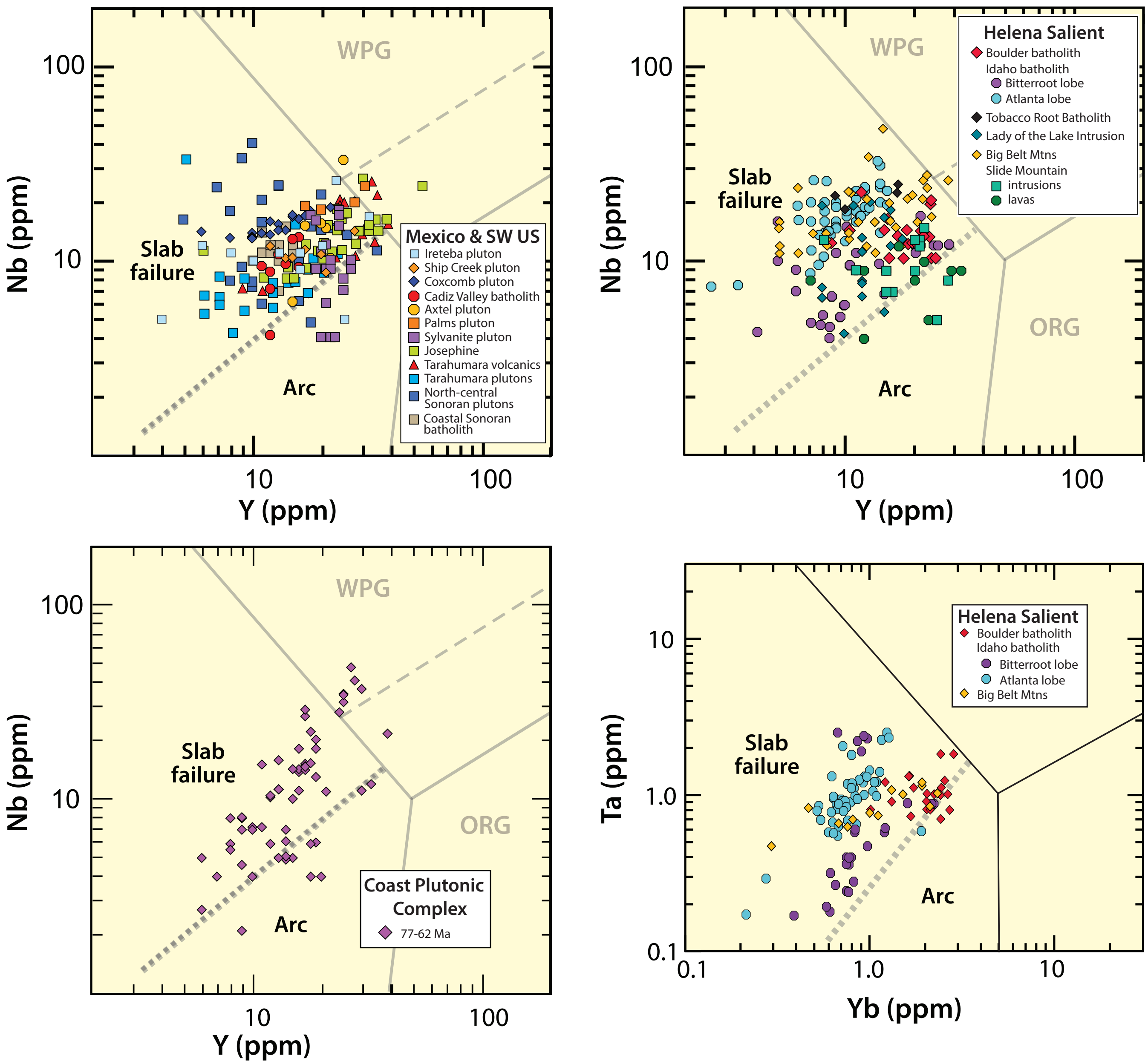


Figure 52. Nb-Y and Ta-Yb discrimination plots for various posttectonic Laramide magmatic suites, illustrating that Laramide magmatism was dominated by slab failure, not arc magmatism. Sources are as in Figure 51. WPG—within-plate granite; ORG—ocean-ridge granite.

occur in the Zongolica fold-and-thrust belt of eastern Mexico (Nieto-Samaniego et al., 2006) and in the Skeena fold-and-thrust belt of British Columbia (Evenchick, 1991). Within the Skeena fold-and-thrust belt (Fig. 31), rocks of western Stikinia, covered with a thick veneer of Middle–Late Jurassic to mid-Cretaceous clastic sedimentary rocks, were thrust northeastward during the Campanian–Maastrichtian, as documented by easterly migrating clastic sedimentation and thrusts of the Sustut foreland basin (Wheeler and Gabrielse, 1972; Evenchick, 1991; Ricketts, 2008). The thrusts root to the west within the Coast plutonic complex, where eastward-vergent ductile thrusts are bracketed to have been active between 87 and 68 Ma (Rusmore and Woodsworth, 1991), with major exhumation of the metamorphic hinterland during the early to middle Paleogene (Wheeler and Gabrielse, 1972; Erdmer and Mortensen, 1993; Armstrong, 1988).

In easternmost Mexico, the Zongolica belt (Figs. 2 and 50, inset) strikes southward to Guatemala, where a west-facing Cretaceous platform that sat on Mesoproterozoic–Triassic basement

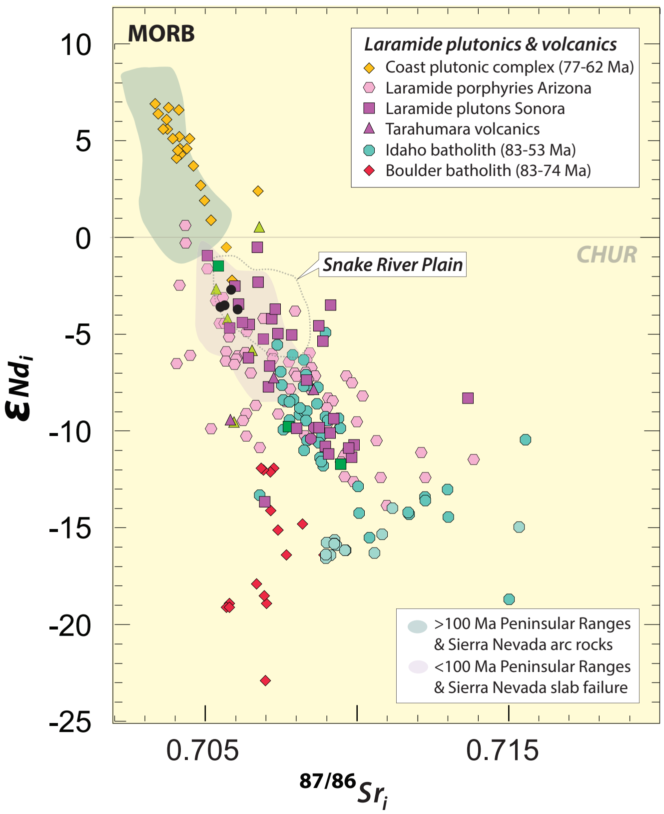


Figure 53. ε_{Nd} vs. $Sr_{initial}$ plot for Laramide magmatic suites. Fields for arc and slab failure suites from the Peninsular Ranges and Sierran batholiths, as well as basalts from the Snake River Plain, are shown for comparison purposes (see Fig. 23). Note that rocks of the Coast plutonic complex are isotopically similar to arc rocks yet have typical slab failure signatures in their trace-element profiles, as shown on previous figures. Data sources are as in Figure 51. MORB—mid-ocean-ridge basalt; CHUR—chondritic uniform reservoir.

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 then exhumed to amphibolite grade a million years later (Martens et al., 2012), presumably after slab failure.

Other examples of the fold-and-thrust belt are well known in Montana (Fig. 28), where rocks of the Belt Supergroup form the structurally highest allochthons in a northeastward-vergent structural stack (Fuentes et al., 2012), in Idaho (Skipp and Hait, 1977, Skipp, 1987), and to the south in Wyoming and Utah (Yonkee and Weil, 2011; DeCelles, 2004), where thrusts are of both Sevier and Laramide age (Hildebrand, 2013). As described earlier, Hildebrand (2009, 2013, 2014) argued that the Laramide event is best interpreted to represent terminal collision of a ribbon continent with North America and its neoautochthon, now located within the Canadian Cordillera due to post–70 Ma meridional migration.

During and after the compressive deformation, the extensive firestorm of plutons was emplaced into the metamorphic hinterland, where they currently outcrop from southern Mexico to the Transverse Ranges of southern California and from the Helena salient northward to Alaska (Figs. 31 and 50). Most of the plutons are syn- to postdeformational and were emplaced into the crust during exhumation and cooling (Miller and Morton, 1980; Armstrong, 1988; Wells and Hoisch, 2008; Miller et al., 2009). Our trace-element geochemistry (Figs. 51, 52, and 53) supports the idea that these rocks resulted from slab failure and do not represent arc rocks.

Understanding that the metamorphic-plutonic belt was the upper plate during the Laramide collision helps to clarify many previously difficult-to-understand relations throughout the Cordillera. For example, fragments of older terranes and structures related to older events, located within the upper plate, outcrop sporadically along and/or across strike, but they are truncated to the east by the Laramide suture. In this way, older thrust belts are separated from their original plate boundary, or slivers of much older accretionary prism rocks are left along the younger plate boundary, similar to relations found along the Caribbean coastal region of Venezuela (Avé Lallement and Sisson, 2005). Thus, as workers try to understand progressively older collisional belts within the Cordillera, they should appreciate not only meridional dispersal along the margin, but also the likely possibility that complete cross sections through individual collisional belts may not be available in many locations, and that compressional structures of different ages could be juxtaposed across a single fault (e.g., Pană and van der Pluijm, 2015).

Overall, we tried to show that the timing of emplacement for the Cretaceous batholiths that we examined was dominantly postcollisional and that they are therefore incompatible with arc magmatism. Plotting trace-element geochemistry, where modern analyses are available, allows us to discriminate between arc and slab failure magmatism. By combining the geochemical discrimination data with the syn- to postcollisional timing, we conclude that most Cretaceous plutons of the North American Cordillera were generated by slab failure magmatism.

DISCUSSION

Involvement of Subcontinental Lithospheric Mantle?

While the general differences in plutonism within Cordilleran batholiths—at least in terms of transverse compositional asymmetry—have long been known (Lindgren, 1915; Buddington, 1927; Larsen, 1948; Moore, 1959: Moore et al., 1961), the plutons of the great Cordilleran batholiths were generally considered to have been generated by the same general process, however mysterious it might have been at the time. Following the

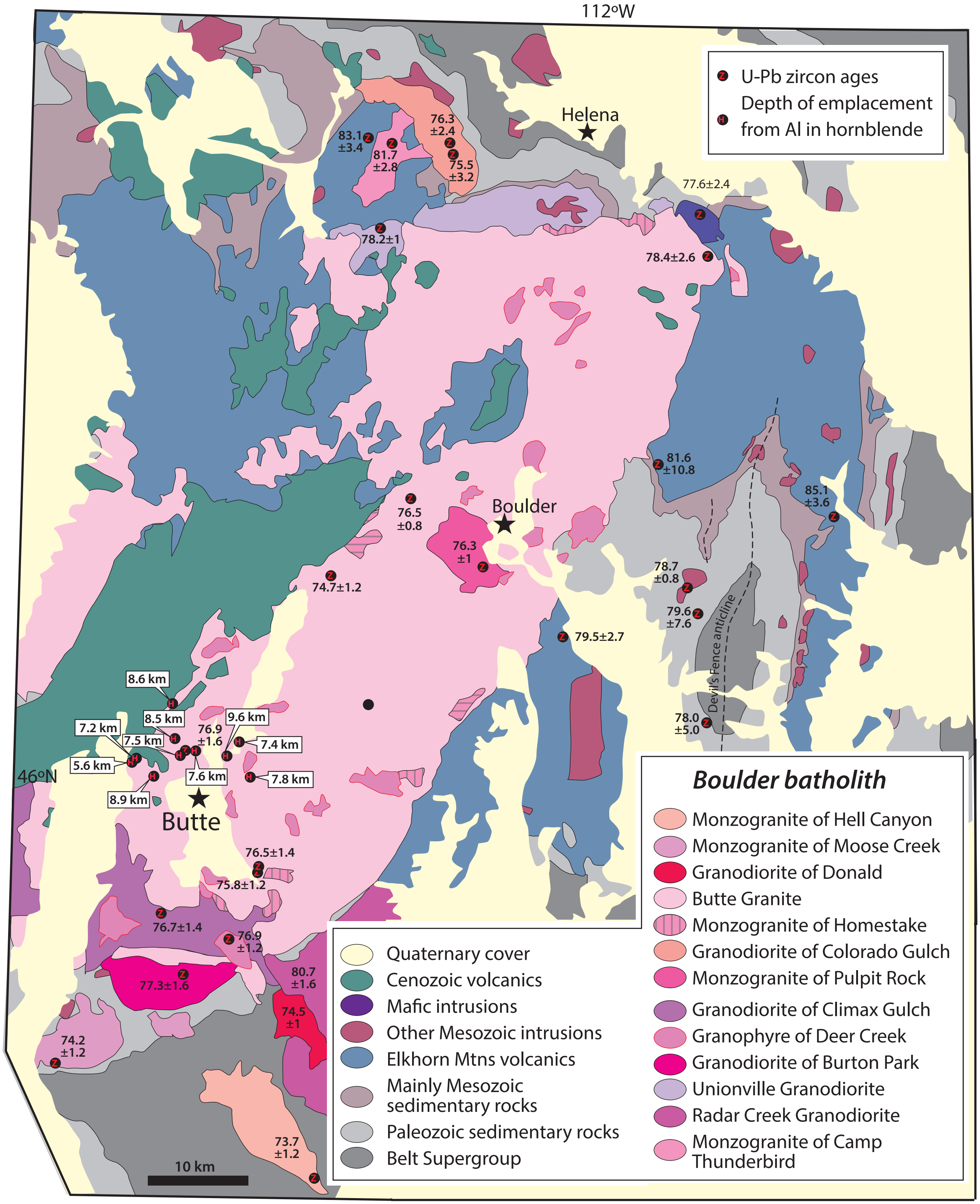


Figure 54. Geologic sketch map of the Boulder batholith and its wall rocks modified from du Bray et al. (2012), showing the various component plutons and their U-Pb zircon ages. Depths of emplacement calculated from samples collected from the southern end of the Butte granite using Al in hornblende are from Houston and Dilles (2013a).

acceptance of plate tectonics, the idea that the plutons were produced by continuous arc magmatism—which nearly all workers hypothesized to have been generated by eastward subduction—is still in vogue today (Ducea and Barton, 2007; Mahoney et al., 2009; Paterson et al., 2014; Cao et al., 2015; Ducea et al., 2015a). The general idea is that older rocks were emplaced into accreted terranes above an eastwardly dipping subduction zone, but that the eastern suite was emplaced into the western margin of autochthonous North America; otherwise, the process of formation was the same: an eastwardly dipping subduction zone that shallowed with time. This led most workers to conclude that overall compositional variations were largely the result of the crustal component becoming more important as the source of the magmatism moved farther east beneath the continent.

Kistler and Peterman (1973) used the geochemical and isotopic differences to suggest that magmas of the western group were derived from the upper mantle, whereas the more eastern facies were derived from intermediate to mafic lower crust of Precambrian age. Kistler (1978, 1990), Krummenacher et al. (1975), and Armstrong et al. (1977) argued that the Sr_i 0.706 isopleth corresponded to the boundary, typically a regional-scale cryptic fault, between Precambrian crust on the east and Phanerozoic eugeosynclinal lithosphere to the west, or in more modern plate-tectonic terms, North America and exotic accreted terranes. Nowadays, the concept that the Sr_i 0.706 isopleth represents the western edge of autochthonous Precambrian basement of North America is perhaps one of the most widely accepted hypotheses in Cordilleran tectonics and paleogeography (Miller et al., 2000). The only real challenge to this paradigm was presented by Drew Coleman (Coleman et al., 1992; Coleman and Glazner, 1998), who utilized Nd and Sr isotopic analyses of plutons, mantle xenoliths, and Cenozoic basalts to suggest that the Sr_i 0.706 isopleth reflects the edge of Proterozoic-age enriched lithospheric mantle, not crust. Earlier, Chen and Tilton (1991) recognized the possibility that plutons of the eastern Sierra Nevada (Sierran Crest) were derived from enriched subcontinental mantle, using the similarities between Sr and Pb isotopic analyses of eastern Sierran plutons and those of ca. 3.5 Ma ultrapotassic volcanic rocks of the central Sierra Nevada (Van Kooten, 1980, 1981), but on the basis of ^{18}O data, they decided to favor assimilation of granulitic lower crust. However, in a more recent study of $\delta^{18}O$ from zircon, quartz, and whole rock within Sierran plutons, Lackey et al. (2008) found many eastern plutons with $\delta^{18}O_{zircon}$ to fall within the range of mantle $\delta^{18}O_{zircon}$ values. For example, Tuolumne plutons have $\delta^{18}O_{zircon}$ ratios of 6.0%–6.6%, Mount Whitney zircons are 5.67%–5.90%, and other plutons emplaced at 96 Ma range as low as 4.21%. They noted that those values supported studies by Coleman and others, cited earlier, that plutons of the Sierran Crest "were generated directly from melting of enriched (Sr_i = 0.706; ε_{Nd} = –4.5) lithospheric mantle beneath the eastern Sierra" (p. 1415). Note that in Sr versus O isotope space, the Sierran Crest plutons are similar isotopically to Sierran pyroxenite xenoliths, which are interpreted by most workers to represent subcontinental mantle lithosphere (Fig. 55).

Crustal Input: Fact or Fiction?

In our earlier contribution on the Peninsular Ranges batholith (Hildebrand and Whalen, 2014b), we noted that the crust beneath the eastern and western sectors of the batholith was more or less the same, in that prior to rifting and opening of the Bisbee-Arperos seaway, the Mesozoic units were continuous across the region (see also Lawton and McMillan, 1999). For example, rocks of the latest Jurassic Peñasquitos Formation of the westernmost Peninsular Ranges (Kimbrough et al., 2014a) and the Cucurpe Formation of Sonora (Mauel et al., 2011) are the same age, have similar basements, were both deformed between ca. 145 and 139 Ma, and were unconformably overlain by rocks in the range 130–125 Ma. As we located the collisional suture between the Santiago Peak–Alisitos arc on the west and the west-facing passive margin on the mainland to the east, we were able to show that the basement to the arc was quite heterogeneous and contained rocks ranging in age from Paleoproterozoic to Early Cretaceous. It is important to realize that the two suites, arc and slab failure, overlap spatially on the ground, indicating that they both came through the same crust (see Hildebrand and Whalen, 2014b, their figures 4 and 5).

Plutons of the Sierran batholith intruded a similar spectrum of rocks (Bateman, 1992) ranging from the latest Jurassic Mariposa Formation, another equivalent of the Peñasquitos Formation (Kimbrough et al., 2014a), to Lower Paleozoic rocks of the Shoo Fly Complex (Harwood, 1992), sandstones of which contain abundant Precambrian detrital zircons (Girty and Wardlaw,

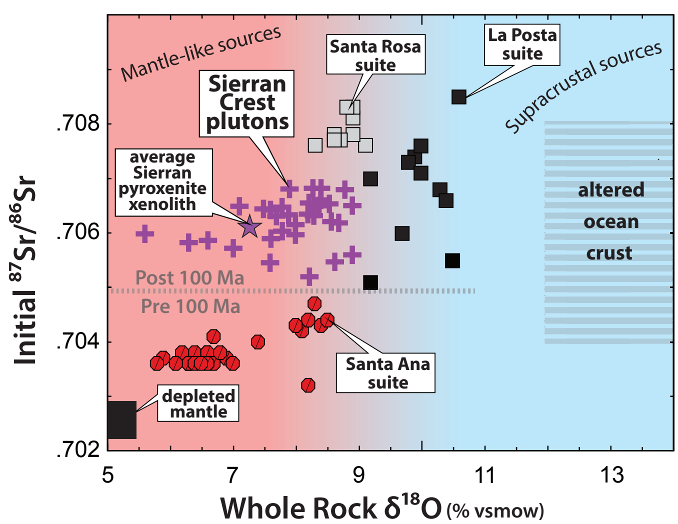


Figure 55. Sr vs. O isotopes for plutons of the Sierran Crest magmatic suite from Lackey et al. (2008) compared with pre– and post–100 Ma plutonic samples from the Peninsular Ranges batholith (Kistler et al., 2014). Note that samples of Sierran Crest plutons plot near the average Sierran pyroxenite xenolith, possibly representing enriched mantle, whereas La Posta plutons could be mixtures derived from enriched mantle and hydrothermally altered ocean crust. vsmow—Vienna standard mean ocean water.

1984, 1985; Harding et al., 2000), suggesting proximity to Precambrian basement.

Chen and Tilton (1991) used Pb isotopes of rocks in the Sierran batholith with Sr isotopic ratios >0.706 to suggest a mean age of 1.8 Ga for the crust that possibly interacted with magmas to form the plutons, although they could not rule out an old subcontinental lithospheric mantle source. Our geochemical plots suggest that the most isotopically evolved magmas within Cordilleran batholiths are slab failure–related and not arc-related magmas (Figs. 23, 45, and 53). These findings require a reevaluation of the various source regions involved. We noted earlier that the crust traversed by magmas of both suites is approximately the same, and, when these similarities are combined with the observation that there is spatial overlap between the eastern and western suites, it seemed to us that the crust cannot have been the principal cause of the compositional differences. This led us to wonder: How much crust is involved in the generation of the plutonic rocks, both in arcs and slab failure regimes? In other words, if the rocks generally inferred to represent arc rocks with the greatest amount of assimilation of continental crust are not in fact arc rocks, but slab failure related, then how much crustal assimilation really occurs during ascent of arc magmas through the crust?

To answer this question, we turned to our compiled data sets. A pronounced feature on our plots of arc magmatism is the general similarity in both trace-element abundances and ratios between the oceanic Aleutian arc and the continental Santiago Peak–Alisitos arc, as well as much younger arcs built on continental crust (Fig. 56). These similarities suggest to us that crustal components might not be as important in the genesis of continental arc rocks as generally hypothesized (Bateman et al., 1963; DePaolo, 1981; Bateman, 1992; Ducea and Barton, 2007; Gaschnig et al., 2011). We therefore tested this idea with more extensive data sets of both oceanic and continental arcs using elements commonly associated with continental crust (Fig. 57). Except for some obvious magmatic fractionation trends, the samples of most suites range broadly but overlap with one another to form a large blob on the various plots, indicating that for most elements shown, there is little difference between oceanic and continental arcs. Some of our slab failure suites, such as the Sevier slab failure bodies, do show evidence for crustal interactions—at least with high-level fluids—but the evidence for crustal interaction in the Sierran or Peninsular Ranges batholiths, as well as most arc rocks, is meager.

Depth of Melting and Interactions with Lithospheric Mantle

Arc magmas are fundamentally different from slab failure magmas in that arc magmas are generated by volatile addition and partial melting of mantle rocks containing plagioclase in the shallow asthenosphere (Ulmer, 2001; Till et al., 2012), whereas slab failure rocks appear to be derived from deeper garnet-bearing assemblages (Gromet and Silver, 1987; Hildebrand and Whalen, 2014b) by pressure release melting of garnet peridotite, direct melting of garnet pyroxenite (Spandler et al., 2008), or combinations of the two (Lambart et al., 2016). Sm/Yb ratios are one of several elemental proxies for depth of partial melting (Putirka, 1999; Putirka et al., 2012). If one uses enriched mantle xenolith compositions from the Sierra Nevada as a source, then the Sm/Yb ratios for partial melts from a garnet peridotite source are >4.6 for all melt fractions (F) >0.1, whereas when F = 0.04, Sm/Yb is 2.4 (Putirka, 2016, personal commun.). As partial melts of spinel peridotite should produce more melt due to larger degrees of partial melting than the deeper garnet peridotites, most partial melts of spinel peridotite will have Sm/Yb less than ~2.5. We plotted a variety of arc rocks and slab failure rocks onto La/Sm versus Sm/Yb space (Figs. 58 and 59), and nearly all arc rocks, no matter whether continental or oceanic, have Sm/Yb <2.5 and are therefore derived from spinel peridotite, whereas nearly all slab failure rocks have higher values, many over the pure garnet peridotite value of 4.6. Others are mixtures of the two. These diagrams suggest that the value of 2.5 divides most arc from slab failure rocks and can be used to discriminate between the two, although, as with other diagrams, we found that rocks with >70% SiO_2 can be problematic, probably due to fluid-related alteration or fractionation of REE-rich accessory minerals, and therefore they should not be used.

Numerous geochemical and isotopic studies of basaltic volcanism in the western United States demonstrated that the rocks reflect assimilation of enriched continental mantle lithosphere and had little, if any, crustal input (Leeman, 1970, 1974; Van Kooten, 1980, 1981; Gibson et al., 1993; Farmer et al., 2002, 2013; Blondes et al., 2008). Many of our slab failure rocks have REE and isotopic values close to the basalts (Figs. 23 and 58), as well as other rocks argued to have involved subcontinental mantle lithosphere (Cousens et al., 2011; Putirka et al., 2012; Farmer et al., 2013), so we explored a model that involved sublithospheric melts rising to the base of, or into, the subcontinental lithospheric mantle, where they either melted enriched subcontinental lithospheric mantle or metabasalt derived from older arc magmatism or even very slightly older slab failure basalt that had reacted with the subcontinental lithospheric mantle and crystallized.

The most mafic melts that are known from the Sierran–Peninsular Ranges slab failure suite are those that cooled to form hornblende gabbros, typified by the well-studied Onion Valley sill complex in the Sierra Nevada (Sisson et al., 1996). They pointed out, using several factors, such as low Mg#, that the magmas of the intrusion could not be direct partial melts of mantle peridotite, but that they could have been derived from deep crystallization of hydrous primary magmas to leave ultramafic cumulates and an evolved basaltic melt, which then rose to final emplacement depths. The Nd and Sr data from the intrusion (Fig. 23) and ultramafic xenoliths in younger magmas also suggested to them that the mafic magmas were derived from enriched sub-Sierran lithospheric mantle. A major question is: What was the origin of the basalt that underplated or infiltrated the subcontinental lithospheric mantle? Was it from older arc magmatism, or was it from older slab failure melts, or both?

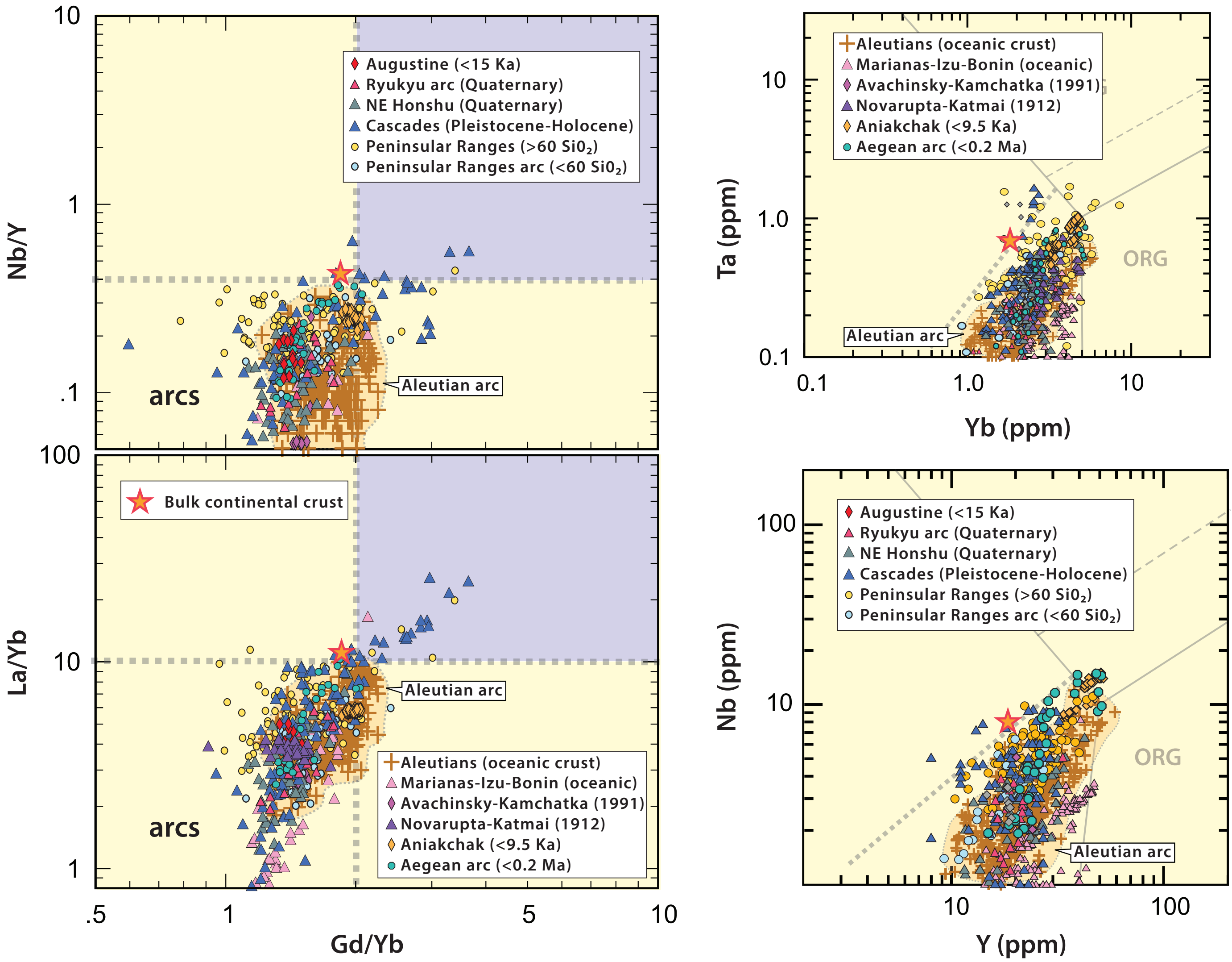


Figure 56. Nb/Y and La/Yb vs. Gd/Yb and Nb vs. Y and Ta vs. Yb discrimination diagrams for active volcanic arc suites and the pre–100 Ma Peninsular Ranges arc suite, illustrating the overall similarity of continental and oceanic arcs. The similarity suggests that assimilation of continental crust is not a dominant process in arcs built on continental crust. Data: Aleutian arc from Kelemen and Behn (2016); Novarupta-Katmai from Hildreth and Fierstein (2012); Aniakchak from Bacon et al. (2014b); Augustine from Johnson et al. (1996); Avachinsky from Viccaro et al. (2012); Aegean arc from Bailey et al. (2009); Ryukyu from Shinjo et al. (2000); and Peninsular Ranges arc from Lee et al. (2007). Aleutians: $n = 411$; other arcs: $n = 750$. ORG—ocean-ridge granite.

Detailed studies on basaltic rocks of the Snake River Plain further inform how slab failure melts might have evolved as they rose into and through the lithosphere. Lavas of the Snake River Plain are widely interpreted to represent a mantle “hotspot” that was overridden by North America as it migrated westward (Smith et al., 2009; Yuan and Dueker, 2005; Waite et al., 2006). Because the entire magmatic system—which includes Miocene and younger rocks of the Columbia Plateau, Oregon High Lava Plains, the Snake River Plain, and the Yellowstone Plateau—represents one magmatic system, and because the plume thus passed beneath exotic terranes and old Precambrian lithosphere as North America moved westward, Pb, Sr, and Nd isotopes collected from the westernmost basalts have plume isotopic signatures, whereas those collected farther east above ancient lithosphere suggest a source in the subcontinental lithospheric mantle (Hanan et al., 2008). Thus, the effects of subcontinental lithospheric mantle and crust on the magmas can be ascertained with high degrees of confidence.

Three-component mixing models, utilizing the oceanic-island basalt–like Steens-Imnaha lava to represent the plume component, old lithosphere like that of the Wyoming craton, and younger Paleoproterozoic-like lithosphere, show that >97% of the variability can be accounted for by progressive incorporation of older lithosphere into the plume source as it migrated

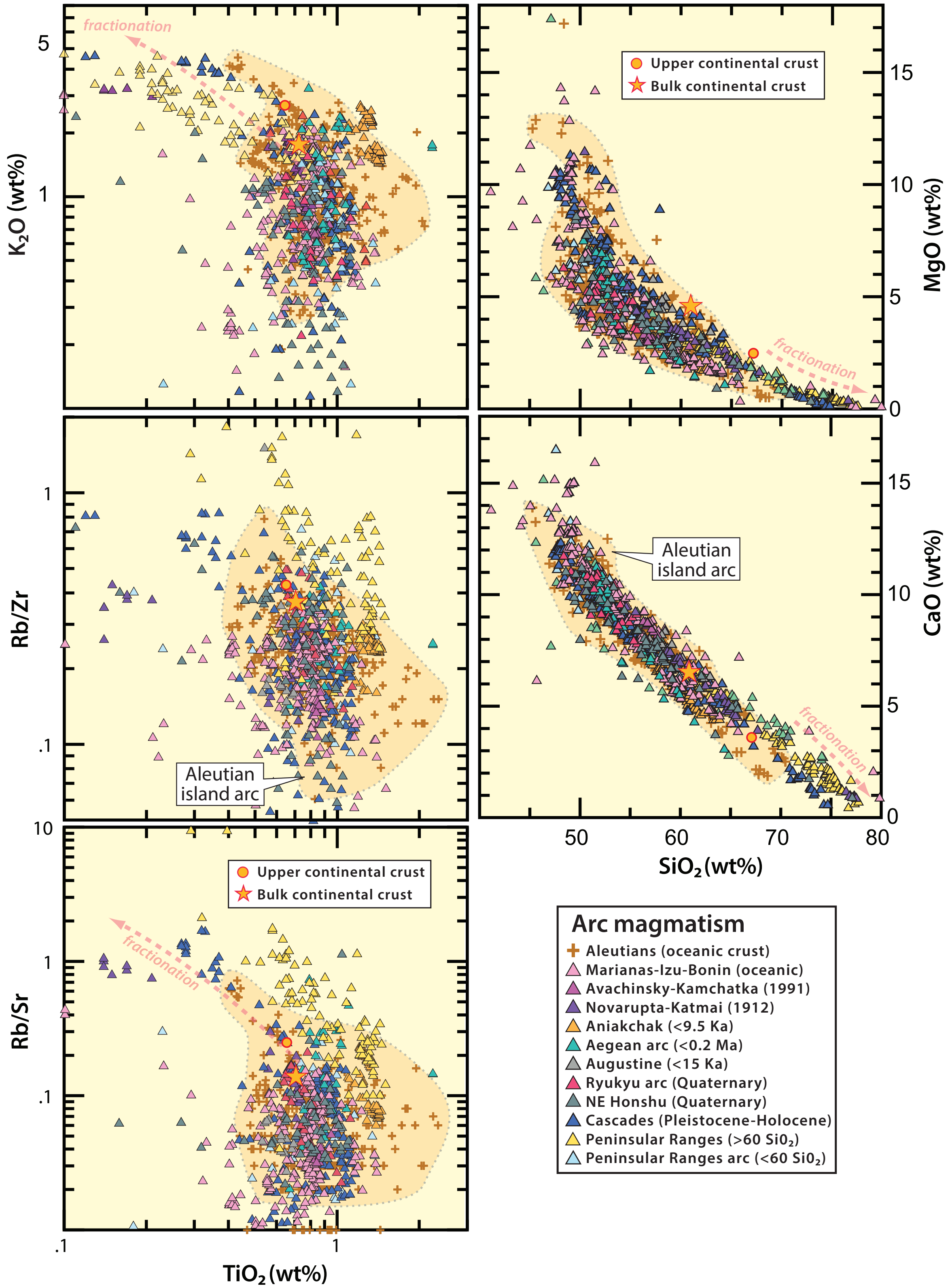


Figure 57. TiO_2 vs. Rb/Sr, Rb/Zr, and K_2O and SiO_2 vs. MgO and CaO for our oceanic- and continental arc–type magmatic suites plus the pre–100 Ma Peninsular Ranges arc suite, illustrating the similarity of rocks erupted through oceanic and continental crust. The similarity in both trace and major elements indicates that continental crustal assimilation may not be a dominant process in arcs built on continental crust. The field of Aleutian arc rocks (>400 samples) erupted through oceanic crust is outlined and shaded pale orange. Also plotted are upper and bulk continental crust compositions of Rudnick and Gao (2003). Predicted variation trends for fractional crystallization are shown. Data sources are as in Figure 56 plus samples from GEOROC database (http://georoc.mpch-mainz.gwdg.de/georoc/) for Izu-Bonin, Marianas, NE Honshu, and Cascades.

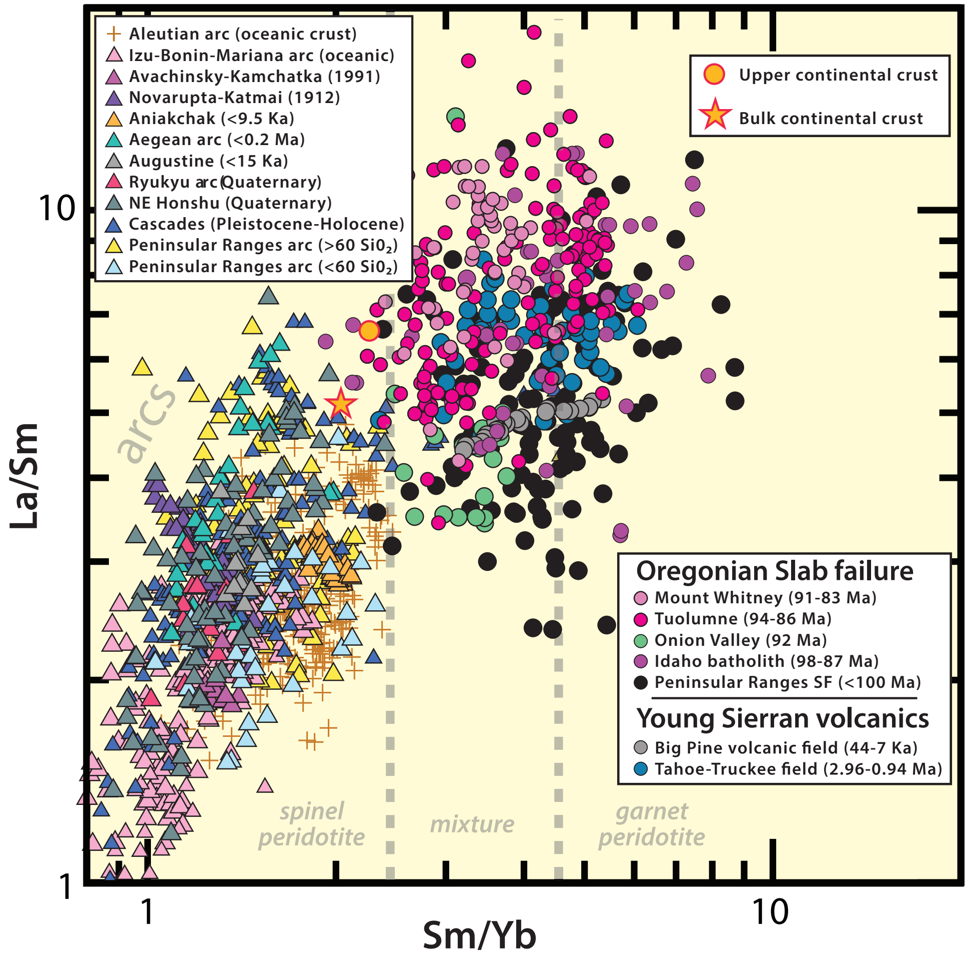


Figure 58. La/Sm vs. Sm/Yb plot for various arc and slab failure igneous suites, illustrating the differences between various arc suites and Oregonian slab failure suites. Virtually all slab failure rocks have Sm/Yb greater than 2.5, whereas most arc rocks have Sm/Yb <2.5. Labeled gray tone subdivisions into different fields that indicate inferred source rocks are derived from the work of Putirka (1999) and based on partial melting models of xenoliths believed to represent sub-Sierran lithospheric mantle. Arc suites are as in previous figures; Mount Whitney data are from Hirt (2007); Tuolumne data are from Memeti (2009); Onion Valley data are from Sisson et al. (1996); Idaho batholith data are from Gaschnig (2015, personal commun.); Peninsular Ranges data are from Lee et al. (2007); Big Pine data are from Blondes et al. (2008); and Tahoe-Truckee data are from Cousens et al. (2011). Estimated compositions of upper and bulk continental crust are from Rudnick and Gao (2003). SF—slab failure.

eastward (Jean et al., 2014). Note that the lower crust beneath the Snake River Plain is old and radiogenic, with $^{87/86}Sr$ as high as 0.83 and epsilon Nd values ranging from –20 to –50, as deduced from xenoliths (Leeman et al., 1985), and so if the deep mantle melts, which contained very low Rb concentrations (Camp et al., 2003), interacted with this crust in any appreciable way, it would be readily apparent.

These results are consistent with isotopic analyses and trace-element geochemistry obtained from the Big Pine volcanic field located along the eastern Sierra Nevada (Ormerod et al., 1991; Blondes et al., 2008); isotopic ratios and geochemistry from the Tahoe-Truckee volcanic field in the northern Sierra (Cousens et al., 2011; Farmer et al., 2013); and isotopic ratios and geochemistry from the Miocene–Pliocene volcanics in the southern Sierra Nevada (Farmer et al., 2002; Putirka et al., 2012) and from ultramafic xenoliths in young Sierran basalts (Domenick et al., 1983; Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1998; Lee et al., 2001a, 2001b; Chin et al., 2012). All existing data are consistent with our overall model that upwelling peridotitic magmas provided sufficient heat to melt metabasalt beneath or within the subcontinental lithospheric mantle. The garnet signature was maintained because the lithosphere was old and sufficiently thick such that deeper melts were unable to traverse and interact with substantial thicknesses of spinel peridotite, although the Sm/Yb ratios (Fig. 58) indicate that there was some interaction.

A probable scenario is that basaltic melts from the upwelling peridotite ponded at the base of the lithosphere, as in the model of Sisson et al. (1996) for the Onion Valley hornblende gabbro. If basaltic melts derived from depleted peridotite arrived in significant quantity at the base of the lithosphere, or even infiltrated it, their crystallates could be remelted, as suggested by Coleman et al. (1992), to produce the common granodiorites of the slab failure suite. Where lithosphere was thick, the source rocks would contain garnet but not plagioclase. The possibility that mantle peridotite may be highly variable and contain substantial amounts of pyroxenite, as indicated by Lambart et al. (2016), makes exact solutions difficult.

Support for our model for much deeper derivation of slab failure magmas also comes from the Coast plutonic complex of British Columbia. There, the Oregonian slab failure rocks of the Coast plutonic complex of British Columbia have a similar slab failure trace-element signature to the Sierra–Peninsular Ranges (Fig. 34) but are markedly different in terms of Sr and Nd isotopes, which in the Coast plutonic complex are much more closely allied with

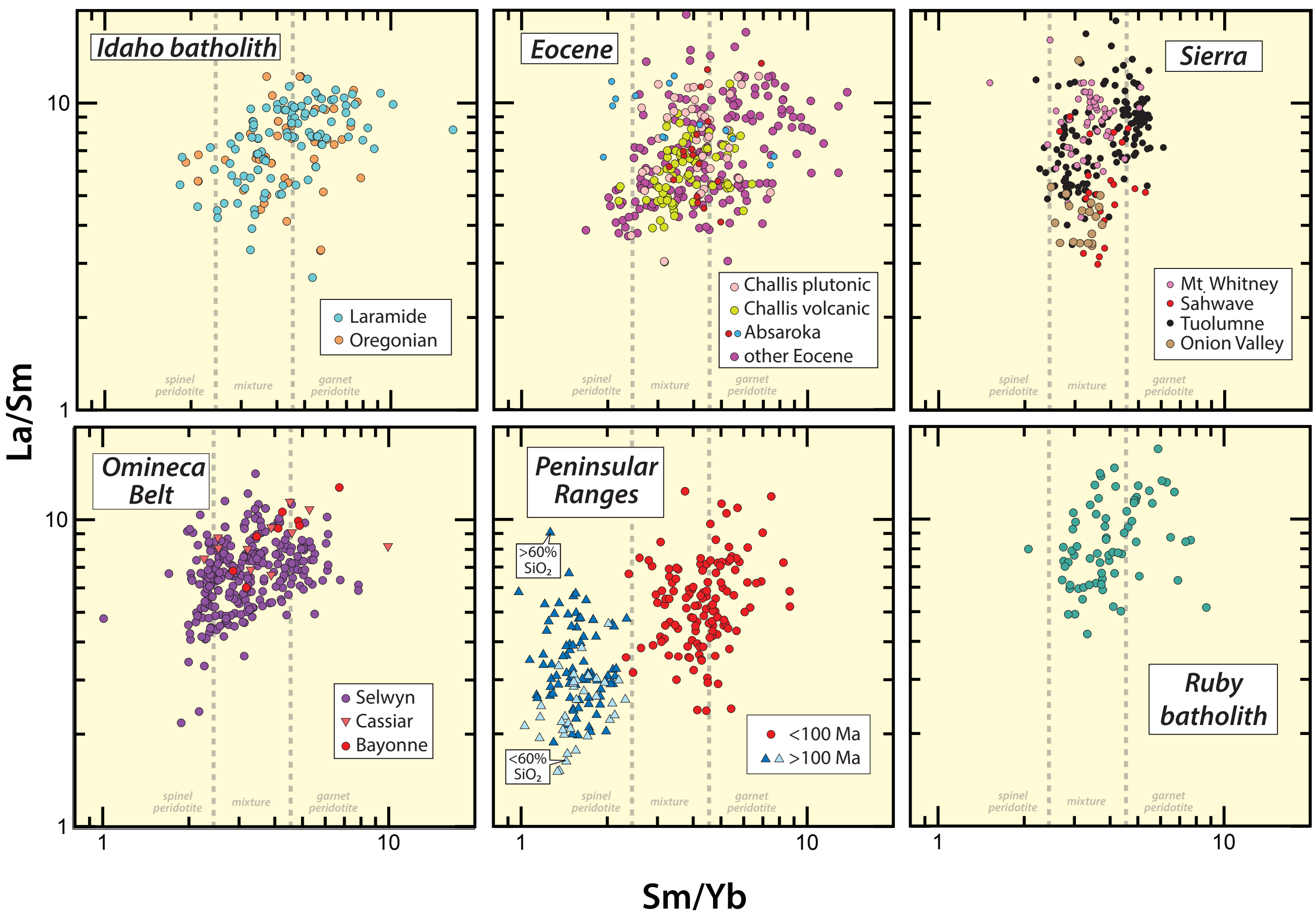


Figure 59. Various slab failure igneous suites plotted on our La/Sm vs. Sm/Yb diagram, illustrating that the suites are dominated by Sm/Yb >2.5, which separates them from arc rocks, which are characterized by Sm/Yb <2.5. We included the pre–100 Ma arc suite from the Peninsular Ranges for comparison. Data are from sources cited in the text. See Figure 56 for additional arc suites. Labeled gray tone subdivisions into different fields that indicate inferred source rocks are derived from work of Putirka (1999) as discussed in text.

asthenospheric melts of arcs (Fig. 53). This suggests to us that the rising asthenospheric melts in the Coast plutonic complex did not interact with old enriched subcontinental lithospheric mantle, but that the subcontinental mantle lithosphere was nevertheless sufficiently thick that the trace-element signatures reflect melting within the plagioclase-free, garnet stability field.

Zoned Intrusive Complexes and the Absence of Crustal Input

If we are correct that slab failure magmas are derived at least in part from subcontinental lithospheric mantle or basalt that accumulated within it or ponded at its base, then zoned intrusive complexes such as the Tuolumne, Mount Whitney, and San Pedro de Martir, where plutons are stacked one beneath the other, and so traversed the same mantle and lithospheric column as they rose, provide sequential sample sets to track progressive mantle involvement and evaluate possible crustal input. The Tuolumne is well dated (Fig. 18) and has the most extensive geochemical and isotopic data set (Memeti, 2009), so we utilized it as a template to better understand the broader problem of the generation, rise, and emplacement of slab failure magmas.

The Tuolumne intrusive complex (Bateman and Chappell, 1979; Bateman, 1992), or suite as it is often called, is composed of four intrusions that get younger toward the center (Fig. 18): Kuna Crest, Half Dome, Cathedral Peak, and Johnson granite porphyry, from margin to core. While all previous workers have considered the intrusions to be nested, Hildebrand (2014) argued that they and all other plutons of the Sierra Nevada were intrusive sheets folded into type 1 interference patterns by younger deformation. The exact shape of the bodies is unimportant here, and the salient feature of the complex is that it contains four well-dated plutons that are younger inward. Here, we do not consider the youngest and most central body, the Johnson granite porphyry, because it is a fractionated leucogranite with abundant evidence of fluid interaction.

Originally, Bateman and Chappell (1979) argued that the compositional zoning within the complex resulted from crystal fractionation of a single voluminous influx of magma. However, subsequent isotopic work (Kistler et al., 1986) ruled out the possibility of relating the compositions to any sort of fractionation scheme, and U-Pb zircon age determinations demonstrated that the complex was emplaced over 10 m.y. from 95 to 85 Ma (Coleman et al., 2004). This not only ruled out the closed fractionation model, but also the two-component mixing scheme favored by Kistler et al. (1986). Instead, Coleman et al. (2004) argued for incremental intrusion of stacked sheets by a crack-seal mechanism. Subsequent U-Pb dating of zircon by several workers confirmed the results of Coleman et al. (2004), and most U-Pb dates are compiled on the geologic sketch map (Fig. 18).

Although Bateman and Chappell (1979) provided an excellent synopsis of the petrographic variations within the complex, recent work by Gray (2003) showed that many textural features originated during late-stage, postmagmatic thermal maturation, as documented by mineral chemistry that indicates significant recrystallization (Gray et al., 2008). This may mean that studies of "magmatic" textures and mineral chemistry in plutons might not yield much information on the origin of the pluton. Because closed system fractionation in each of the three plutons was relatively limited (Fig. 26), the overall geochemical variations in the complex probably arose well below the level of emplacement (Coleman et al., 2012). We therefore focus on the overall geochemistry.

U-Pb ages (Fig. 18) tell us that there were a minimum of four main magmatic pulses that formed the complex. There may have been many more, and local field evidence suggests that there were (Coleman et al., 2005, 2012), but available geochronology and geochemistry are not sufficiently detailed to resolve finer features throughout the complex. Here, we plotted modern geochemistry and isotopic ratios from the oldest three plutons on variation diagrams (Figs. 60–62).

Major elements versus SiO_2 are generally well behaved and form linear trends on variation diagrams (Fig. 60), with Na_2O and K_2O being the major exceptions. Note that as SiO_2 increases, both Na_2O and K_2O increase, which is atypical.

Trace-element plots are informative and, for the most part, except for transition metals such as V, are decoupled from most major-element trends (Fig. 61), as noted by Gray et al. (2008). We used Sm/Yb as a measure of melting depth and plotted this ratio versus our typical slab failure discrimination ratios, La/Yb, Sr/Y, and Nb/Y (Fig. 62), finding a marked positive correlation with decreasing age of intrusive phases. We also found the Na/Ti ratio, another measure of depth of melting in the mantle (Putirka, 1999; Spandler et al., 2008), to increase with Sm/Yb, as expected (Fig. 62). Interestingly, both Sr and Nd isotopes correlate with our depth-of-melting proxies (Fig. 62), as both are more evolved in younger plutons. Wide variations in Rb/Sr, Sr/Ba, Zr/Hf, and La, Rb, and Yb are observed, but they vary consistently with decreasing age (Fig. 61).

Although it appears, based on Sm/Yb and Na/Ti, that the youngest body, the Cathedral Peak pluton, was derived from deeper levels than the older bodies, that is not necessarily the case. As partial melts rise, the resultant more deeply derived melts are progressively diluted, not only by fractional melting at shallower depths, but also by incorporating melts from the margins of the melt column (O'Hara, 1985). If the magma column remains fixed over time, the more easily melted fractions of the walls will be "used up," and so there will be less material influx into the melt column. Thus, even though the geochemistry suggests that younger bodies were derived from greater depth, it is not necessarily so if the magma used the same pathway through time.

Oxygen isotopes from the Tuolumne intrusive complex (Fig. 63), as well as nearly all post–100 Ma plutons in the Sierra Nevada (Fig. 55), have values close to, and overlapping mantle values (Lackey et al., 2008), so that significant crustal assimilation and mixing with evolved melts of continental crust are precluded. Instead, as pointed out earlier, Lackey (2005) suggested that the magmas were derived from old lithospheric mantle. Thus, it appears that the variations within the magmas are artifacts from their mantle source regions, perhaps modified during their subsequent rise through the lithospheric mantle, but not from the crust where they were emplaced. Some slab failure suites, such as deep-seated La Posta plutons of the Peninsular Ranges, have $\delta^{18}O$ values >10 (Taylor and Silver, 1978), which indicate possible derivation from altered oceanic crust (Fig. 55), but values from plutons of the Sierran Crest plutonic suite overlap with, or are just slightly higher than, mantle values.

Granodiorites from the Mantle?

If we are correct that radiogenic isotopes in slab failure magmas are derived from enriched subcontinental lithospheric mantle, then the more-evolved nature of Sr and Nd isotopes in the plutons suggests that with time, greater quantities of evolved isotopes are scavenged from the subcontinental lithospheric mantle, as evidenced by the increasingly enriched ratios in younger plutons (Fig. 62). If continental crust is uninvolved in the genesis of Sierran slab failure melts, as indicated by oxygen isotopes, then this presents a paradox, for if crustal processing does not create the high SiO_2, then it must be a characteristic of the melts arriving in the crust, yet how can melts rising out of the mantle be more siliceous than basalt, especially when partial melts of garnet peridotite yield Ti-enriched basalts (Walter, 1998; Davis et al., 2011)? In other words, how do you get granodiorites from sources that produce melts of basaltic composition?

If the $\delta^{18}O$ results correctly reflect a mantle signature, and if asthenospheric mantle peridotite is ruled out on compositional grounds, then there seem to be only four remaining possibilities: (1) melting of metamorphosed and subducted mid-ocean-ridge basalt (MORB); (2) melting and assimilation of inordinately enriched subcontinental lithospheric mantle; (3) melting of older arc or slab failure basalts at the base of, or within, the subcontinental lithospheric mantle; and (4) some combination of 1, 2, and 3. Melting of enriched subcontinental lithospheric mantle can readily produce the isotopic characteristics of Sierran plutons,

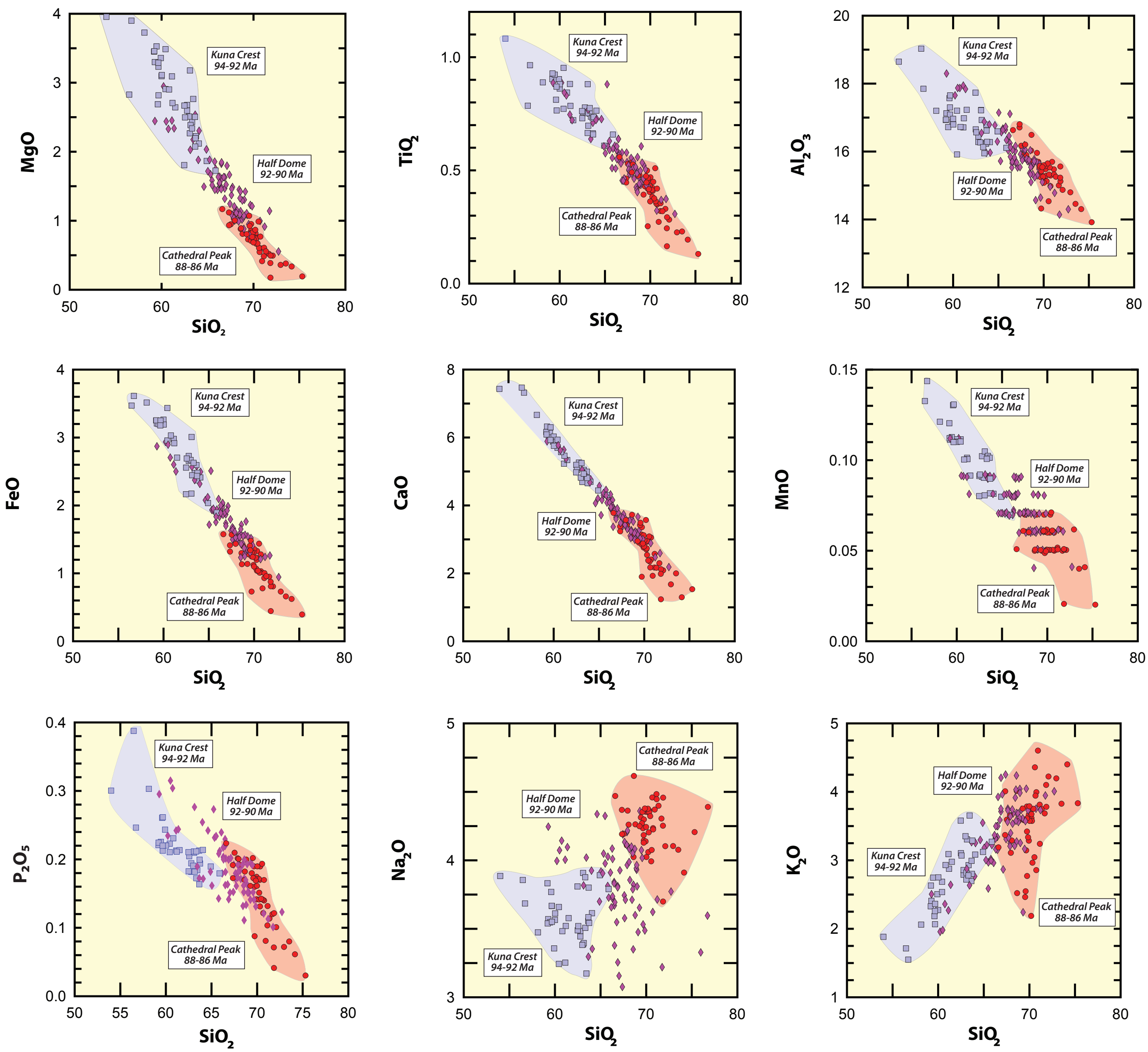


Figure 60. Major-element Harker variation diagrams (in wt% oxide) for Kuna Crest, Half Dome, and Cathedral Peak phases of the 100–84 Ma postcollisional Tuolumne intrusive suite, Sierra Nevada batholith. The Kuna Crest and Cathedral Peak samples are outlined by colored fields. Data are from Memeti (2009).

as we have already seen with compositions of younger volcanic rocks and their xenoliths, but the volcanic rocks are fundamentally basaltic, albeit with varying degrees of alkaline enrichment (Van Kooten, 1980, 1981; Blondes et al., 2008; Putirka and Busby, 2007; Putirka et al., 2012; Farmer et al., 2002, 2013; Cousens et al., 2011). Thus, we are left with the subducted oceanic crust or melting of underplated and/or intermingled metabasalt such as garnet pyroxenite and/or garnet amphibolite.

During the Cretaceous, the oxygen isotopic ratios of seawater and oceanic crust appear to have been the same as today (Muehlenbachs, 1998; Valley et al., 2005), so values determined from MORB today can be used for the Cretaceous. Today's oceanic crust has mantle $\delta^{18}O$ values of ~5.7, but where hydrothermally altered, it has higher values, around 10 (Eiler, 2001; Bindeman et al., 2005). When oceanic crust is subducted, the volatile component might be driven off due to metamorphic reactions,

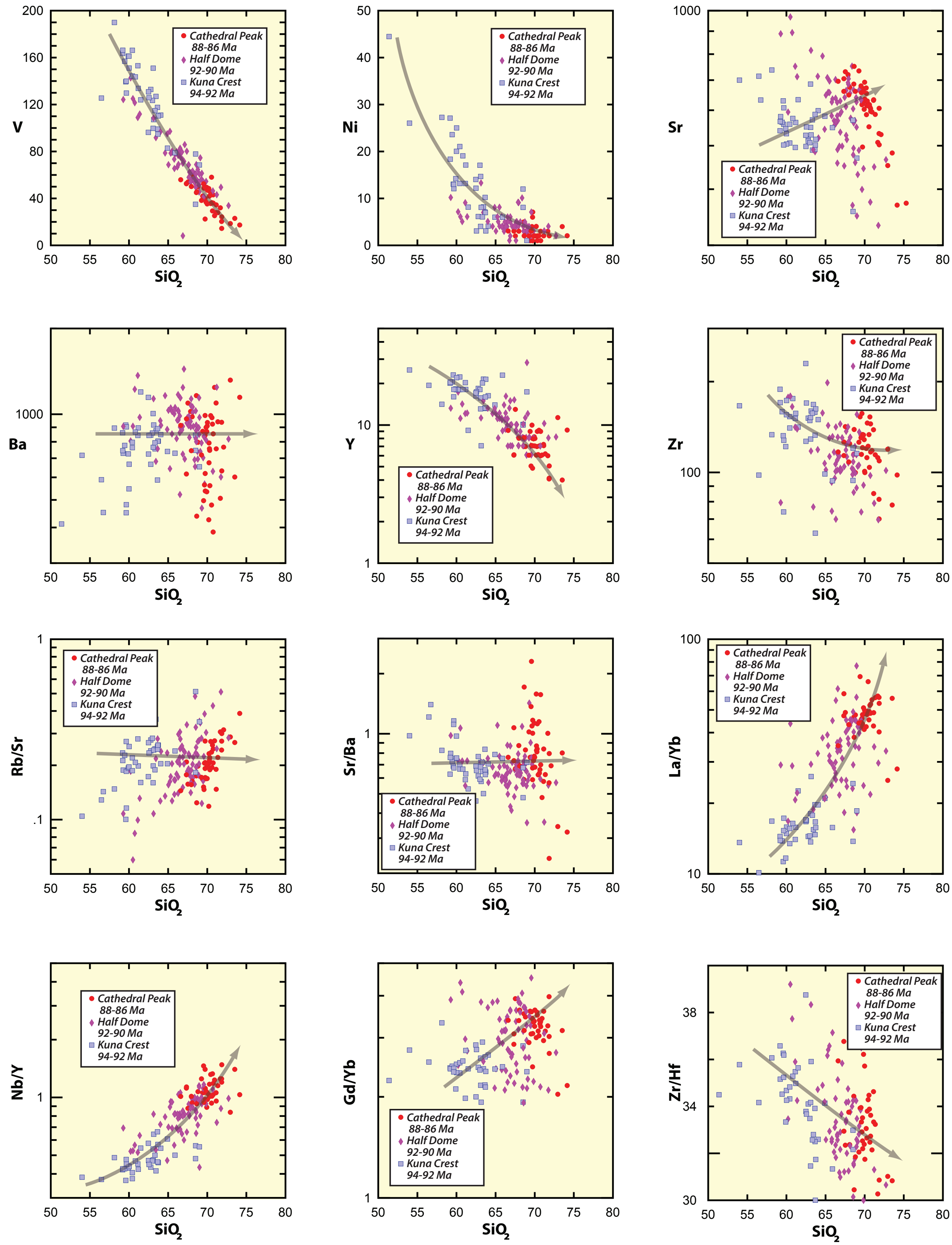


Figure 61. Harker variation diagrams for Kuna Crest, Half Dome, and Cathedral Peak phases of the Tuolumne intrusive suite. Inferred trends with increasing silica and decreasing age are shown by gray lines. Data sources are as in Figure 60. SiO_2 in wt%, others in ppm.

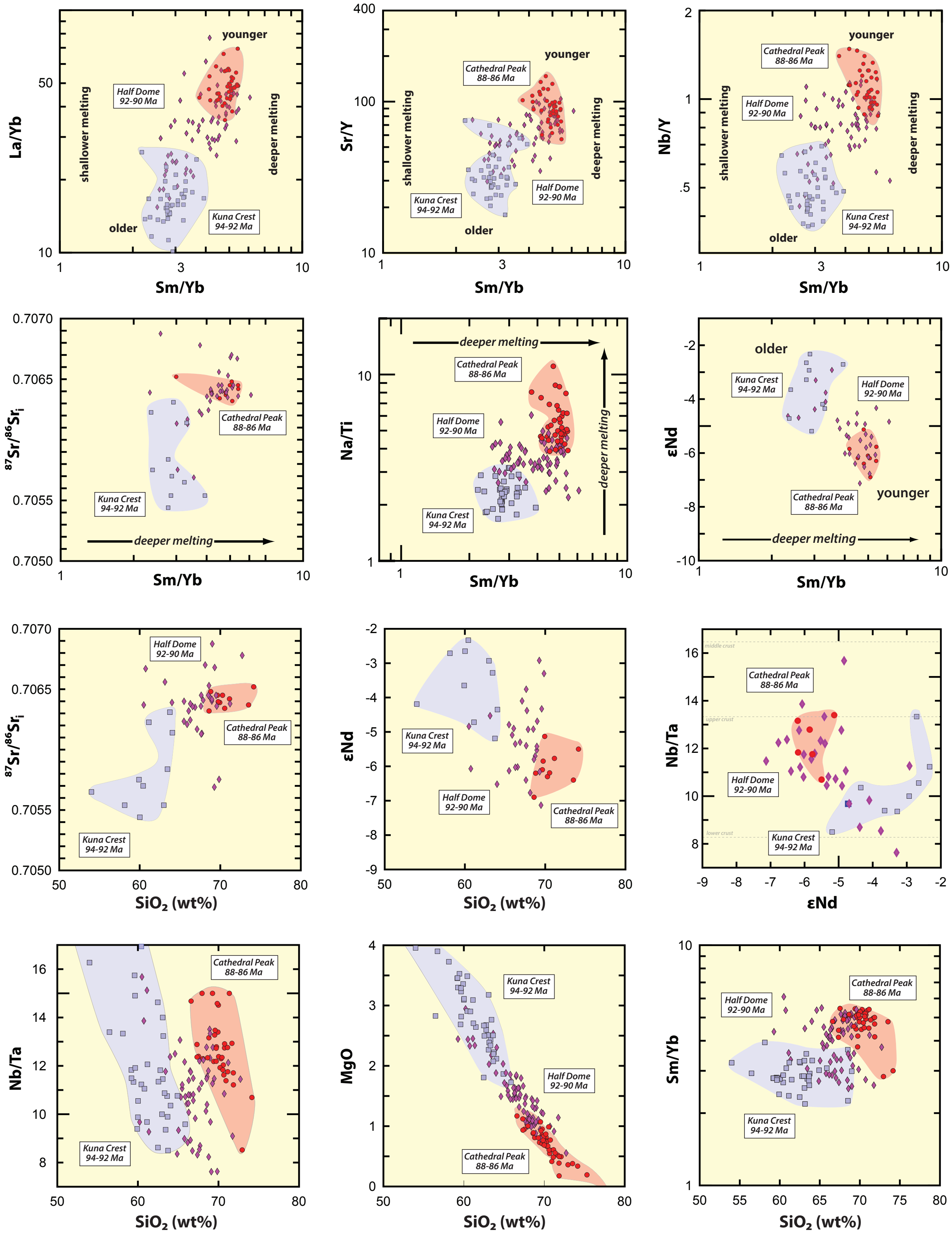


Figure 62. Harker-type plots for 94–92 Ma Kuna Crest granodiorite, 92–90 Ma Half Dome granodiorite, and 88–86 Ma Cathedral Peak granodiorite of the Tuolumne intrusive suite, illustrating decoupling of rare earth elements (REEs), high field strength elements (HFSEs), Nd and Sr isotopes from major-element compositions (see MgO vs. SiO_2 for example). Outlined fields are as in Figures 60 and 61. Data are from Memeti (2009). SiO_2 in wt%.

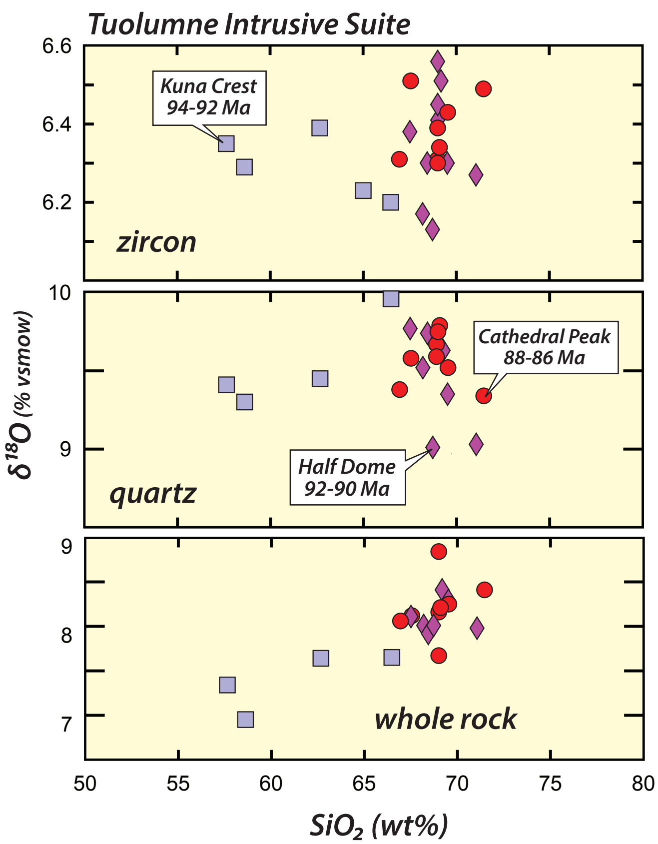


Figure 63. SiO_2 vs. $\delta^{18}O$ for zircon, quartz, and whole rock from Kuna Crest granodiorite, Half Dome granodiorite, and Cathedral Peak granodiorite of the Tuolumne intrusive suite. The $\delta^{18}O$ values approximate mantle values. Data are from Lackey et al. (2008). vsmow—Vienna standard mean ocean water.

but a detailed study of stable isotopes in Grecian eclogite-facies metagabbros and metabasalt showed that minerals from the metabasalts were strongly domainal on the scale of decimeters and had appreciably higher $\delta^{18}O$ values consistent with low-temperature alteration by seawater, while metagabbros had $\delta^{18}O$ values typical of pristine oceanic crust as well as oceanic crust that was altered at high temperature (Putlitz et al., 2000). Additionally, most eclogitic xenoliths in younger basalt generally have values of $\delta^{18}O$ that range from 5 to 6 (Eiler, 2001). Given that most of the 6–7 km thickness of oceanic crust is gabbro that is somewhat heterogeneous, but can have $\delta^{18}O$ as low as 2.5 and commonly in the 4–6 range (Putlitz et al., 2000), bulk oceanic crust might be in the 6–7 range. In any case, depending on the relative amounts of each component and the depth of melting, melts derived from subducted oceanic crust might be expected to have slightly higher $\delta^{18}O$ values than pure MORB values.

Another constraint on the petrogenesis of slab failure magmas is that they do not generally have a Eu anomaly, which suggests that there was no residual plagioclase in the source. In experiments with gabbro, Green and Ringwood (1967) found that at about pressure >2 GPa and temperature >1100 °C, plagioclase did not coexist with garnet and pyroxene. In another set of experiments (Rapp et al., 1991), olivine-normative amphibolite and an alkali-rich basalt yielded minor residual plagioclase at 16 kbar, but exclusively garnet-clinopyroxene-rutile at 22 kbar and above. Thus, at slab failure depths of >100 km, there is unlikely to be plagioclase in eclogite of the descending slab, or metabasalt at the base of, or within, the lower subcontinental lithospheric mantle.

Because hornblende and biotite are ubiquitous in slab failure plutons, the source melts are hydrous, and so breakdown or melting of a hydrous phase, most likely amphibole, should be implicated in the source region. Sisson et al. (1996) concluded that magmas of the 92 Ma Onion Valley hornblende gabbro in the Sierra Nevada contained 4%–6% H_2O and that other magmas there probably contained 3%–4% H_2O. This water could have been produced from dewatering of subducted continental rift and margin sediments.

Several researchers have performed experiments on melting of amphibolite and garnet pyroxenite of MORB parentage and have developed a broad understanding of their partial melts, as well as their major- and trace-element contents at varying degrees of partial melting under various pressures (Rapp and Watson, 1995; Klein et al., 2000; Klemme et al., 2002; Kogiso et al., 2004; Spandler et al., 2008). When MORB is returned to the mantle, it forms eclogite at pressures above 2.0 GPa (Ringwood and Green, 1966). Rapp et al. (1991) examined vapor-absent melting of natural olivine-normative amphibolites, three low-K MORB-like rocks, and an alkali basalt. Resultant melts produced by 10%–40% melting were tonalitic-trondhjemitic at all pressures from 8 to 32 kbar and were highly depleted in heavy rare earth elements (HREEs) with La/Yb of 30–50 when garnet was present in the residue.

Sen and Dunn (1994) used amphibolite as a starting composition for dehydration melting experiments and found that melts at 2 GPa were compositionally similar to adakites and that REE abundances at 10%–15% melting approximated those of adakites. However, they did not produce the high Sr/Y ratios typical of adakites, so they argued that if adakites are partial melts of oceanic crust, then in order to match their distinctive characteristics, they must interact with mantle and/or crust.

Melting experiments by Rapp and Watson (1995) used four different basaltic starting compositions and examined changes in the relative proportions of melt and coexisting residue from 1000 °C to 1150 °C, over pressures from 8 to 32 kbar. They found highly siliceous melts of granitic to trondhjemitic composition at 5%–10% partial melting at 8–16 kbar, but the residue contained plagioclase. At 32 kbar and 1100–1150 °C, they found that trondhjemitic-tonalitic, granodioritic, quartz dioritic, and dioritic partial melts resulted from 20%–40% partial melting and left a garnet-clinopyroxene residue. Thus, they showed that adakites and TTG suites with high Sr/Y and La/Yb could be produced by 10%–40% melting of partially hydrated metabasalt in the presence of garnet between 1000 °C and 1100 °C. Although subduction zones are typically too cool to produce these partial melts, melting of hot, young oceanic crust near spreading ridges (Peacock, 1996) or above and adjacent to zones of deep mantle upwelling would satisfy those conditions.

Direct slab melts should interact with mantle peridotite as they ascend, and Rapp et al. (1999) studied this experimentally by allowing oceanic crustal melts to infiltrate and react with peridotite. At nearly 4 GPa and high melt-to-rock ratios, they found that the interaction produced high Mg# adakites, but at melt-to-rock ratios close to unity, the melts were completely consumed by reaction with the peridotite. Only when additional heat was added to the system did melt remain. Trace-element abundances in hybrid slab melts were higher than in pristine slab melts because melt was progressively consumed by reaction; however, elemental ratios such as Sr/Y and La/Yb remained constant. They also showed that 30% melts of hydrothermally altered MORB closely resemble the trace-element contents of adakites, except for Zr and Ti, which they attributed to residual accessory phases. One of their principal conclusions was that slab-derived melts can metasomatize the overlying mantle peridotite as they are consumed, which means that, although the mantle mineralogy will control the overall composition of younger melts, the incompatible trace elements from the slab will be available to be scavenged from the peridotite by younger melts.

Some of the potentially most applicable melting experiments come from studies of melting to generate oceanic-island basalt. Research of particular importance is the work on silica-deficient and silica-excess pyroxene (Kogiso et al., 2004). Kogiso et al. (2004) looked at phase equilibria and identified a garnet-pyroxene thermal divide in CaO-MgO-Al_2O_3-SiO_2 projections. Partial melts that form on either side of the divide have distinct phase assemblages and produce either nepheline-normative or quartz-normative compositions, depending on which side of the divide they formed. In natural systems, they, as well as Lambart et al. (2013), found the divide to appear at pressure >2 GPa, and the two sides to produce melts that lie either above or below 48% SiO_2 (Fig. 64).

Lambart et al. (2013) followed up on the melting of pyroxenes in the mantle and confirmed that the thermal divide becomes effective at around 2 GPa, but that it depends on the bulk composition, notably FeO, TiO_2, Na_2O, and K_2O contents. They found that interactions between pyroxenite melts and upper-mantle peridotites produced a wide variety of pyroxenites in the upper mantle, and so the various reactions between both might be critical to understanding mantle melts, but there have been few studies of the resultant second-generation compositions.

Thus, tonalitic, trondhjemitic, and even granodioritic melts with the appropriate trace-element characteristics of slab break-off magmas can be generated by melting garnetiferous, plagioclase-absent rocks such as meta-oceanic crust, under appropriate conditions. What is not clear is precisely where the bulk of melting occurs and the geometry of mantle upwelling during slab failure. Precisely how the upwelling mantle interacts with the overlying oceanic lithosphere just prior to complete failure is simply unknown. Also, the experimental data do not rule out the creation of basaltic melts from the upwelling peridotite that could then pond at the base of the lithosphere, or even infiltrate it, as in the model of Sisson et al. (1996) for the Onion Valley hornblende gabbro. They argued that the basaltic melts there had low Mg# and low incompatible trace-element concentrations, so they were neither primitive nor primary, and, on the basis of high-pressure hydrous experimental data, that they resulted from an earlier stage of differentiation. If basaltic melts derived from depleted peridotite arrived in significant quantity at the base of the lithosphere, their crystallates could be remelted, as suggested by Coleman et al. (1992), to produce the common granodiorites of the slab failure suite. Where lithosphere was thick, rocks there would contain garnet and not plagioclase.

Our model for the origin of slab failure magmatism is simple and is fundamentally akin to mantle plume, or "hotspot," magmatism (Morgan, 1972), except that instead of an upwardly migrating cylinder, it is a long, linear upwelling of mantle material. Both involve dynamic upwelling and decompression, which are physical processes, and so we use plumes as an initial proxy for slab failure magmatism.

We envision that once the subducting slab fails, asthenospheric mantle upwells adiabatically through the tear in the downgoing plate to form a linear mantle plume. The rising mantle melts when it intersects the solidus and continues to melt during its ascent (McKenzie and O'Nions, 1991), similar to the way in which melting to create MORB starts in the garnet stability field and continues as mantle rises to shallower levels (Salters and Hart, 1989; Kinzler, 1997; Jean et al., 2010), or similar to mantle plumes, which also involve melting in the garnet stability field, as evidenced by high Sm/Yb ratios (Salters and Hart, 1989; Hirschmann and Stolper, 1996; Niu et al., 1999; Putirka, 1999).

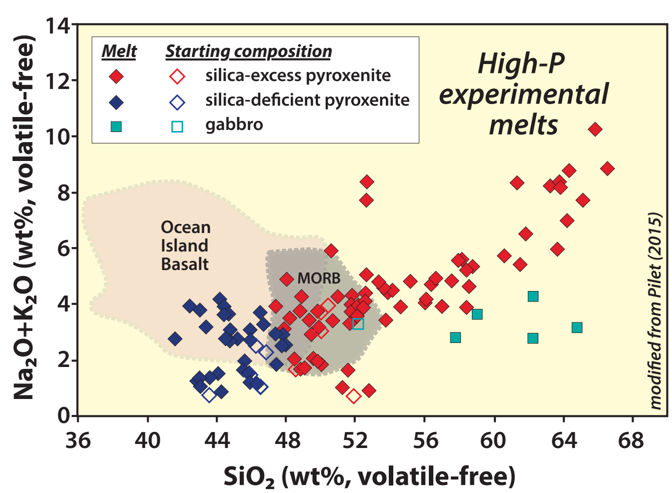


Figure 64. SiO_2 vs. (Na_2O + K_2O) plot of high-pressure (*P*) experimental data for partial melting of silica-excess and silica-deficient pyroxenite and gabbro. Also shown are the compositional fields for oceanic-island basalt (OIB) and mid-ocean-ridge basalt (MORB). Figure is modified from Pilet (2015).

The rising magma stops and perhaps pools as it reaches the base of the lithosphere (Sakamaki et al., 2013), which acts as a lid on the melting column (Ellam, 1992; Putirka, 1999; Humphreys and Niu, 2009; Ferguson et al., 2013), but significant volumes might infiltrate the lower lithosphere, where they react with it and crystallize. Previously accumulated basaltic melts, perhaps largely garnet amphibolite, either within the subcontinental lithospheric mantle or at its base, can then be melted to produce magmas of the requisite composition.

Slab Window Magmatism and Adakites

We noted in our earlier contribution on the Peninsular Ranges (Hildebrand and Whalen, 2014b) that slab failure rocks were compositionally similar to adakites and suggested that our model for slab failure magmatism might shed light on magmatism associated with ridge subduction and formation of slab windows or gaps (Uyeda and Miyashiro, 1974; Dickinson and Snyder, 1978; Thorkelson and Taylor, 1989; Nelson and Forsyth, 1989; Thorkelson, 1996). These features form when the asthenospheric window of a spreading ridge is subducted and, driven by plate divergence and slab pull, widens and descends into the mantle (Thorkelson, 1996). There is evidence that melting beneath mid-oceanic ridges starts in the garnet stability field and then rises to shallow levels within the spinel peridotite field before forming oceanic crust (Salters and Hart, 1989; Hirschmann and Stolper, 1996). In the case of ridge subduction, the ridge is pulled down into the mantle, and so magmas are derived from increasingly greater depths and should be compositionally different than typical MORB, reflecting greater amounts of residual garnet, and they should be compositionally akin to our slab failure magmas. Not only should they show the same geochemical features as slab failure magmas, but they should also display similar Nd and Sr isotopic features, depending on whether or not they rise into ancient, enriched subcontinental lithospheric mantle or not.

Adakites were originally recognized as peculiar high-Sr/Y and high-La/Yb magnesian andesites within arcs (Kay, 1978; Defant and Drummond, 1990). They were generally interpreted to represent melting of the subducting slab (Defant et al., 1991, 1992; Yogodzinski et al., 1995; Martin, 1999; Martin et al., 2005; Gómez-Tuena et al., 2007), but other models, such as delaminated crustal melts (Xu et al., 2010) and underplated basalt (Atherton and Petford, 1993), were also proposed. Largely based on SiO_2 content, Martin et al. (2005) recognized two distinct groups of adakites: (1) a high-SiO_2 group, which they considered to be derived directly from melting of subducted oceanic crust and subsequently modified by reactions with peridotite as they traversed the upper mantle; and (2) a low-SiO_2 group, which they interpreted to represent partial melts of asthenosphere metasomatized by slab melts.

A quick survey of classic adakitic locales reveals that most high-Sr/Y and high-La/Yb adakite-like volcanic rocks were erupted in regions of ridge subduction and slab window creation beneath arcs. Examples include the Aleutians (Levin et al., 2002), Kamchatka (Portnyagin et al., 2005), Japan (Morris, 1995), the Antarctic Peninsula (Hole, 1990), southern South America (Breitsprecher and Thorkelson, 2009), Panama (Johnston and Thorkelson, 1997), mainland Mexico (Orozco-Esquivel et al., 2007), Baja California (Saunders et al., 1987), and the western United States (Atwater and Stock, 1998; Cousens et al., 2011). When spreading ridges are subducted, melting will take place progressively deeper and will ultimately occur within the garnet peridotite field. Therefore, the resultant magmas are likely to be dominated by a garnet signature and resemble slab failure magmas. Melting of garnet pyroxenite, that is, meta-MORB, should also yield grossly similar partial melts, but they apparently lack the high-Sr/Y values of slab failure rocks (Sen and Dunn, 1994). Given the high heat flow of spreading ridges, it is likely that the resultant magmas are various mixtures of pyroxenite and peridotite sources (Yaxley and Green, 1998; Rapp et al., 1999). In order to evaluate these concepts, we plotted Neogene examples of magmatism attributed to ridge subduction on our slab failure diagrams.

Rocks of the Antarctic Peninsula preserve a long history of arc magmatism, followed by a period of late Miocene–Holocene alkalic magmatism (Smellie, 1987), which is best explained by ridge subduction (Barker, 1982; Breitsprecher and Thorkelson, 2009). Hole (1988, 1990) studied the geochemistry of the youngest rocks and suggested that the magmas were derived by low degrees of partial melting of a garnet-bearing source beneath a slab window. We plotted analyses of rocks younger than 6 Ma on some of our standard plots (Fig. 65), where they appear similar to our slab failure suites. Nd and Sr isotopes suggest that old, enriched subcontinental lithospheric mantle is not present beneath the peninsula.

Within southern South America, the Chile Ridge, a spreading ridge located between the Nazca and Antarctic oceanic plates, was subducted and overridden by the South America plate starting at ca. 24 Ma (Ramos and Kay, 1992; Gorring et al., 1997; Breitsprecher and Thorkelson, 2009). Resultant magmatism, occurring in Argentina, migrated northward with time and formed two discrete groups: (1) the 24–12 Ma Austral volcanic zone of southernmost Argentina, and (2) the Pliocene and Quaternary basalts of the Patagonian plateau (Stern and Kilian, 1996; Stern et al., 1990). The younger plateau basalts were divided into two stages by Gorring and Kay (2001): a voluminous 12–5 Ma sequence of tholeiitic lavas, and a 7–2 Ma less-voluminous sequence of alkaline lavas, both of which they attributed to melting in the garnet stability field. We plotted geochemical and isotopic data from the rocks of the Austral volcanic zone and rocks of both plateau sequences on Figures 66 and 67. While the trace elements plot in expected fields on our plots, the isotopic data were surprisingly primitive, considering that at least the eastern parts of the volcanic field were supposedly erupted through Proterozoic lithosphere of the Deseado massif. The most likely interpretation for the lack of a strong subcontinental lithospheric mantle signature is that intense Jurassic extension and mafic to siliceous magmatism (Panza and Haller, 2002), related to rifting and subsequent

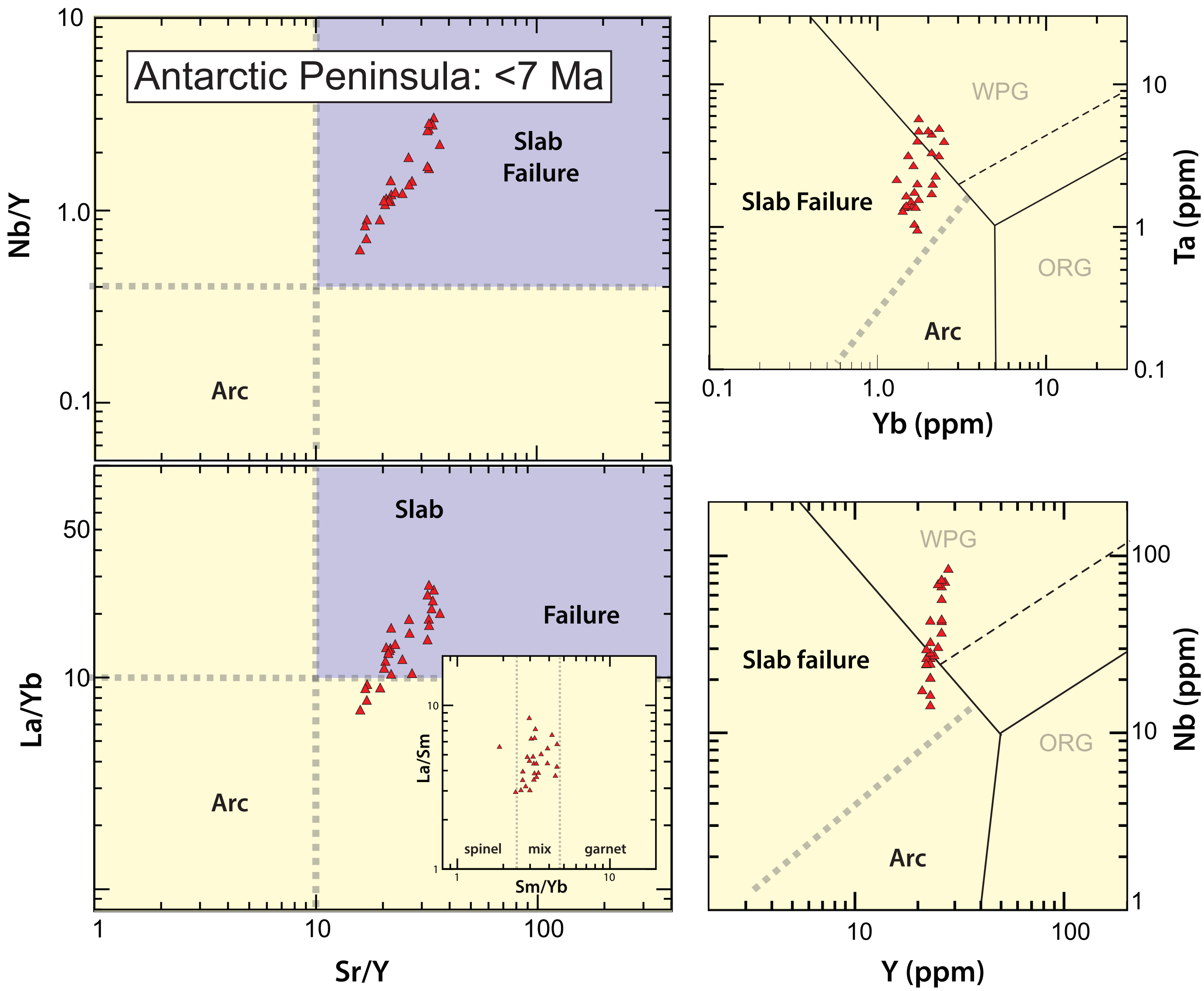


Figure 65. Antarctic Peninsula <7 Ma slab window mafic volcanic rocks plotted on our slab failure discrimination diagrams. Data are from Hole (1988, 1990). WPG—within-plate granite; ORG—ocean-ridge granite.

opening of the South Atlantic Ocean, severely modified any old, enriched subcontinental lithospheric mantle. Interestingly, one of the oldest volcanic rocks in the field, a 12 Ma adakitic lava dome, contains mostly Paleozoic and Mesozoic xenocrystic zircons and only a few older grains—all interpreted to be detrital zircons scavenged from subjacent basement (Orihashi et al., 2013).

During the late Miocene, the Cocos-Nazca Ridge was subducted beneath the Central American arc, which led to marked changes in magma compositions within the arc at ca. 6 Ma (Defant et al., 1992; Johnston and Thorkelson, 1997; Abratis and Wörner, 2001). Compositional changes, which some have attributed to hotspot magmatism (Gazel et al., 2009) and others to formation of a slab window (Abratis and Wörner, 2001), occurred mainly within the Panamanian sector of the arc, with transitional compositions extending to Costa Rican volcanoes. We plotted whole-rock analyses from Panamanian volcanic rocks younger than 6 Ma on our standard slab failure plots (Fig. 68). Most samples plot within our slab failure fields, suggesting that they were derived from garnet-bearing rocks or mixtures of garnet and spinel peridotite, although a few samples show arc affinities and were likely derived mostly from spinel peridotite. Nd and Sr isotopic ratios for these rocks are typical of those that rose through depleted subcontinental lithosphere, and they generally match those of arc magmas to the northwest in Costa Rica (Fig. 69).

Within the Trans-Mexican volcanic belt, magmatic compositions also changed abruptly during the late Miocene at ca. 7.5 Ma, "when calc-alkaline subduction-related magmatism was replaced by mafic, alkaline OIB–type volcanism" (p. 149),

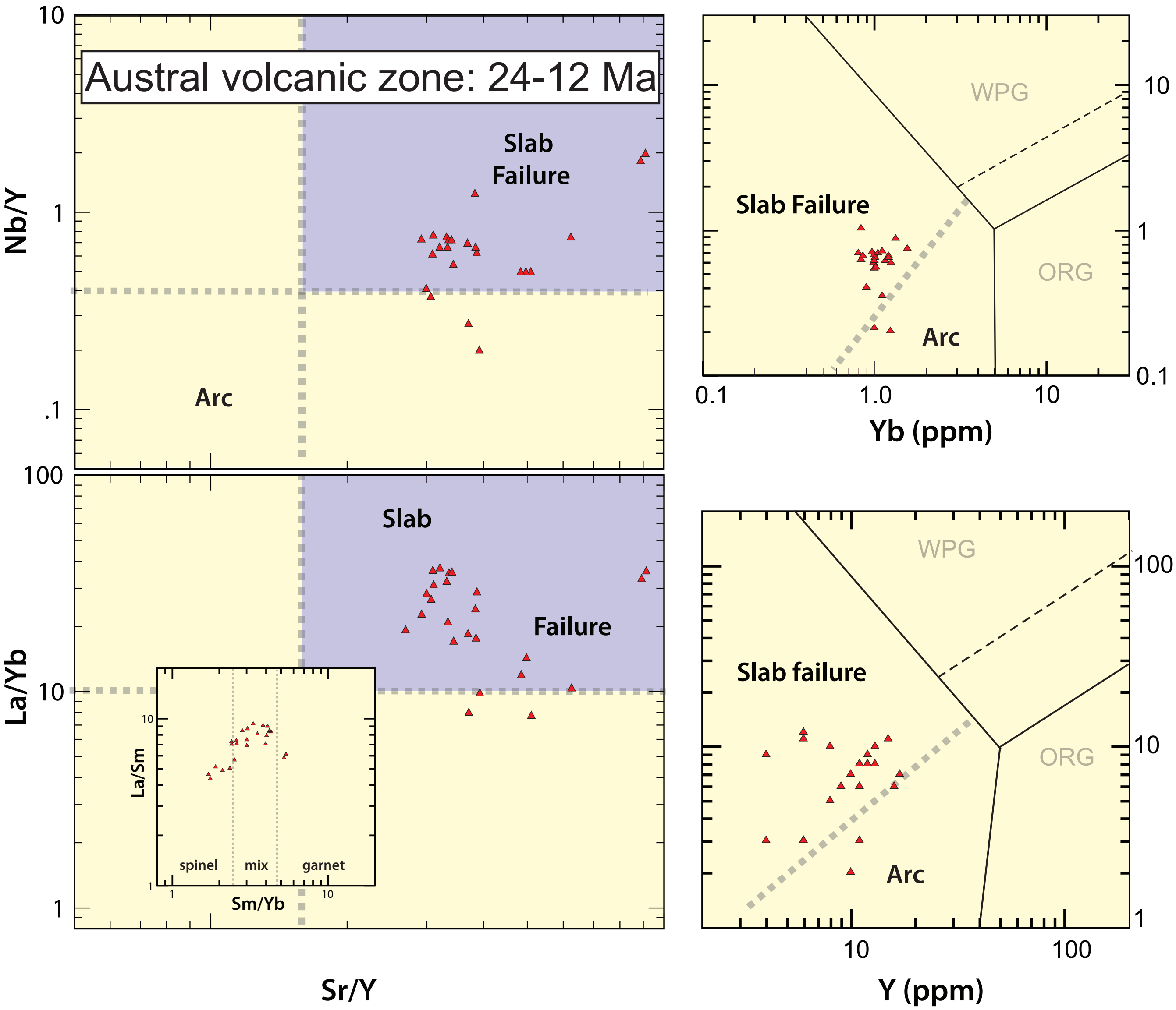


Figure 66. Austral volcanic zone 24–12 Ma slab window mafic volcanic rocks plotted on slab failure discrimination diagrams. Data are from Stern and Kilian (1996) and Stern et al. (1990). WPG—within-plate granite; ORG—ocean-ridge granite.

related to slab failure or development of a slab window along the northern margin of the Cocos plate (Ferrari, 2004; Orozco-Esquivel et al., 2007). We plotted trace-element data from the Mexican examples younger than 7.5 Ma on our standard plots (Fig. 70), where they display geochemical characteristics more typical of slab failure magmatism than arc magmatism. Like the Panamanian example, Nd and Sr isotopic analyses suggest a lack of old, enriched subcontinental lithospheric mantle beneath that part of Mexico (Fig. 69).

The island of Honshu in Japan contains both arc rocks and adakites, with arc rocks dominating the northeastern part of the island (Kimura and Yoshida, 2006) and slab window rocks dominating the southwest (Morris, 1995). We plotted both suites on Figure 71, where the differences between the two suites are obvious and typical. While the southwestern rocks are commonly considered to be part of the magmatic arc, they are compositionally very different and were likely generated by subduction of the spreading ridge in the Shikoku Basin (Okino et al., 1994) between the Izu-Bonin arc and the Kyushu-Palau remnant arc. High-Mg andesites of the Setouchi volcanic belt were erupted during the Miocene and may represent the initial products of ridge subduction (Tatsumi, 2006). The arc rocks of northeastern Honshu have primitive Nd and Sr isotopic ratios, except for the southernmost volcanoes, which erupted magmas with more evolved ratios suggestive of old enriched subcontinental lithospheric mantle (Fig. 69). These southern volcanoes are one of the

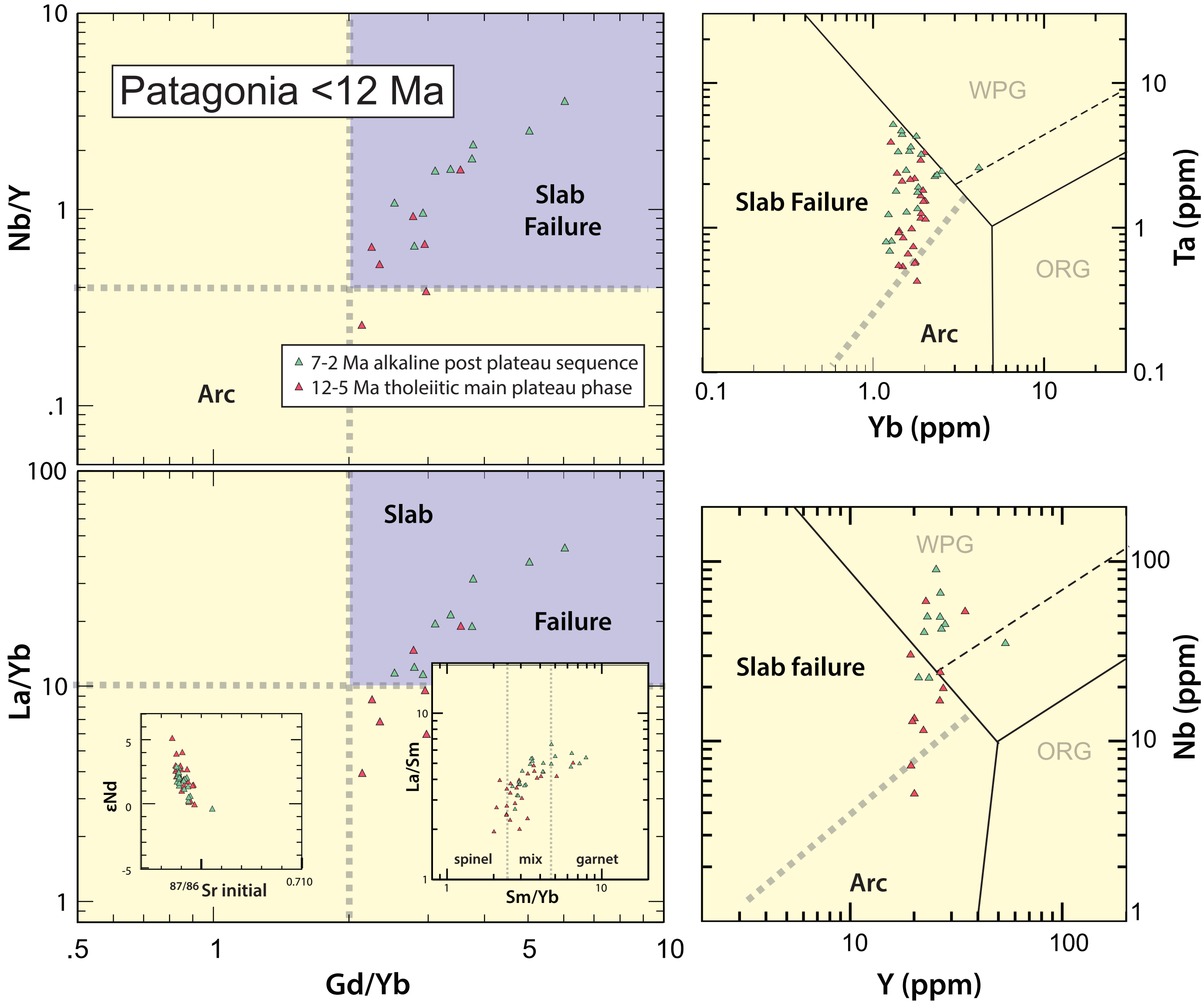


Figure 67. Patagonian slab window mafic volcanic rocks (<12 Ma) plotted on our slab failure discrimination diagrams. Data are from Gorring and Kay (2001). WPG—within-plate granite; ORG—ocean-ridge granite.

few places where arc rocks have markedly negative epsilon Nd and evolved Sr isotopic systematics. They might be influenced by melts derived from a slab tear related to the collision of the Izu-Bonin arc with Japan.

Another possible occurrence of a slab window exists in eastern California, where the northerly march of the Mendocino edge of the Juan de Fuca plate beneath western North America allowed asthenosphere to well up in its wake (Atwater and Stock, 1998). While the edge is typically drawn as a more or less straight E-W line (Engebretson et al., 1985; Atwater and Stock, 1998; Cousens et al., 2011; du Bray et al., 2014), the precise configuration of the slab window and its position through time are poorly constrained. First, this is because several smaller oceanic plates, such as the Monterrey and Arguello, were situated near the triple junction of the Juan de Fuca, Pacific, and Farallon plates (Atwater and Stock, 1998), so that there might have been irregular-shaped windows of indeterminate size and orientation (see, e.g., Dickinson, 1997).

Second, following the Laramide event, the southern margin of North America apparently nearly trended ~300° (current coordinates), based on geological arguments by Hildebrand (2015) and crustal thickness variations determined by Shen and Ritzwoller (2016). This geometry provides actualistic explanations for both the northerly migrating magmatism during the Cenozoic in the southern Arizona–eastern California area (Glazner and Supplee, 1982), and coeval, but opposite, south-southwestward migration of magmatism throughout the Great Basin (Henry and John,

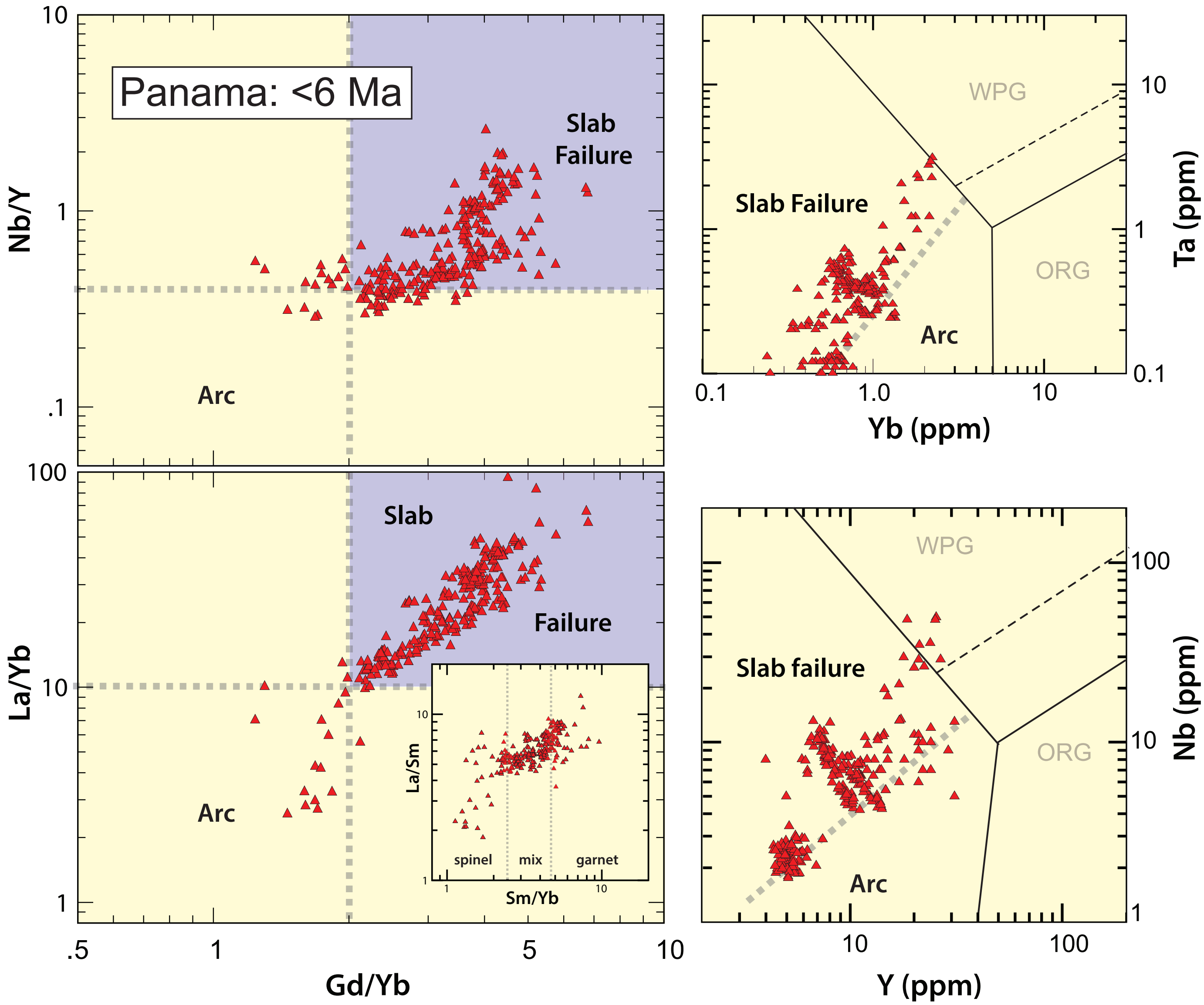


Figure 68. Panamanian slab window mafic volcanic rocks (<6 Ma) plotted on our slab failure discrimination diagrams. Data are from GEOROC, filtered by latitude (http://georoc.mpch-mainz.gwdg.de/georoc/). WPG—within-plate granite; ORG—ocean-ridge granite.

2013). In this scenario, there must have been a northerly trending break in the lower-plate lithosphere to accommodate the contrasting magmatic migrations in adjacent regions of western North America (Fig. 72). By restoring the slip on the San Andreas and related faults (Powell, 1993), as well as extension in the Mojave Desert and Colorado River corridor, the two regions of opposed magmatism were separated along strike of the southern North American margin, which suggests that the ultimate cause of the two magmatic zones was a lower-plate phenomenon. Although this break was located within the lower, descending plate, it likely manifested itself in the upper North American plate as the southward continuation of the zone along which most of the Cordillera migrated northward after 70 Ma (Hildebrand, 2015).

Within the western region, magmatism—presumably related to initial impingement and subduction of the East Pacific Rise and associated ridges related to the Arguello and Monterey microplates—occurred during the mid-Miocene in what is now southern California–northern Mexico (Hurst, 1982; Dickinson, 1997; Atwater and Stock, 1998). Pliocene–Pleistocene magmatism considered to be asthenospheric upwelling along the southern edge of the Juan de Fuca plate occurred in the Lake Tahoe–Truckee area of east-central California. If correct, then the slab windows must have passed between the two areas. For these reasons, we looked at the geochemistry and Nd-Sr isotopes of several Neogene volcanic suites in eastern California, including the Miocene–Pliocene volcanic rocks (Fig. 73) of the southern Sierra

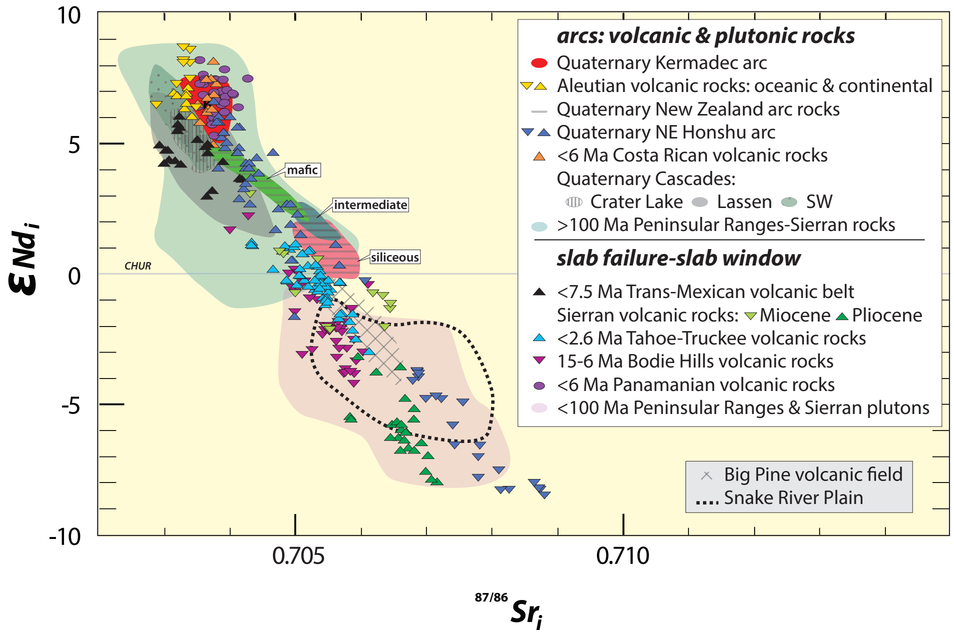


Figure 69. ε_{Nd} vs. $Sr_{initial}$ plot for arc, slab failure, and slab window volcanic and plutonic suites, illustrating the differences and similarities between slab failure/window and arc rocks in our studied suites. Slab window/failure rocks may have values both above and below chondritic uniform reservoir (CHUR), whereas arc rocks are dominated by positive ε_{Nd}. The more isotopically evolved slab failure/window rocks involved old enriched subcontinental mantle lithosphere. New Zealand and Kermadec data are from Gamble et al. (1993), and Honshu data are from Kimura and Yoshida (2006). Fields for Quaternary Cascades are from Bacon et al. (1994).

Nevada (Farmer et al., 2002; Putirka and Busby, 2007; Putirka et al., 2012), the 15–6 Ma volcanic rocks (Fig. 74) of the Bodie Hills (John et al., 2012; du Bray et al., 2016), and Pliocene–Pleistocene volcanic rocks (Fig. 75) of the Lake Tahoe–Truckee area (Cousens et al., 2011). The data plot mostly in our slab failure fields and are evolved with respect to Nd and Sr isotopes (Fig. 69), as are slab failure magmas of the Sierran Crest magmatic suite (Fig. 23), which indicate that the slab window magmas interacted with enriched subcontinental lithospheric mantle.

In summary, Neogene adakites appear to be mixtures of slab and asthenospheric melts caused by melting of hot oceanic crust adjacent to a spreading ridge and varying degrees of mantle-derived melts (Kay et al., 1993). On the basis of slightly elevated oxygen isotopes from adakites in many of the locations discussed herein, Bindeman et al. (2005) suggested that if adakites come from slab melting, they must be mixtures of basalt and gabbro, or if not, they might be melts derived from melting of basalt and/or related mafic cumulates, such as plagioclase-poor garnet amphibolite, that ponded at the base of the crust.

The process of ridge subduction generates a descending slab window that allows melts, ultimately derived from both garnet peridotite and garnet pyroxenite, to rise into the arc. Melting of eclogitic oceanic crust is implicated in the generation of high-SiO_2 adakites, whereas low-SiO_2 rocks are derived from garnetiferous mantle peridotite. The thickness of the lithosphere likely plays an important role: Where it is thick, rising melts might have little or no chance to interact with mantle peridotite, but where it is thin, melts of garnetiferous rock would be diluted by melts of spinel peridotite. Just as with slab failure melts, perhaps slab window melts are sufficiently voluminous that they are able to maintain their garnet signature even as they rise through sections of spinel peridotite. In cases where the lithosphere is thick, cumulates and basalt derived from arc magmatism (Sisson et al., 1996; Bindeman et al., 2005), and ponded at the base of the lithosphere, might be garnetiferous, and rising asthenosphere might provide sufficient heat to melt them. While there might well be exceptions, the compositional similarities between slab failure and slab window magmatic rocks are striking, and it appears that adakites are related to our slab failure suite of magmas developed at convergent margins.

Origin of Archean TTG Suites

Archean sodic leuco-granitoids, usually termed the TTG series or suite, have been estimated to represent at least two thirds of surviving Archean continental crust (Condie, 1981; Martin, 1994). Recognition of the geochemical similarities between the Archean TTG series and potential modern equivalents (adakites; Martin, 1986, 1987; Defant and Drummond, 1990; Drummond and Defant, 1990) spawned an enormous body of literature concerning TTG petrogenesis and implications for the crustal development of Earth (Drummond et al., 1996; Martin, 1994, 1999; Martin and Moyen, 2002; Martin et al., 2005, 2014, and references therein). Key geochemical features of Archean TTG suites and modern adakites are high La/Yb and Sr/Y values, which was termed the adakitic signature by Moyen (2009). Most researchers accept that the distinguishing geochemical features of TTGs and adakites reflect partial melting of mafic protoliths under garnet-stable, plagioclase-unstable pressure-temperature conditions, but how and where this occurred remain controversial. Using concepts developed to explain adakites in modern arcs, some workers have argued that TTGs formed from partial melting of garnetiferous oceanic lithosphere beneath arcs (Martin, 1986, 1994, 1999; Drummond and Defant, 1990). Other researchers suggested that they were generated in thick continental crust similar

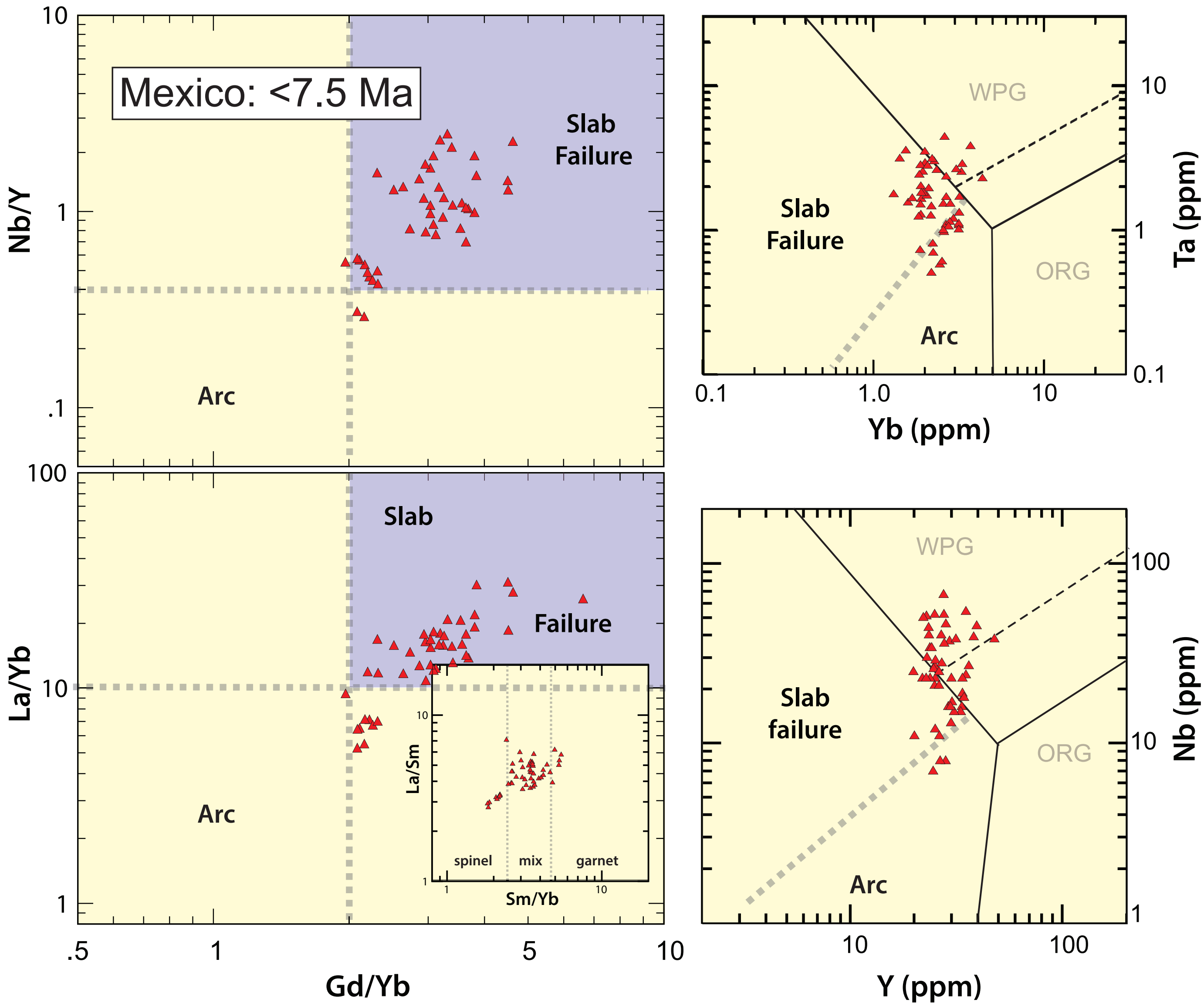


Figure 70. <7.5 Ma slab window mafic volcanic rocks from Mexico plotted on our slab failure discrimination diagrams. Data are from Orozco-Esquivel et al. (2007). WPG—within-plate granite; ORG—ocean-ridge granite.

to that considered by many workers to exist beneath Cordilleran batholiths (Whalen et al., 2002).

At its core, the "slab-melt" TTG petrogenetic model employs partial melting of a "conveyer-belt–like" supply of subducted oceanic crust to generate the preserved voluminous Archean TTG suites. This model, based on modern plate tectonics, implies an ongoing process that occurred at destructive plate margins over extended time periods, providing continuous upwelling of TTG melt. A feature not addressed in this model is how these increments of slab-melt–derived TTG material were amalgamated into the voluminous plutons or batholiths that form over two thirds of Archean cratons. In portions of Archean cratons where careful geological mapping is supported by high-precision U-Pb zircon analyses, such as the Wabigoon portion of the Western Superior craton, entire TTG suites were shown to have been emplaced over periods of time as short as 2–3 m.y. (Whalen et al., 2002, 2004). This indicates that voluminous TTG magmatism formed during short-lived magmatic events or pulses rather than as products of a long-lived ongoing process, as implied in the continuous slab-melting model. Furthermore, where exposures permit, it can be seen that Archean greenstone belts were intruded by, or are in tectonic contact with, voluminous TTG bodies, and were not deposited unconformably upon them (Folinsbee et al., 1968; Myers and Watkins, 1985; McGill and Shrady, 1986; Lamb, 1987; Kusky, 1989, 1998; Barley and Groves, 1990; Kusky and Hudleston, 1999; Percival et al., 2012;

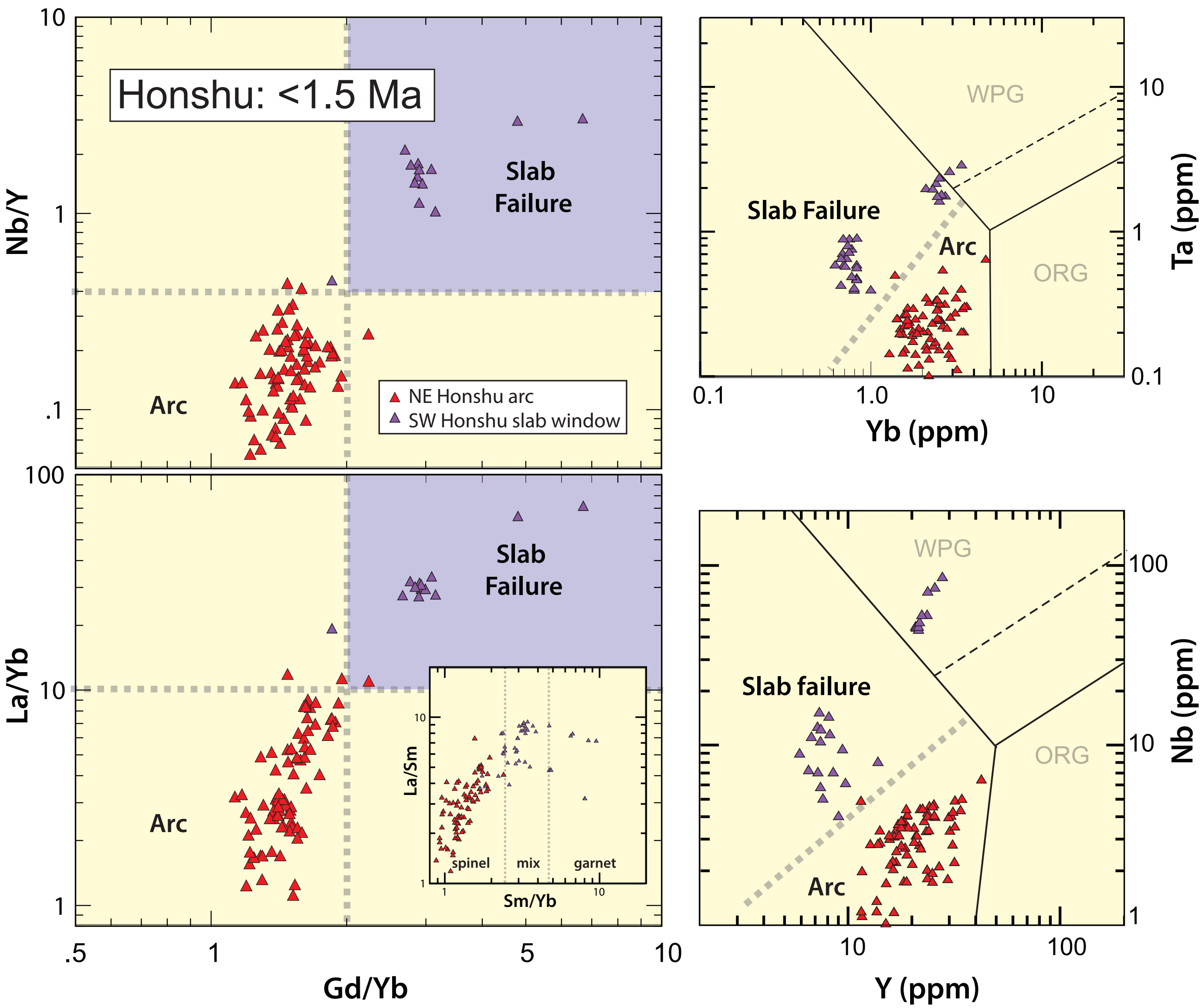


Figure 71. <1.5 Ma slab window and arc volcanic rocks of the Honshu arc, Japan, plotted on our slab failure discrimination diagrams. Data are from Kimura and Yoshida (2006). WPG—within-plate granite; ORG—ocean-ridge granite.

Hildebrand et al., 2014). On a large scale, Archean greenstone belts of the extensively studied Western Superior subprovince are best envisioned as kilometer-scale rafts swimming in a TTG sea.

In our view, the compositional similarities of rocks that form TTG suites (Fig. 12) are so like slab failure (and slab window) suites that a similar origin is warranted. A slab failure model for Archean TTG suites can readily explain both their timing and contact relationships with associated greenstone belts.

Formation and Growth of Continental Crust

Modern estimates of the bulk composition of the continental crust are andesitic (Rudnick and Gao, 2003). Because there previously appeared to be few, if any, realistic alternatives for the origin of continental crust, nearly all models for its creation called upon arc magmatism as the principal mechanism for its creation and growth (Rudnick, 1995; Hawkesworth and Kemp, 2006; Lee et al., 2007; Jagoutz and Schmidt, 2012, 2013; Jagoutz and Kelemen, 2015), although some workers understood there are additional possibilities (e.g., Tarney and Jones, 1994).

Models that require the upper crust to have differentiated from new bulk crust of basaltic composition invariably invoke large-scale lower-crustal delamination because the estimated amount of fractional crystallization necessary to create upper crust is ~86% (Kay and Kay, 1991; Hawkesworth and Kemp, 2006; Jagoutz and Kelemen, 2015; Lee and Anderson, 2015,

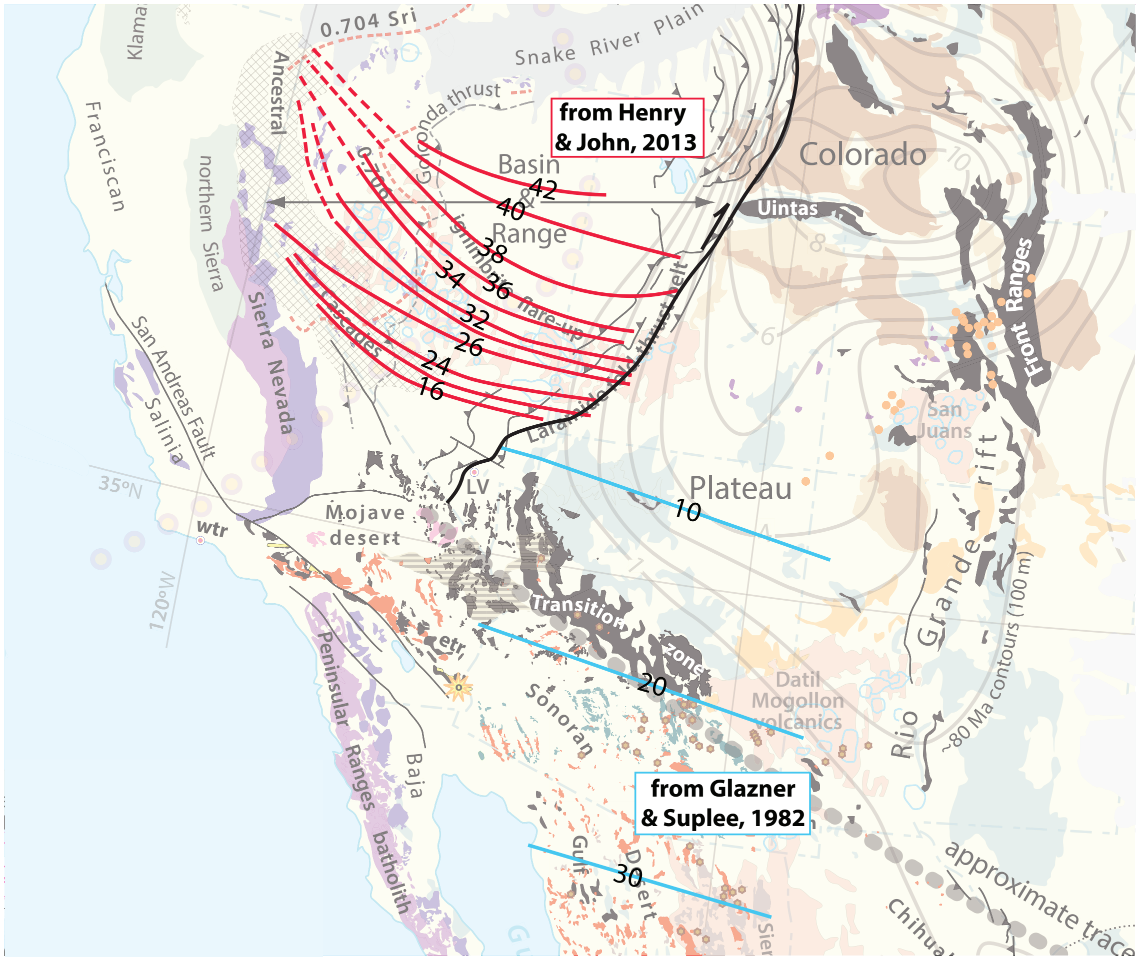


Figure 72. Opposing chrontours (Ma) of Cenozoic magmatism in western and southwestern North America. Southward-migrating magmatism is from Henry and John (2013), whereas northward-migrating magmatism is from Glazner and Supplee (1982). These trends make little sense in terms of easterly subduction beneath North America, but they fit well with a west-northwesterly oriented margin of North America as suggested by Hildebrand (2015). In that scenario, the opposing magmatic belts developed in the upper, North American plate by northward subduction of Pacific plates and reflect a segmented lower plate. See Figure 50 for base map and legend.

and references therein) and, based on cross sections of older arcs such as Kohistan, greater than 2:1 for bulk crust (Jagoutz and Schmidt, 2013). Thus, this process would leave huge volumes of residue in the lower crust, which must have been removed, presumably by gravitational delamination, as such a thick crust is not observed. Also, because the generally accepted model for the bulk of crustal formation is arc magmatism (Rudnick, 1995), the model also implies that the bulk of this differentiation and crustal loss must have happened beneath arcs at convergent margins (e.g., Lee and Anderson, 2015). However, this solution is problematic because the mafic residue must first be converted to eclogite, which requires pressures typically unattainable in arc crust (Green and Ringwood, 1967; Poli, 1993; Hacker, 1996), which, as we have seen, is not typically thick prior to collision. Other variations, such as weathering (Liu and Rudnick, 2011), are also attempts to explain the change in bulk crustal composition from basalt to andesite.

Even a cursory glance at many of our discrimination diagrams (Figs. 12, 76, and 77) would lead to the recognition that the inferred bulk composition of continental crust lies between arc and slab failure rocks. The slab failure magmatism is more siliceous than arc magmatism, and since slab failure magmatism is largely derived from melting of basaltic rocks (MORB, subcontinental lithospheric mantle, basal lithospheric cumulates) that came directly from the mantle, slab failure magmatism is a net contributor to growth of continental crust. Also, because slab failure produces volumes of magma that rival, and probably even exceed, arc magmatism, it may produce larger volumes of continental crust. We believe that, given its siliceous bulk composition (typically 60%–70% SiO_2), invoking slab failure magmatism to

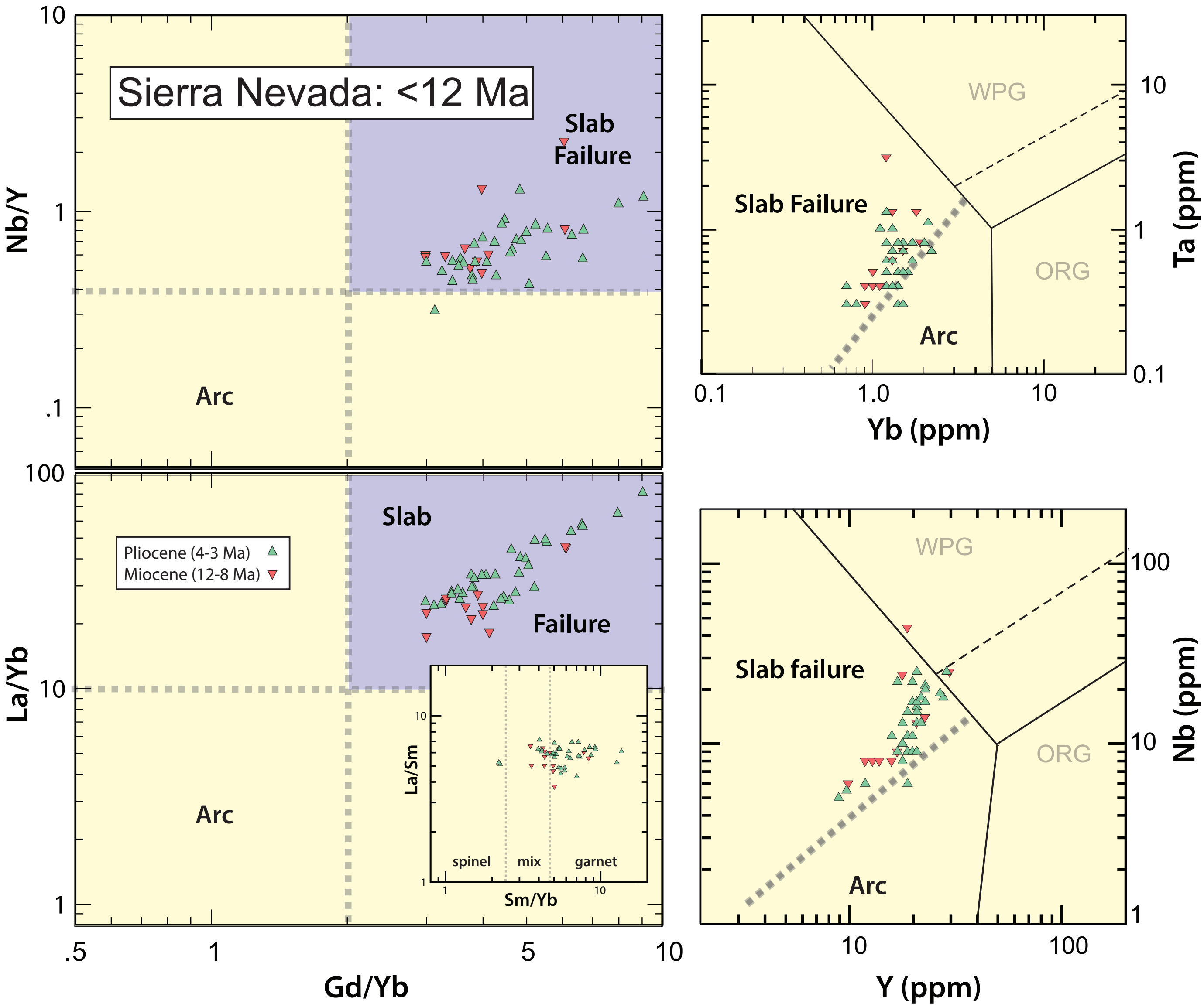


Figure 73. Sierra Nevada slab window mafic volcanic rocks <12 Ma plotted on our slab failure discrimination diagrams. Samples are subdivided into two age groups, 4–3 and 12–8 Ma. Data are from Farmer et al. (2002). WPG—within-plate granite; ORG—ocean-ridge granite.

generate crust may be the "missing ingredient" that precludes the necessity for large amounts of mafic restite to founder and sink into the mantle in order to create the modeled bulk composition of continental crust (Rudnick and Gao, 2003).

Supporting our model for crust formation are two observations: (1) On the basis of their compositions, Precambrian TTG suites (Fig. 76), as described earlier, are best seen as ancient episodic equivalents of slab failure magmatism; and (2) ages of metamorphism and detrital zircons (Fig. 78) suggest that crust formation, at least since the early Archean, was strongly episodic, with major pulses broadly coincident with amalgamation of supercontinents (Cawood et al., 2013; Hawkesworth et al., 2016). If arc magmatism was the dominant mechanism for generation of continental crust—and if it all was not created early in Earth's development—then continental crust should have grown continuously, minus that recycled into the mantle by various mechanisms. While it might be attractive to suggest that the peaks represent preferential preservation (Hawkesworth et al., 2016), arcs are upper-plate phenomena, and during and after collision, it is the upper plate that is preferentially eroded, so it is unlikely that they would be preferentially preserved. If the peaks in U-Pb ages are artifacts of supercontinent formation, then every collision during amalgamation would have experienced slab failure, and the concentrations of U-Pb ages would likely represent zircons derived largely from cooled slab failure magmas. Interestingly, there could well be some preservation bias in the rock record,

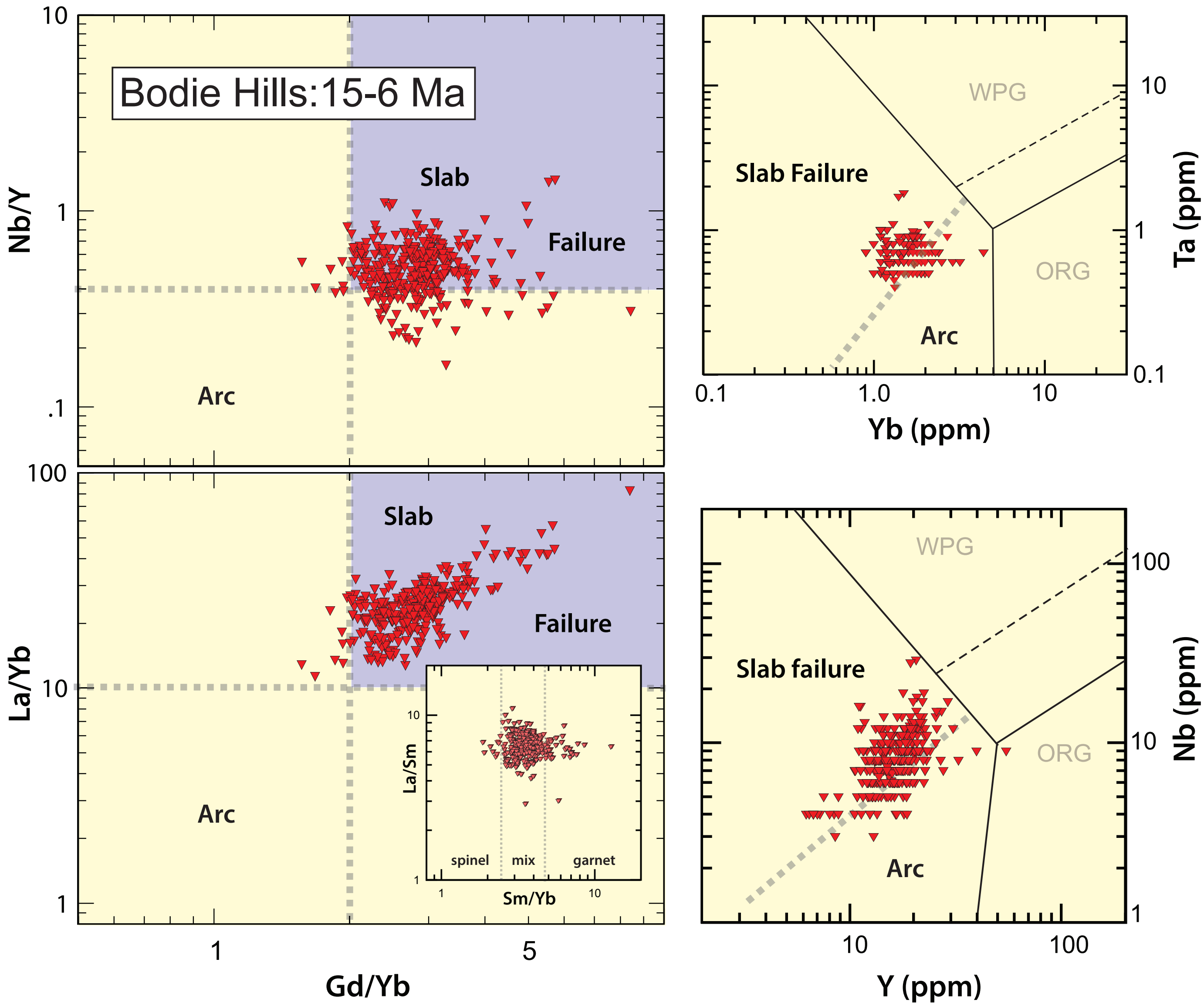


Figure 74. 15–6 Ma volcanic rocks with <70% SiO_2 from the Bodie Hills plotted on our slab failure discrimination, illustrating similarity to slab failure/window magmatism. Data are from John et al. (2012) and du Bray et al. (2016). WPG—within-plate granite; ORG—ocean-ridge granite.

for during collision and slab break-off, large portions of the formerly magmatically active, rift facies of the lower plate could be recycled into the mantle (Hildebrand and Bowring, 1999); thus, older magmatic events and their zircons would be supplanted within the crust by much younger zircons related to slab failure. We conclude that slab failure magmatism best explains the peaks in detrital zircon ages.

Mineralization

Several types of mineralization are generally presumed by researchers to be related to magmatism at convergent margins. Principal among them are the voluminous and economically important Cu-Mo and Cu-Au porphyry deposits, long considered to be formed by arc magmatism (Sillitoe, 1972; Sawkins, 1972). Today, researchers continue to associate porphyry deposits with arc magmatism, although they commonly suggest that they formed from potassic magmas in mature arcs with thick crust (Jamali and Mehrabi, 2015; Wilkinson, 2013; Richards et al., 2012). However, nearly 30 yr ago, Solomon (1990) suggested that many porphyry deposits might have formed from magmatism during collision and subduction reversal, and by the late twentieth and early twenty-first centuries, several workers agreed with the hypothesis and argued that porphyry deposits, especially Cu-Au types, were the product of syncollisional slab failure magmatism (McDowell et al., 1996; de Boorder et al., 1998; Xia et

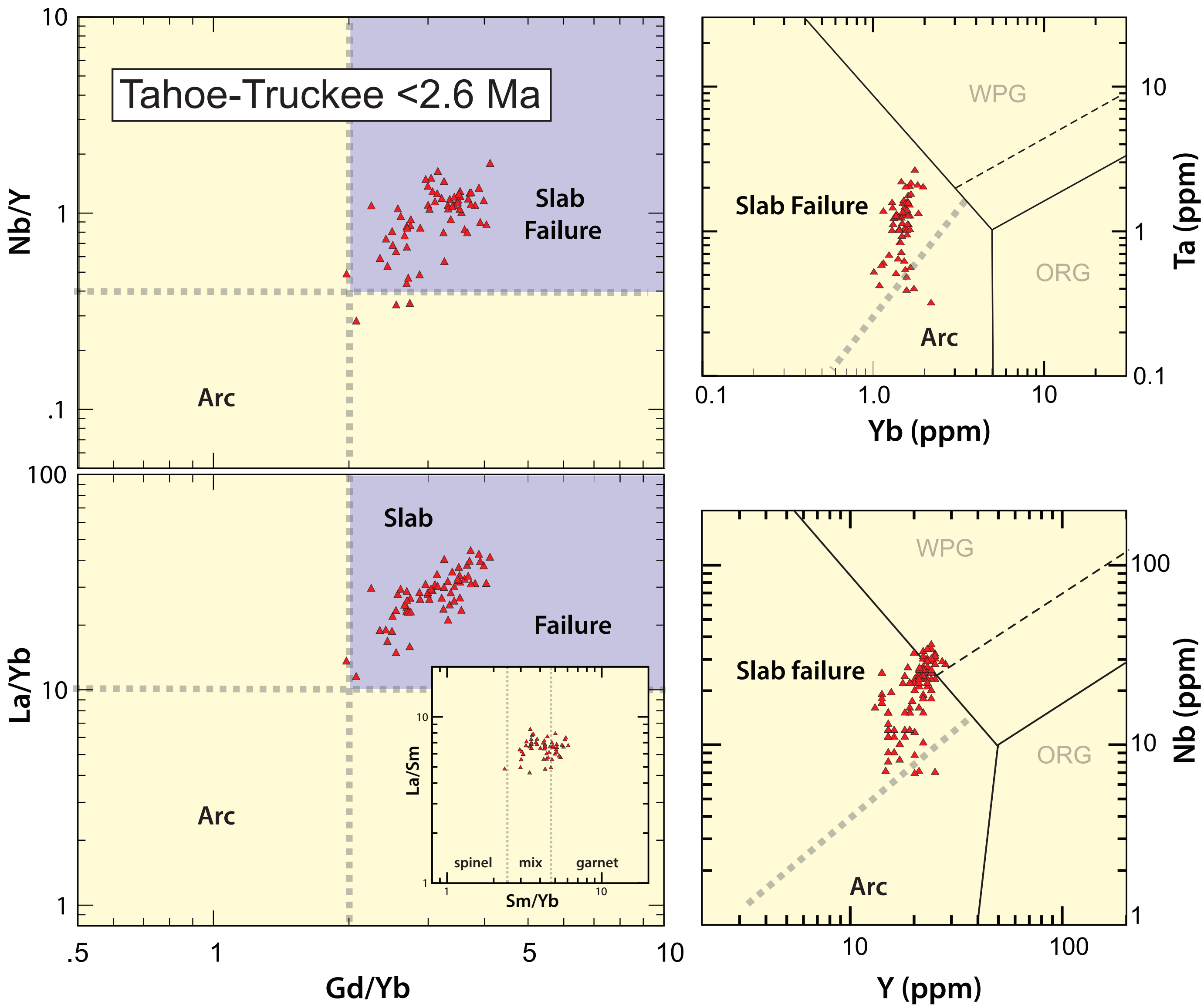


Figure 75. Tahoe-Truckee <2.6 Ma slab window mafic to felsic volcanic rocks plotted on our slab failure discrimination diagrams. Data are from Cousens et al. (2011) and Farmer et al. (2013). WPG—within-plate granite; ORG—ocean-ridge granite.

al., 2003; Cloos et al., 2005; Cloos and Housh, 2008; Hildebrand, 2009; Hou et al., 2015).

Within our study area, large numbers of porphyry deposits occur in the Sonoran segment and, along with their plutonic sources, formed between ca. 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), which suggested to Hildebrand (2013) that they formed from slab failure during the Laramide event. Our geochemical compilation of Laramide plutons (Figs. 51 and 52) supports this hypothesis, not only for the North American Southwest, but also for its prefaulting continuation in the Idaho-Montana region, where the rich deposits near Butte, Montana, are temporally and spatially associated with the Boulder batholith (Houston and Dilles, 2013a, 2013b) and plutons of the surrounding area (Taylor et al., 2007). Other Laramide mineral deposits, including Cu-Au porphyries and vein-type gold deposits associated with slab failure plutons, occur through the Coast plutonic belt northward to the Yukon Territory (Godwin, 1975; Allan et al., 2013) and Alaska (Harlan et al., 2017).

Because the upper plate in collisional belts is commonly deeply eroded, and so arc rocks are commonly underrepresented, we thought it important to examine an area where both arc magmatism and slab failure magmatism are present, mineralized, and have modern geochemical analyses. The Kerman batholith of central Iran satisfies these criteria and so provides a test of the validity of our geochemical discriminators and their exploration

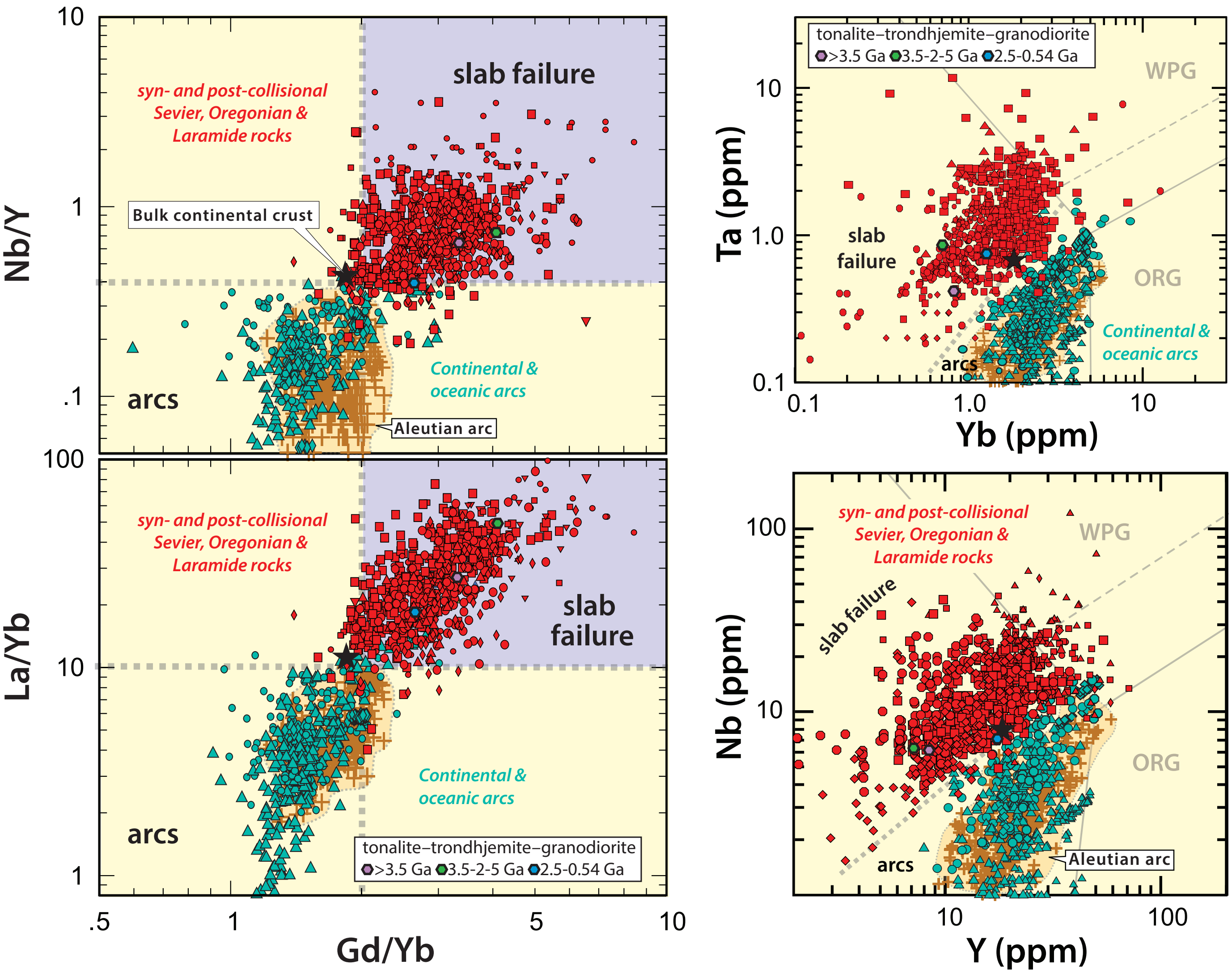


Figure 76. Summary discrimination plots showing the bulk of the trace-element analyses from both arc and slab failure suites (Sevier, Oregonian, and Laramide) used throughout this paper, illustrating the two discrete groups, which we interpret to have different sources and causes. Average values of Precambrian tonalite-trondhjemite-granodiorite (TTG) suites (Martin et al., 2005) are shown to illustrate their similarity to slab failure magmatism. Note that the estimated composition of bulk continental crust (Rudnick and Gao, 2003) plots consistently between arc and slab failure suites, which suggests to us that continental crust was, and is, formed by mixtures of both arc and slab failure magmatism. Aleutians, N = 411; other arcs, N = 750; slab failure, N = 876. WPG—within-plate granite; ORG—ocean-ridge granite.

potential. According to Shafiei et al. (2009), "Pre-collisional Eocene–Oligocene arc diorites, quartz diorites, granodiorites, and volcanic equivalents in the Kerman arc segment in central Iran lack porphyry Cu mineralization and ore deposits, whereas collisional middle-late Miocene adakite-like porphyritic granodiorites without volcanic equivalents host some of the world's largest Cu ore deposits" (p. 265). We plotted their analyses on our standard discrimination diagrams (Fig. 79), where the two groups are readily separated into arc and slab failure magmatism, as our model predicts. Thus, we conclude that our discrimination plots should prove useful in mineral exploration and that slab failure plutons are excellent exploration targets for porphyry mineralization.

Pegmatites, rich in lithium, cesium, and tantalum (LCT pegmatites of Bradley and McCauley, 2013), appear to be concentrated in collisional orogens and provide nearly a third of the world's Li, nearly all of the Ta, and the bulk of the world's supply of Cs (Bradley et al., 2012; McCauley and Bradley, 2014). Oregonian slab failure plutons of the <100 Ma postcollisional slab failure La Posta suite of the Peninsular Ranges batholith have associated 98–93 Ma LCT pegmatites in the Pala district of San Diego County (Foord, 1976; Snee and Foord, 1991). Other examples occur associated with slab failure plutons of the Sevier event within the southern Omineca belt of the southern Canadian Cordillera (Brown, 2003). A factor of possible relevance to the association of slab failure plutons and Ta is the

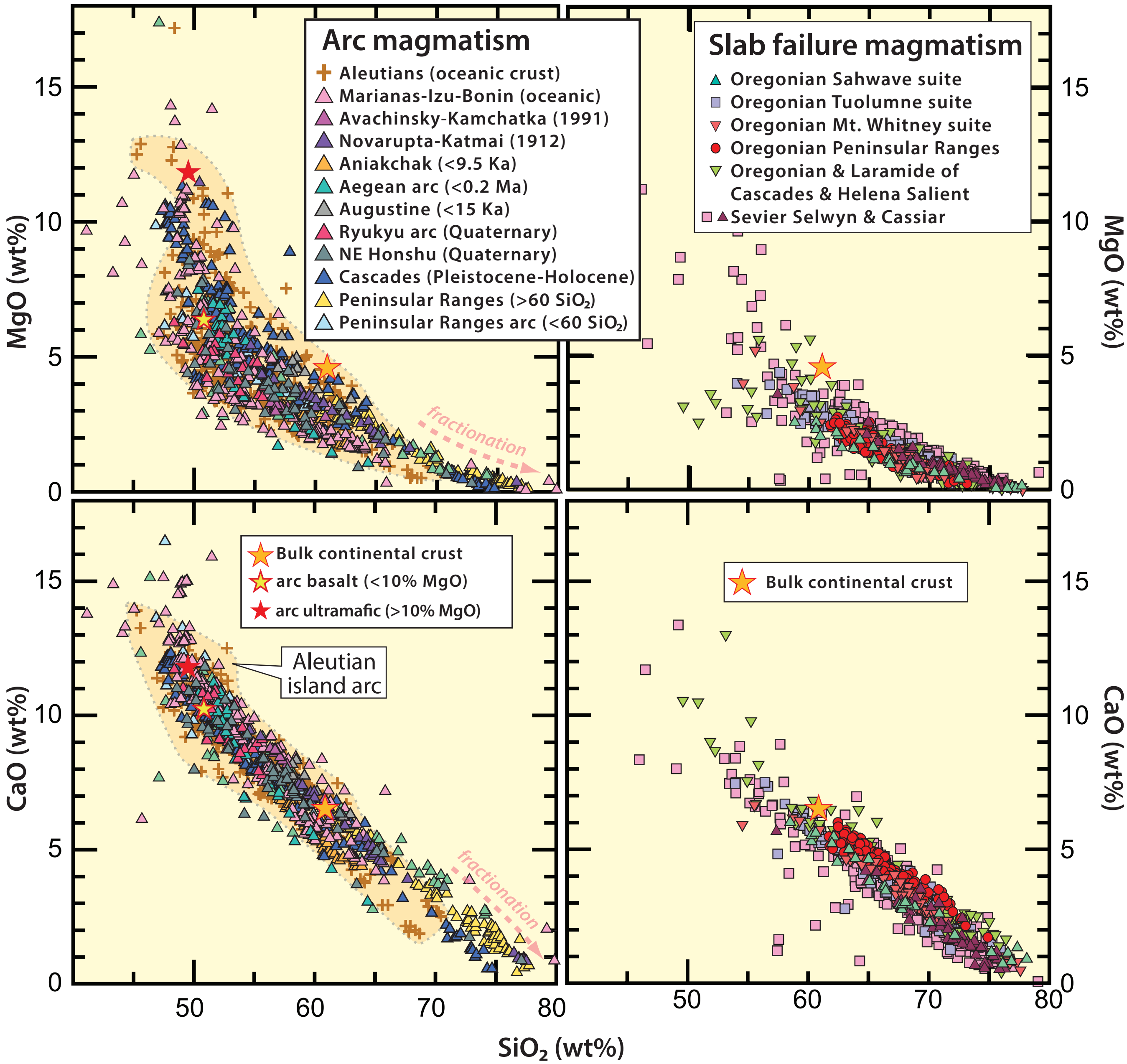


Figure 77. To further test our model for formation and growth of continental crust, we plotted major-oxide contents from samples from both our arc and slab failure reference suites on Harker-type variation diagrams. On these plots, most arc analyses fall above and to the left of estimated bulk continental crust of Rudnick and Gao (2003), whereas most slab failure rocks plot below and to the right, suggesting that continent crust is created by a mixture of arc and slab failure magmatism, as indicated on the previous plots utilizing trace elements and ratios. Note that the estimated MgO content for bulk continental crust is located slightly above and to the right of both arc and slab failure trends, so if our mixing model is correct, then perhaps a better value for bulk crustal MgO content would be about 3 wt%.

general Ta and Nb enrichment in slab failure suites relative to arc suites (Fig. 76), a feature we attribute to a lack of residual rutile and titanite during partial melting.

Slab failure plutons of the Sevier event, now located within the southern Omineca belt and Selwyn Basin of the Canadian Cordillera, are extensively mineralized. Plutons in the Selwyn Basin are part of the Tintina gold belt (Goldfarb et al., 2000, 2007; Hart et al., 2004a, 2004b), and abundant intrusion-related gold deposits are associated with plutons of the Bayonne suite within the southern Omineca belt (Logan, 2002b).

Overall, mesozonal and catazonal Oregonian plutons (100–84 Ma) of the Peninsular Ranges, Sierran, and Idaho batholiths are

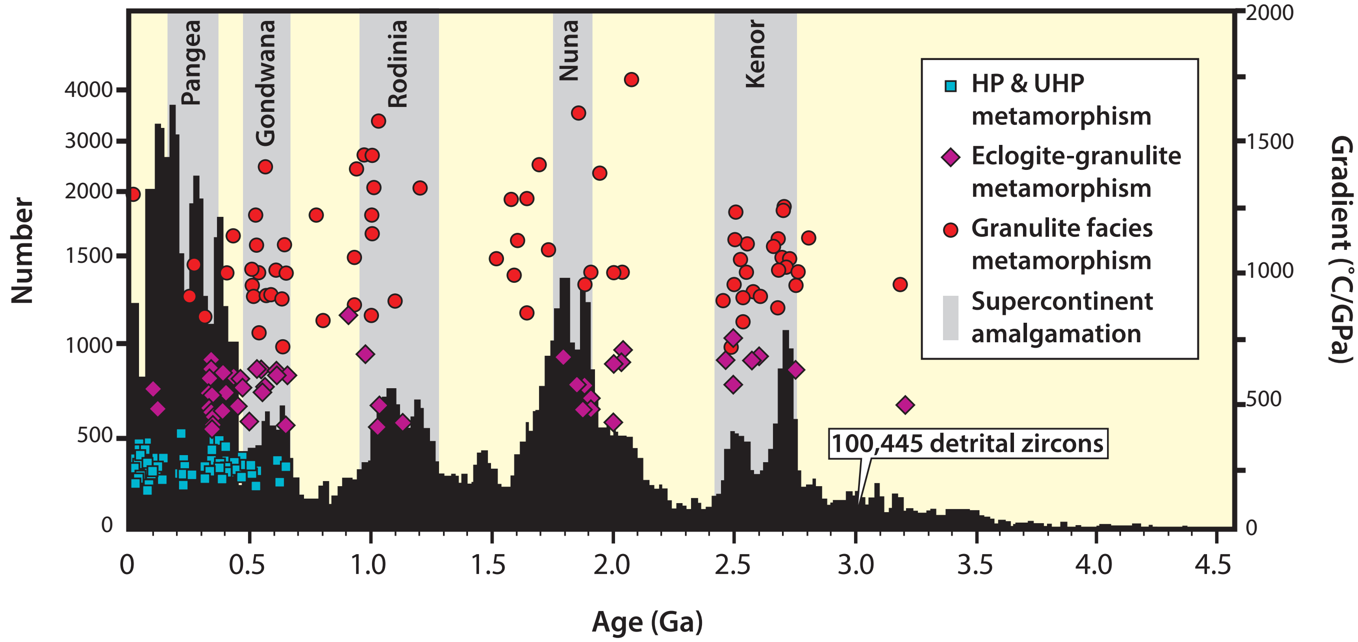


Figure 78. Figure showing the crystallization ages of >100,000 detrital zircon ages coupled with ages of various types of metamorphism through geologic time (modified from Hawkesworth et al., 2016). The origin of the peaks and valleys on this plot is contentious, largely because it is difficult to interpret using crustal growth models, where the bulk of crust is created by arc magmatism, because that is a continuous process and would not lead to the observed peaks and valleys, especially since arcs are the upper plates in collisions and are subject to severe erosion during postcollisional exhumation. However, they are precisely what one would expect if considerable quantities of continental crust are made from slab failure magmatism, as we argue here, for the peaks coincide broadly with periods of supercontinent assembly (labeled in gray fields), which require collisions and resultant slab failure magmatism. We conclude that continental crust is made from both arc and slab failure magmatism. HP—high pressure; UHP—ultrahigh pressure.

poorly mineralized, although there are exceptions (Taylor et al., 2007). Epizonal slab failure plutons of both the Laramide and Sevier events are commonly mineralized, which suggests that the level of emplacement and, probably, meteoric water are important for the development of economic mineralization in slab failure suites.

CONCLUSIONS

(1) Our survey of arcs built on continental crust indicates that they are generally not regions of thick crust. The Andean arc is the exception, but it appears to have been built atop thickened crust formed during the Late Cretaceous–early Cenozoic Laramide event (Hildebrand and Whalen, 2014a). Arc crust gets thickened when the leading edge of a rifted continental margin is subducted beneath the arc (e.g., Froitzheim et al., 2016), in other words, during collisions.

(2) Within what is now the North American Cordillera, we recognize several regional Cretaceous deformational events, which we interpret to represent collisions either within the Cordilleran ribbon continent or between the ribbon and North America. The oldest event discussed here—which followed the Neocomian Nevadan-Brookian collision—is the opening of the Bisbee-Arperos seaway at ca. 135 Ma along the western side of the ribbon continent. The seaway may have extended continuously from at least southern Mexico to the Arctic Ocean. At ca. 122 Ma, prior to closure of the Bisbee-Arperos seaway, a promontory on the eastern side of the ribbon continent collided with cratonic North America to generate the first tectonic thickening of the North American passive margin since its formation in the Early Cambrian. This event was the Sevier orogeny, and it appears to have had a westerly subduction polarity, with deformation confined to the western United States. At 100 Ma, the Bisbee-Arperos seaway closed above a westwardly dipping subduction zone, and the western edge of neoautochthonous North America was partially subducted beneath a 130–100 Ma continental arc complex during what is known as the Oregonian event. Rocks deformed during the event extend from Zihuatenejo to Alaska. The final event is the ca. 80 Ma Laramide orogeny, which represents an outboard collision of another ribbon-like terrane or continent with the Americas. Although the orogen likely extends from Tierra del Fuego to Alaska, here we dealt only with the North American sector. There, during the later parts of the collision, some 1300 km of northward meridional migration, presumably propelled by northward movement of the Kula plate, drove the Sevier hinterland northward from the Great Basin area to the Canadian Cordillera, where it resides today as the Omineca belt. During the latest Paleocene and Eocene, the thickened collisional Laramide hinterland collapsed

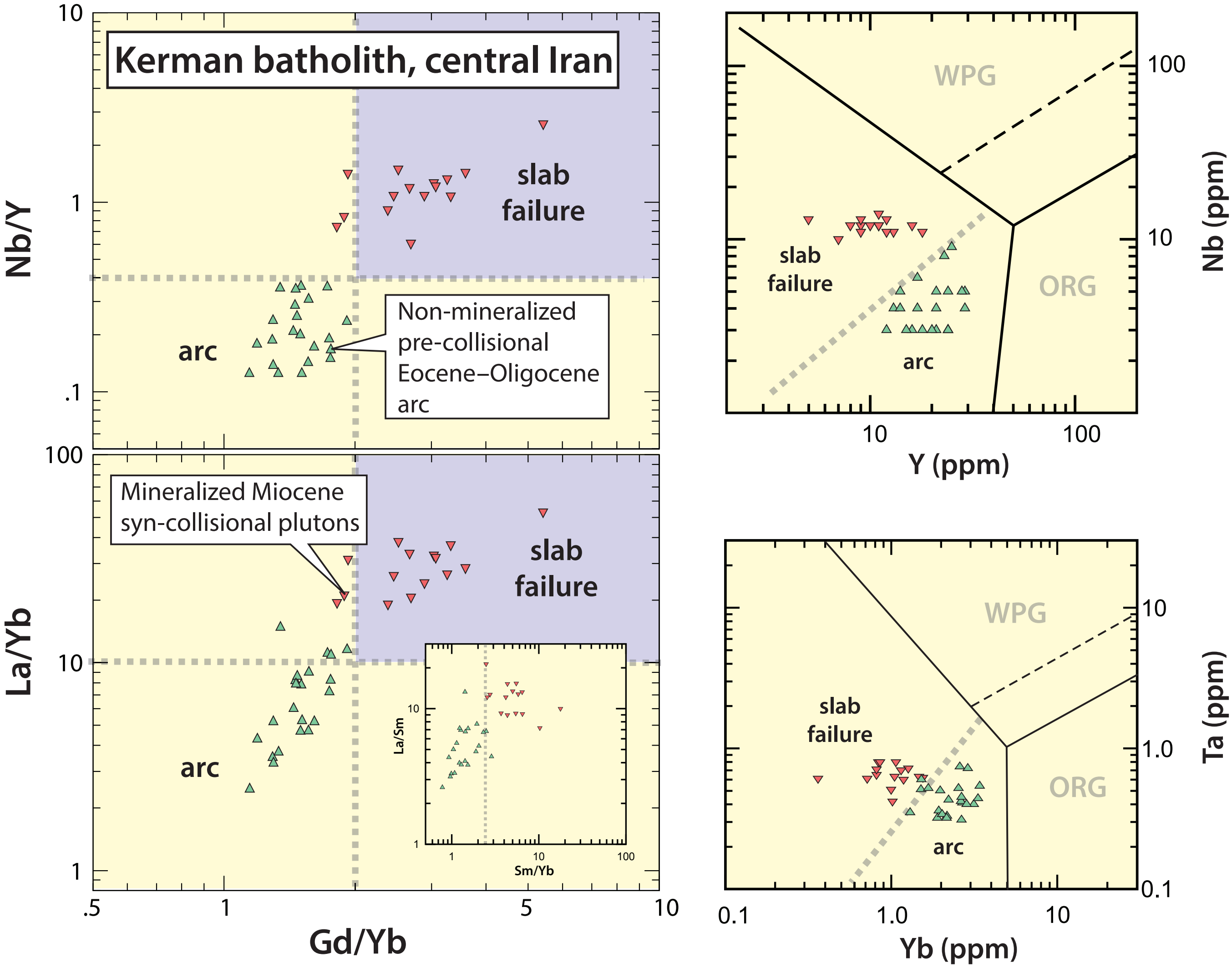


Figure 79. Geochemical analyses from nonmineralized Eocene–Oligocene arc rocks and mineralized syncollisional plutons from the Kerman batholith of central Iran (Shafiei et al., 2009), illustrating the applicability of our various discrimination diagrams for identifying mineralized slab failure plutons. WPG—within-plate granite; ORG—ocean-ridge granite.

gravitationally, and upper-crustal extension was dominated by an extensive network of normal faults.

(3) We examined magmatism during and after each of three Cretaceous compressional events within the Cordillera. Not only did we note their syn- to posttectonic setting, but we also utilized enhanced geochemical criteria developed by Hildebrand and Whalen (2014b) for discriminating arc-related magmatism from slab failure magmatism. The templates for these criteria were constructed empirically from samples of pre- and postcollisional plutonic rocks of the Peninsular Ranges batholith, where we found that various ratios of rare earth elements and high field strength trace elements allowed us to differentiate arc from slab failure magmatism, and which we further verified with geochemical analyses of much younger samples from recognized arc and slab failure settings. We interpret these trace-element ratios to reflect the effects of residual garnet (La/Yb, Gd/Yb, and Sm/Yb), residual garnet without residual plagioclase (Sr/Y, no Eu anomaly), and residual garnet without residual titanite or rutile (Nb/Y and Ta/Yb).

(4) We found that our discrimination diagrams, which were originally based largely on rocks of the Peninsular Ranges batholith ranging in SiO_2 from 60% to 70%, are less reliable for mafic cumulate rocks, as well as plutonic and volcanic rocks containing >70% SiO_2.

(5) Contrary to the batholithic paradigm, in which most major batholiths are generated in arcs, our compilation of existing geochemical and isotopic data from Cretaceous batholiths of the North American Cordillera, combined with the temporal

relationships of magmatism to deformation indicate that the bulk of plutonism was generated after collision by slab failure magmatism, not arc magmatism. These include the well-known La Posta and Sierran Crest magmatic suites of the Peninsular Ranges and Sierran batholiths, the Idaho, Boulder, and related batholiths of the Helena salient, younger than 120 Ma plutons of the Omineca belt and Selwyn Basin within the eastern Canadian Cordillera, and extensive tracts of plutonic rocks within the Coast batholith of British Columbia and its equivalents in Alaska.

(6) Nearly all Oregonian plutons (younger than 100 Ma) in the Sierra Nevada analyzed to date have $\delta^{18}O_{zircon}$ values close to, and overlapping with, mantle values (Lackey et al., 2008), so that significant crustal assimilation and mixing with evolved melts of continental crust are precluded. Their Nd and Sr isotopic values match those of ultramafic xenoliths collected from younger volcanic rocks in the Sierra Nevada and are interpreted to represent old subcontinental mantle lithosphere (Ducea and Saleeby, 1998; Lee et al., 2001a, 2001b). Their Nd and Sr ratios are also the same as young basaltic lavas of the Big Pine volcanic field, Miocene and Pliocene basalts of the Sierra Nevada, and central and eastern basaltic rocks of the Snake River Plain, all of which gained their isotopic signatures from old subcontinental mantle lithosphere. Thus, many, but not all, slab failure rocks have more evolved Nd and Sr ratios than most arc rocks. This suggests to us that the well-known 0.706 Sr isopleth represents the western boundary of slab failure plutons, rather than a feature or division in the subjacent crust, which most workers have interpreted to represent an accretionary boundary. Our interpretation is supported by the observation that in both the Sierran and Peninsular Ranges batholiths, the slab failure and arc plutons overlap in their distribution. Some slab failure suites, such as those of the Coastal batholith of British Columbia, have typical slab failure trace-element signatures, but less-evolved, arc-like Nd and Sr, so they apparently did not interact with old subcontinental mantle lithosphere.

(7) Because it appears that most of the plutons formerly considered to represent the interaction of arc magmatism with "mature" cratonic crust are probably slab failure plutons, we evaluated the extent of crustal assimilation in continental arc rocks by plotting several geochemical attributes generally considered to represent crustal input, such as K_2O, Rb/Sr, and Rb/Zr, from a variety of both continental and oceanic arcs. Although each arc contains a wide variety of compositions, the values for continental and oceanic arcs overlap, which suggests to us that crustal assimilation is not a major process in continental arcs. This does not mean that mantle-derived magmas do not assimilate crust, because crustally derived xenocrysts such as zircons provide evidence for some assimilation. We emphasize that with slab failure magmas out of contention, crustal assimilation is not nearly as important as previously thought.

(8) In our previous contributions, we noted that adakites of arc terranes had similar trace-element profiles to slab failure magmas and, as suggested by others, were generated during ridge subduction and consequent development of slab windows. As the physical process of slab window formation is functionally the same as slab failure in that both involve upwelling of asthenospheric melts through a gap in subducting lithosphere, we looked at the geochemistry of magmas erupted during several well-documented examples of ridge subduction. In terms of our "slab failure elements," the slab window rocks are remarkably similar and suggest that slab window magmatism also involved melting of mantle at greater depths than arc magmatism. The possibility of melting oceanic lithosphere exists in ridge subduction as well as within slab failure.

(9) We noted the episodic nature of Archean and Proterozoic TTG suites, as well as their geochemical similarities to slab failure magmatic suites. This suggests to us that TTG suites represent Precambrian slab failure magmatism.

(10) Our data challenge the generally accepted hypothesis that the bulk of continental crust was created by arc magmatism, because the average composition of bulk continental crust as determined by Rudnick and Gao (2003) plots between arc and slab failure magmatism on most of our geochemical plots. This suggests to us that the majority of continental crust was generated by a combination of arc and slab failure magmatism. In the standard arc-dominated crustal-development model, significant quantities of mafic-ultramafic cumulates are required to be left in the lower crust and later delaminate, because in arcs, the principal magma rising from the mantle is basalt. However, in the slab failure model, significant quantities of magma as siliceous as granodiorite rise from the mantle as a result of slab failure, so there is no need for voluminous ultramafic-mafic cumulate residua. When coupled with the likelihood that Precambrian TTG suites represent slab failure magmatism, it appears to us that slab failure magmatism might have produced an equal, or larger, quantity of continental crust than arc magmatism, although the precise ratios are difficult to estimate due to the uncertainties in resolving the volume of crustal recycling. Slab failure magmatism is an episodic process that occurs during and after collisions, and hence supercontinent formation, so that it might account for the observed major peaks in the detrital zircon record (Hawkesworth et al., 2016).

(11) Wide varieties of economic mineral deposits are associated with slab failure plutons. The most notable are volumetrically large Cu-Au porphyry deposits associated with epizonal plutons; mesozonal and catazonal slab failure bodies typically do not contain economic mineralization. Pegmatites, containing much of the world's supply of lithium, cesium, and tantalum, also appear to be associated with slab failure plutons. These observations suggest that upper-crustal fluids play an important role in alteration and mineralization.

(12) Slab failure is an overlooked phenomenon of plate tectonics, commonly underdescribed, or left out of basic geological models and textbooks, yet it occurs in every collision and is not only responsible for the width of orogens and their exhumation, abundant magmatism and associated mineralization, and recycling of continental crust into the mantle (Hildebrand and Bowring, 1999), but it is also a long active mechanism for generating new continental crust.

ACKNOWLEDGMENTS

This is yet another in our series of self-funded studies on Cordilleran batholiths, so without the kind generosity of so many geologists who shared spreadsheets of their published and unpublished data, this study simply could not have been completed in any reasonable time frame, if at all. We thank Lawford Anderson, Charlie Bacon, Ed du Bray, Scott Bryan, Alan Chapman, Fukun Chen, Mark Cloos, Ron Cole, Weecha Crawford, Rob Creasor, Frank Dudas, Cynthia Dusel-Bacon, Don Francis, Rich Gaschnig, Sally Gibson, James Girardi, Carlos Gonzalez-León, Matt Gorring, Bill Hirt, Wei-Qiang Ji, David John, Peter Kokelaar, Thomas Kruijer, Jade Star Lackey, Cin-Ty Lee, Hao-Yang Lee, Reed Lewis, Val Memeti, George Morris, Doug Morton, Hugh O'Brien, Kirsten Rasmussen, Ioan Seghedi, Erin Shea, John Shervais, Josh Shwartz, Cliff Taylor, Erin Todd, Andy Tulloch, Kuo-Lung Wang, Li-Yun Zhang, and Junhong Zhao for sharing.

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APPENDIX 1

The Belt-Purcell Problem

As stated in the text, an area critical to understanding the overall development of the Cordillera, and one often given as proof that there was no Laramide collision, is the Helena salient of Idaho and Montana (Fig. 28), where rocks were involved in several different events, and so an understanding of the relations is challenging. Furthermore, the traditional view held by most researchers, is that rocks of the Belt Supergroup were deposited upon North America and therefore are not exotic. Recently, other researchers (Johnston, 2008; Hildebrand, 2009, 2013) argued that rocks of the Belt Supergroup formed part of the upper plate beneath which the leading edge of North America was subducted during the Laramide event and are far traveled with respect to their current location. Thus, a more extensive discussion of the area should be beneficial to understanding the nature of the Laramide event. Here, we address some of the complexities in order to integrate rocks of the area into our overall tectonic framework.

Rocks of the Blue Mountains superterrane outcrop in the westernmost regions of the Helena salient and are generally interpreted to represent Upper Paleozoic–Mesozoic terranes consisting of sedimentary shelf deposits, ophiolitic slices and slivers, accretionary prism, magmatic arc rocks, and successor basin fill (Mullen, 1985; Avé Lallement, 1995; Dorsey and LaMaskin, 2008), whereas, based largely on isotopic data, rocks to the east are widely interpreted to represent cratonic lithosphere, presumed to be North American (Armstrong et al., 1977; Fleck and Criss, 1985, 2004; Manduca et al., 1992; Lund et al., 2008; Blake et al., 2009). Metasedimentary rocks of the Neoproterozoic Windermere Supergroup and the Mesoproterozoic Belt Supergroup dominate the inferred North American margin and only locally are seen to sit unconformably upon older Precambrian basement (Doughty and Chamberlain, 2007; Lewis et al., 2007) of uncertain affinity, but probably not part of NW Laurentia (Vervoot et al., 2016).

Most, if not all, of the rocks of the Belt Supergroup and their local basement constitute a huge allochthon (Cook et al., 1987; Yoos et al., 1991; Constenius, 1996; Fuentes et al., 2012) that forms the structurally highest, and thus farthest traveled, thrust sheet within the Cordilleran fold-and-thrust belt (Fig. 28). The allochthon is internally broken by thrust, normal, and strike-slip faults, but it is ~450 km long, continuous for nearly 500 km across strike, and at least 14–16 km thick, and it was transported northeastward over Cretaceous shales of the Laramide foredeep basin (Mudge and Earhart, 1980; Mudge, 1982; Sears, 1988, 2001; Cook and van der Velden, 1995; Burton et al., 1998; Fuentes et al., 2012). Palinspastic reconstructions using balanced cross sections in the frontal fold-and-thrust belt demonstrate that the Belt allochthon was transported eastward a minimum of 100 km (Fuentes et al., 2012) to 200 km (Price and Sears, 2000).

The exposures of microfossil-bearing metasedimentary rocks in the Little Belt Mountains (Fig. 28) led most workers to argue that rocks of the Belt Supergroup were deposited on the western margin of North America. A pinkish to orange quartzite, termed the Neihart quartzite, lies unconformably on Late Archean to Paleoproterozoic meta-igneous rocks (Vogl et al., 2004; Mueller et al., 2002), leading some workers to suggest that the Neihart quartzite is the autochthonous basal unit in the Belt Supergroup (Schieber, 1989). However, the quartzite is nonfossiliferous and is unknown within the allochthon, which makes correlations uncertain.

Price and Sears (2000) provided an analysis of the displacement on thrusts cutting the Belt Supergroup, and, because they considered rocks of the Neihart Formation in the Little Belt Mountains to be autochthonous strata of the Belt Supergroup, they were forced to argue that bulk displacement on the thrusts went from >250 km to the north to <20 in the Helena area and concluded that the displaced rocks to the north rotated clockwise ~30° around the Little Belt Mountains. Given that rocks in the thrust belt just to the south also have total displacements of hundreds of kilometers (Skipp and Hait, 1977; Burton et al., 1998; DeCelles, 2004; DeCelles and Coogan, 2006), an opposite rotation would have to have occurred there, which makes the entire scheme unlikely.

There are potential solutions to the conundrum: (1) The Neihart quartzite and possible rocks of the Belt Supergroup in the Little Belt Mountains are not part of the Belt succession within the allochthon; (2) the Little Belt Mountains are an allochthonous block that includes crystalline basement similar to thrusts just to the south (Skipp, 1987), and it was later dropped down along normal faults after thrusting ceased; and (3) rocks of the Belt Basin were formerly more widespread along the western margin of North America, and rocks of the more westerly allochthon were derived from the North American margin much farther south and then transported northward, where they were juxtaposed against the rocks of the Little Belt Mountains.

Solution three is problematic because rocks of the Belt Supergroup were probably exotic relative to North America (Hildebrand, 2009, 2013), based on: (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 unroofed during the late Cambrian (Link and Thomas, 2009; Link and Janecke, 2009), a peculiar occurrence for the outer part of a miogeocline; (3) the 1.2–1.0 Ga metamorphism and deformation found in Belt metasedimentary rocks (Nesheim et al., 2009; Zirakparvar et al., 2010), which are unknown in cratonic northwestern North America; (4) the detrital zircon suites in the upper half of the Lemhi subbasin, which contain a unimodal population of largely juvenile, 1740–1710 Ma detrital zircons that reflect the presence of a voluminous and unrecognized arc (the Great White arc) to the south (Link et al., 2016); and (5) the unconformably overlying Cambrian Flathead Sandstone, which contains a unimodal 1789 Ma population of detrital zircons that requires an unknown ca. 1790 Ma source (Link et al., 2016). There is possible support for solution two in the Volcano Valley fault (Fig. 28), which is a normal fault that juxtaposes sections containing basal Paleozoic sandstone sitting on 4800 m of sedimentary rocks of the Belt Supergroup against those where the basal Paleozoic sandstone sits directly on Paleoproterozoic crystalline rocks (Reynolds and Brandt, 2005, 2007). Nevertheless, the bulk of data support the concept that the autochthonous Precambrian sedimentary rocks are not Belt Supergroup nor do they correlate with them.

Possibly sitting unconformably atop rocks of the Belt Supergroup is another problematic and poorly understood group of sedimentary rocks known as the Windermere Supergroup. The Neoproterozoic Windermere Supergroup is many kilometers thick, and it includes a wide variety of coarse clastic rocks, in part glaciogenic, along with sparse volcanic rocks, shales, arkosic grits, and minor carbonate (Ross, 1991). Some workers (Stewart, 1972; Burchfiel and Davis, 1975; Lund, 2008) have 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 do not contain thick sections of evaporites nor extensive tracts of volcanic rocks, they are dissimilar to younger rift-related sequences. Additionally, they are as many as 200 m.y. older (Lund et al., 2003; Fanning and Link, 2004; Macdonald et al., 2010; Keeley et al., 2013) than the development of the passive margin, so they would not have retained enough heat to match the rate of early Paleozoic subsidence (Bond and Kominz, 1984; Devlin and Bond, 1988).

Whatever the ultimate origin of rocks contained within the Belt-Purcell Supergroups, it is clear that they are far traveled and only arrived at their present location during the Late Cretaceous Laramide event. Field mapping and U-Pb dating along the Salmon River suture indicate that rocks of the Blue Mountains superterrane (Figs. 28 and 50) and those of the Belt-Windermere package were juxtaposed prior to 110 ± 5 Ma, as indicated by the age of and isotopic data from the intrusive Little Goose complex, which cuts the contact (Manduca et al., 1992, 1993; Giorgis et al., 2008). Subsequent deformation, interpreted as transpressional deformation within the Idaho shear zone, is older than 90 Ma, based on the U-Pb ages of granitic pegmatites that cut the fabric (Giorgis et al., 2008). K/Ar and $^{40}Ar/^{39}Ar$ ages in the area fall in the range 93–85 Ma and partially overlap with the emplacement of the 98–87 Ma metaluminous plutons of the Idaho batholith (Lund and Snee, 1988; Manduca et al., 1993; Snee et al., 1995; Giorgis et al., 2008), suggesting that the plutons were emplaced during Oregonian exhumation.

Located just east of the older suture and shear zone is the Cretaceous Coolwater culmination, an oval tectonic window (Fig. 28) where rocks beneath the basal thrust contact of the Belt-Purcell allochthon are exposed. Detailed studies by Lund et al. (2008) showed that metasedimentary rocks beneath the thrust were deposited after 98 Ma and prior to intrusion of an orthogneiss dated at 86 Ma. Given that the rocks of the Belt-Purcell allochthonous block are continuous for over 400 km to the east, where they also sit structurally upon Cretaceous rocks, it seems reasonable that the entire allochthon, including local slivers of basement, was derived from west of the Coolwater culmination and transported over 400 km to the northeast after 86 Ma, presumably during the <82 Ma Laramide event. This interpretation differs from that of Lund et al. (2008), who suggested that rocks within the Coolwater culmination were underthrust from the west, but as deformation within the Salmon River corridor had ceased by 90 Ma, that model is untenable.

Not only was there large-magnitude E-NE transport of the Belt-Purcell–Blue Mountains superterrane during the Laramide, but as discussed earlier, after 70 Ma, there was also ~1300 km of northward migration as well. Hildebrand (2015) pointed out that robust paleomagnetic data from the Blue Mountains terranes indicate 1760 ± 460 km of northerly movement after ca. 90 Ma (Housen and Dorsey, 2005), which also must apply to rocks of the Belt-Windermere block, as they were one block at that time.

In his 2015 contribution, Hildebrand used the Lewis and Clark line and Texas lineament as piercing points to argue that the southern margin of North America was oriented just north of west (~300°) after the Callovian breakup of the central Atlantic until the Late Cretaceous, when it was overridden obliquely by rocks of the Cordilleran ribbon continent. On the basis of a linear band of abundant faults, folds, and intense cleavage (Wallace et al., 1990; Sears, 1988) that affects the Belt allochthon along strike of the Lewis and Clark Line, he suggested that rocks of the Belt Supergroup were thrust over the Lewis and Clark zone, which, prior to northward migration of the Cordillera, marked the approximate southern margin of North America. As the Belt allochthon extends for ~250 km north of the Lewis and Clark line, it is likely that other rocks of the Cordilleran ribbon continent were thrust similar distances to the north, and, indeed, a huge zone of Eocene exhumation extends west of the Belt allochthon for at least 200 km north of the Lewis and Clark line (Fig. 31). Within this region, termed here the southeastern British Columbia extended terrane, high-grade metamorphic rocks in the Valhalla complex reached granulite facies with peak pressures of 8 ± 1 kbar during the Laramide event at 72–67 Ma (Spear and Parrish, 1996). Also, along the western margin, rocks beneath the Okanagan Valley fault reached pressures of 6 kbar during Late Cretaceous compressional deformation (Bardoux, 1993). In the Grand Forks complex, peak metamorphic pressures reached 5.6 kbar at 59–50 Ma prior to extensional collapse (Cubley and Pattison, 2012). These metamorphic pressures, plus fault reconstructions, suggest that, in places, the crust of the area was effectively doubled and was as thick as 50 km following Laramide deformation (Johnson and Brown, 1996). Similar pressures existed in the North Cascades, where the Chelan complex yielded pressures of 10 kbar (Dessimoz et al., 2012), and the Skagit gneiss complex of the High Cascades reached pressures >9–10 kbar (Gordon, 2009; Miller et al., 2009). Just to the north of the Lewis and Clark line in Washington, there is the Priest River complex, where 2.65 Ga Archean quartzofeldspathic gneiss is overlain by quartzite and migmatitic pelitic paragneiss, which in turn are structurally overlain by 1577–1511 Ma augen gneiss (Doughty et al., 1998; Doughty and Price, 1999; Doughty and Chamberlain, 2008). These authors reported metamorphic monazite and xenotime ages of ca. 72 Ma, which date peak metamorphic conditions. The 1577 Ma age for the gneisses above the décollement is unknown on the western North America craton (Hoffman, 1989), and therefore it likely represents exotic crystalline basement now sitting atop rocks aggregated with the North American craton and its thin veneer of autochthonous cover. All in all, the data suggest that the Laramide suture extends from the Belt allochthon more or less westerly across southern British Columbia before turning northward along the eastern side of the exhumed Coastal batholith (Fig. 31), and that Eocene extensional collapse in southern British Columbia reflects regions of doubled crustal thicknesses formed by attempted subduction of what was, at the time, the leading edge of the North American margin during the Laramide event.

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