

# Basin and Range volcanism as a passive response to extensional tectonics

Keith Putirka\* and Bryant Platt\*

Department of Earth and Environmental Sciences, California State University, Fresno, 2576 E. San Ramon Avenue, MS/ST24, Fresno, California 93740-8039, USA

## ABSTRACT

A long-standing issue in Cordilleran geology involves the nature of Basin and Range (western USA) volcanism, and whether such magmatism provides a trigger for extensional deformation, or if volcanic activity is a passive response to extension. We use space-time-composition patterns across the central and southern Basin and Range, and where appropriate, reconstructed latitudes and longitudes of volcanic rocks, to show that volcanism is fundamentally a passive process. Our analysis suggests that Basin and Range volcanism is initiated by the transition from a subduction to a transform boundary (now manifest as the San Andreas fault), which causes a slab window to open, as the subducting Farallon plate falls away. Accordingly, volcanic activity follows the northward-migrating Mendocino Triple Junction (MTJ). In the wake of the MTJ, continental mantle lithosphere is heated over a time scale of 10–12 Ma; it then rapidly degrades (or is removed) 17–20 Ma after MTJ arrival at any given latitude, and is replaced by asthenosphere. In the central Basin and Range, MTJ migration triggers the well-documented structural migration of the Sierra Nevada away from the Colorado Plateau. Not only is volcanism triggered by the tectonic transition, but in the central Basin and Range volcanism also migrates west, following the initiation of upper crustal extensional faulting; the lag between the onset of extension and initiation of volcanism is a remarkably consistent 2 Ma. Within this framework, the initial stage of eastward-concentrated volcanism is dominated by felsic magmatism, which then gradually degrades to a more mafic composition. This pattern, and temporal relations indicate that once lithospheric extension begins, between 2 and 5 Ma are needed to develop conduits through which nonviscous magmas can transit the

crust, and that especially high amounts of extensional strain favor the eruption of  $K_2O$ -rich, low-degree partial melts. Our observations indicate that it is unlikely that mantle plume processes initiated Basin and Range volcanism within the study area, and that decreases in  $SiO_2$  and increases in incompatible element abundances to the east within the Cordillera are best explained by continental mantle lithosphere that thickens to the east, which reduces average melt fractions.

## INTRODUCTION

The interplay between magmatism and tectonics in the western United States (Fig. 1) is a topic of long-standing debate; proposed models for Basin and Range magmatism include thermal mantle plumes (Best and Brimhall, 1974; Parsons et al., 1994; Wang et al., 2002), continental extension (e.g., Gans et al., 1989; Christiansen et al., 1992; Wernicke et al., 1996), migration of the Mendocino Triple Junction (Dickinson and Snyder, 1979; Glazner and Supplee, 1982), lithosphere removal (e.g., Feldstein and Lange, 1999; Humphreys, 1995, 2009), or some combination thereof (Ormerod et al., 1988). Some have even suggested that the Basin and Range is a region in which plate tectonic theory fails to adequately explain volcanic activity (e.g., Cañon-Tapia and Walker, 2004). Whether mantle processes play an active role in promoting volcanism (e.g., a mantle plume; Wang et al., 2002) or magmatism is a passive response to tectonism and crustal extension (McQuarrie and Oskin, 2010) is a fundamental issue of Cordilleran evolution. To test whether plate tectonic actions control volcanism, we add a geochemical perspective to structural and palinspastic reconstructions of Snow and Wernicke (2000), McQuarrie and Wernicke (2005), and McQuarrie and Oskin (2010), and test the models in Putirka et al. (2012) using a broader context. We do this by comparing temporal distributions and chemical compositions of volcanic rocks to time-space patterns of extensional deformation (Snow and Wernicke, 2000;

McQuarrie and Wernicke, 2005) and an evolving plate boundary (Atwater and Stock, 1998).

Our observations indicate that the onset of Basin and Range volcanism is controlled by the arrival of the Mendocino Triple Junction (MTJ; Fig. 2) at any given latitude and that volcanic rock compositions change dramatically in the 5–20 Ma after MTJ arrival. We also show that volcanism initiation and volcanic rock compositions are sensitive to the onset of extensional strain. These results indicate that Basin and Range volcanism is a passive response to external tectonic forces.

## Methods

Geochemical and geochronologic data from the GEOROC (Geochemistry of Rocks of the Oceans and Continents; [georoc.mpch-mainz.gwdg.de](http://georoc.mpch-mainz.gwdg.de)) and NAVDAT (North American Volcanic and Intrusive Rock Database; [www.navdat.org](http://www.navdat.org)) databases were used to reconstruct the eruptive locations of volcanic centers. To determine the extrusive location of the volcanic rocks, we employ the paleogeographic reconstruction of the central Basin and Range province presented in Figure 13 of Snow and Wernicke (2000), hereafter referred to as SW00, which incorporates strain markers from field studies. The SW00 reconstruction provides two grids, one depicting the modern geography and the other the reconstructed geography, which we employ to infer the azimuth and amount of extension ( $D_{tot}$ ; tot is total) at any locality. The average extension rate at a given point ( $D_{tot}/\Delta t$ ) can then be determined once the time of extension initiation,  $t_i$ , is determined ( $\Delta t = t_i - t_{present}$ ). Glazner and Bartley (1984) were perhaps the first to show that extensional tectonism throughout the southern Basin and Range migrates northward with time, with an apparent relationship to volcanic activity (Glazner and Supplee, 1982; see also review by Sonder and Jones, 1999). Subsequent field studies confirm that view (see discussion herein, Volcanism as Controlled by Crustal Extension). For example, extension began at 27 Ma in southeast California (33°N;

\*Emails: Putirka: kputirka@csufresno.edu; Platt: line58@mail.fresnostate.edu.

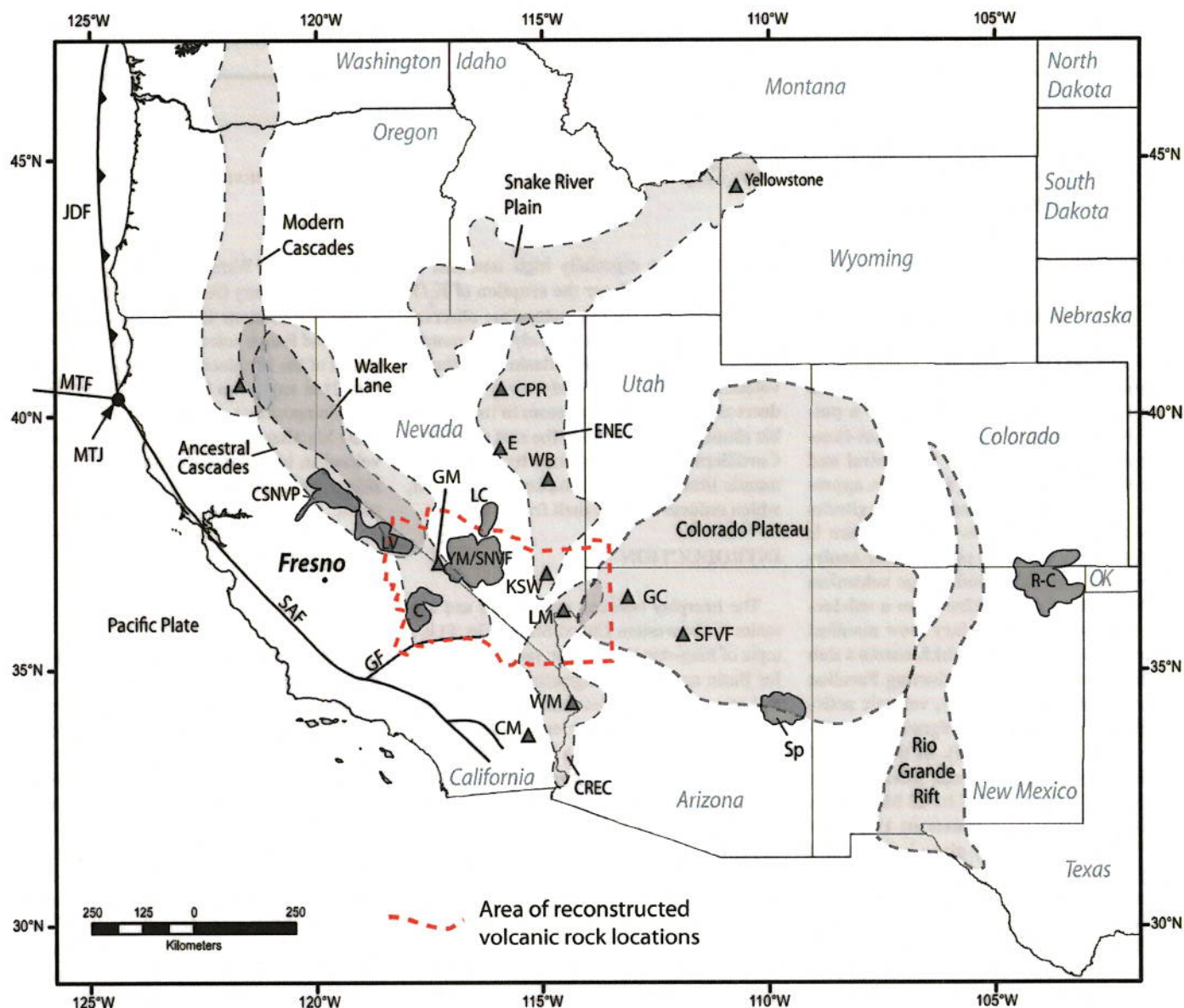


Figure 1. Map showing volcanic provinces and the locations of features discussed in the text. The dashed red line encompasses that part of the central Basin and Range for which we reconstruct paleolatitudes and paleolongitudes for volcanic rocks, using Snow and Wernicke (2000). C—Coso Volcanic Field; CREC—Colorado River Extensional Corridor (Howard and John, 1987); CM—Chocolate Mountains; CPR—Carlin-Piñon Range; CSNVP—Central Sierra Nevada volcanic province; E—Eureka, Nevada; ENEC—Eastern Nevada extensional corridor; GC—Grand Canyon volcanics (Uinkaret volcanic field); GF—Garlock fault; GM—Grapevine Mountains; JDF—Juan de Fuca plate; KSW—Kane Springs Wash; L—Lassen volcanic field; LC—Lunar Crater; LM—Lake Meade; LV—Long Valley; MTF—Mendocino Transform fault; MTJ—Mendocino Triple Junction; R-C—Raton-Clayton volcanic field; SAF—San Andreas fault; SFVF—San Francisco volcanic field; Sp—Springerville Volcanic Field; WB—Window Butte; WGB—western Great Basin; WM—Whipple Mountains; YM/SNVF—Yucca Mountain–Southern Nevada volcanic field.

Jacobsen et al., 2002), at 17–18 Ma near Lake Meade (36°N; Duebendorfer and Simpson, 1994), at 15–16 Ma at Death Valley (36.5°N; Wernicke et al., 1988; Miller and Pavlis, 2005), and at 9.8 Ma at Sonora Pass (38.4°N; Busby et al., 2008; Busby and Putirka, 2009). We regress these points to obtain the equation:

$$t_i = 132.5 - 3.2Y, \quad (1)$$

where  $t_i$  is in years and  $Y$  is degrees latitude; the correlation coefficient for Equation 1 is 0.9996. Equation 1 allows us to interpolate between known initiation points, and to define an extension rate for each grid in SW00. As extension

continues to the present ( $\Delta t = t_i - t_{\text{present}} = t_i - 0 = t_i$ ), the extension rate at any point on the grid is  $D_{\text{tot}}/t_i$ . The extension ( $D_{\text{volc}}$ ; volc is volcanic) undergone by a rock of a given age ( $R_A$ ) is thus  $R_A(D_{\text{tot}}/t_i)$  when  $R_A < t_i$ , or  $D_{\text{volc}} = t_i(D_{\text{tot}}/t_i) = D_{\text{tot}}$  when  $R_A \geq t_i$ . Although we use Equation 1 for interpolation purposes only, an interesting

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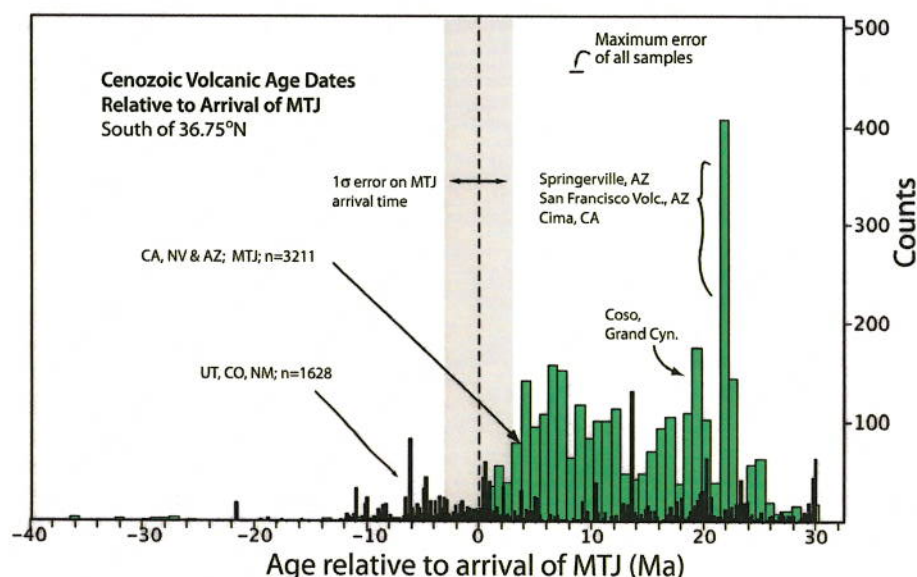


**Figure 2.** Histogram of volcanic rock ages relative to the arrival of the Mendocino Triple Junction (MTJ), calculated using Equation 3, and Equation 1 of Putirka et al. (2012). Practically all volcanic activity south of 37°N, in California (CA), Nevada (NV), and Arizona (AZ), occurs after the MTJ passes a given latitude. In contrast, volcanism in the eastern Cordillera (Colorado [CO], New Mexico [NM], Utah [UT]) shows much less, if any temporal connection to MTJ arrival. In the CA-NV-AZ region, volcanic activity may steadily decline over a 20–30 Ma period; activity at Coso, Cima, and the San Francisco and Springerville volcanic fields is almost certainly overrepresented due to high sampling rates at those fields. Cyn—canyon.

test involves transtensional deformation at the Lassen volcanic center, which began ca. 3.5 Ma (Muffler et al., 2008). The southern terminus of the Clynne and Muffler (2010) map of the Lassen National Park and vicinity is 40.33°N, which, when entered into Equation 1, yields an initiation time of 3.4 Ma (and 2.9 Ma for the latitude of Lassen Peak, at 40.5°N), nearly identical to the field-based estimate of Muffler et al. (2008).

Because air-fall and ash-flow tuffs can be far traveled, we have attempted to filter out tuffs that do not form a spatial cluster, for example, peripheral to a volcano or caldera system. We also applied a geochemical filter to identify volcanic rocks affected by K-metasomatism; such altered rocks generally have  $K_2O/Na_2O > 5$  (Brooks and Snee, 1996); we exclude all rocks with  $K_2O/Na_2O > 3$  (using an even more stringent filter of 2.5 or 2 does not affect our results).

In Putirka et al. (2012), it was shown that  $K_2O$  contents are a key discriminator between Cascade and Basin and Range magmas; the latter group tends to have higher  $K_2O$  contents at any given  $SiO_2$  content. Furthermore, in Putirka and Busby (2007) it was suggested that high  $K_2O$  contents may signal high crustal extension rates. To test these ideas, we applied the equation

(Putirka et al., 2012) that defines the boundary between Cascade and Basin and Range compositions, the K-index (Fig. 3):

$$K\text{-index} = K_2O - 0.12[SiO_2] + 5.1. \quad (2)$$

To identify whether volcanic rocks have a mantle lithosphere or asthenosphere source, we rely on prior geochemical studies of Basin and Range basalts. DePaolo and Daley (2000), building upon work by Fitton et al. (1988, 1991) and Kempton et al. (1991), showed that asthenosphere and subcontinental mantle lithosphere sources can be differentiated using  $^{143}Nd/^{144}Nd$  and  $^{87}Sr/^{86}Sr$  ratios and the ratios of large ion lithophile elements (LILEs) to high field strength elements (HFSEs) (i.e., La/Nb, Ba/Nb, or Sr/P). LILEs are enriched in subduction-related fluids, whereas HFSEs are depleted, leading to high LILE/HFSE ratios (e.g., high La/Nb, Ba/Nb) at arcs; ancient, subcontinental mantle lithosphere contains similar enrichments, and reveals its time-integrated enrichments in the form of low  $^{143}Nd/^{144}Nd$  and high  $^{87}Sr/^{86}Sr$ . For example, asthenospheric partial melts have low La/Nb (<1.7) and high  $^{143}Nd/^{144}Nd$  (>0.513), whereas Cordilleran mantle lithospheric partial melts range to high La/Nb (>7) and low  $^{143}Nd/^{144}Nd$  (<0.512) (Perry

et al., 1987; Fitton et al., 1988, 1991; DePaolo and Daley, 2000) due to Mesozoic-aged subduction-related enrichments (Humphreys et al., 2003; Lee, 2005; Dickinson, 2006). Provided that volcanic rocks avoid crustal contamination, La/Nb, Ba/Nb and  $^{143}Nd/^{144}Nd$  ratios indicate the nature of the source region because these ratios are unaffected by fractional crystallization or degree of partial melting. Estimated asthenosphere and various continental mantle lithosphere source compositions are adapted from Putirka et al. (2012), and the composition of sub-Cordilleran asthenosphere is approximated using mid-oceanic ridge basalts from the East Pacific Rise (Perfit et al., 1996; Hays, 2004; Sims et al., 1999).

## RESULTS AND DISCUSSION

### Reconstructed Volcanic Rock Locations

The SW00 model yields considerable displacements with respect to longitude but remarkably small latitudinal displacements. For example, latitudinal displacements are as great as 110 km for a few Miocene samples in the Walker Lane, but 90% of latitudinal displacements ( $n = 1956$ ) are <25 km, and the mean latitudinal offset is just 13.7 km. These small displacements are a function of principally east-west extension outside the Walker Lane, while within the Walker Lane most volcanic rocks (e.g., Coso, Big Pine, Long Valley) are too young to have been displaced more than a few kilometers. In contrast, longitudinal displacements are considerable, averaging 79.4 km across the reconstructed area and ranging to 280 km for Miocene rocks of the Sierra Nevada. Our maximum displacement (280 km) and mean extension rates (23 mm/yr maximum;  $10.6 \pm 4.3$  mm/yr average) both are within ranges determined by Wernicke et al. (1988), and our processing reproduces fixed points (cities) in SW00 to within  $\pm 5$  km. Consequently, we restrict our analyses of longitudinal variation to the central Basin and Range, where we have reconstructed paleopositions, but we consider a broader array of Basin and Range data when only evaluating latitudinal positions.

### Migrating Mendocino Triple Junction and Onset of Volcanism

Dickinson and Snyder (1979) and Glazner and Supplee (1982) proposed that volcanic activity may be linked to development of the Mendocino Triple Junction (MTJ) and a developing slab window (Atwater and Stock, 1998). In Putirka et al. (2012), it was determined that volcanism in the Sierra Nevada and Walker Lane closely followed

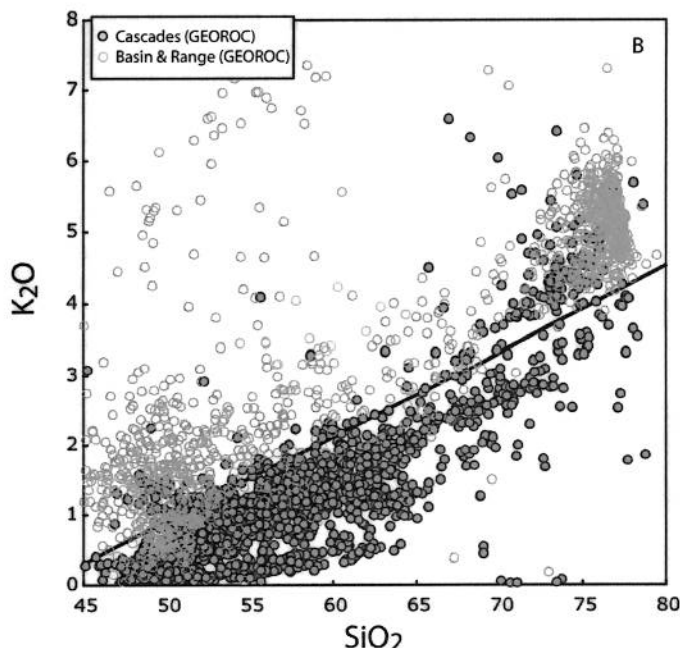


**Figure 3.** A comparison of  $K_2O$  versus  $SiO_2$  for volcanic rocks of the Basin and Range and Cascade provinces. In Putirka et al. (2012), it was shown that in California (CA), Basin and Range (or Walker Lane) rocks have higher  $K_2O$  than the Cascade volcanics. The boundary of Putirka et al. (2012, caption to Fig. 2 therein), calibrated for CA volcanics, successfully separates Cascade and Basin and Range suites as classified by GEOROC (Geochemistry of Rocks of the Oceans and Continents; [georoc.mpch-mainz.gwdg.de](http://georoc.mpch-mainz.gwdg.de)) at the 90% level. Higher  $K_2O$  is likely a result of smaller degrees of partial melting beneath thicker lithosphere in the interior of the Cordillera.

MTJ arrival. We test this issue further by comparing volcanic activity across the southern and central Basin and Range. We calculate the time before and/or after MTJ arrival,  $t^{B/AMTJ}$ , as:

$$t^{(B/A)MTJ} = t^{MTJ \text{ arrival}} - t^{\text{eruption}}, \quad (3)$$

where the term  $t^{MTJ \text{ arrival}}$  is the time that the MTJ arrives at the latitude at which a given sample was erupted (Putirka et al., 2012), and  $t^{\text{eruption}}$  is the volcanic age from NAVDAT. When  $t^{B/AMTJ} < 0$ , the prevailing tectonic setting is subduction;  $t^{B/AMTJ} > 0$  signifies transform motion at the plate boundary. We restrict our analysis to rocks erupted south of  $36.75^\circ N$ , because north of this latitude volcanism is affected by the south- and west-migrating northern Basin and Range arc front (Cousens et al., 2012). Because latitudinal corrections for Basin and Range extension are negligible, we plot present-day latitudes for all samples.

Histograms of age dates (Fig. 2) show that MTJ arrival is well correlated with the timing of volcanic activity within 600 km of the

plate boundary (i.e., within California [CA], Nevada [NV], and Arizona [AZ]), and is more poorly correlated with volcanism farther to the east (Utah [UT], Colorado [CO], New Mexico [NM]). In the CA-NV-AZ region, volcanic activity is nearly absent prior to MTJ arrival, then increases significantly, reaching a peak at 6 Ma post-MTJ arrival. While no waning trend is well defined, volcanic activity appears to have died out by 25–30 Ma after passage of the MTJ. It is significant that peaks at 18–22 Ma post-MTJ arrival are related to the Coso (CA) and Springerville (AZ) volcanic fields and most likely reflect sampling bias. NAVDAT associates an age date for every geochemical analysis, and the Coso ( $n > 300$ ) and Springerville ( $n > 450$ ) fields are oversampled relative to other fields. Reducing the number of samples from those fields to the average number of samples for the other volcanic fields yields a waning trend of volcanic activity between 6 and 26 Ma post-MTJ arrival. Regardless, the age date distribution provides decisive support for a developing slab window or slab gap

model (Dickinson and Snyder, 1979) in the CA-NV-AZ region. To the extent that age dates in NAVDAT reflect activity, volcanism within the CO-UT-NM region is roughly half that of CA-NV-AZ, with no comparable correlation of age dates relative to MTJ arrival (Fig. 2), indicating that the slab window is well developed beneath CA-NV-AZ and not beneath CO-UT-NM because the latter region is too far east to be affected by slab window processes.

Our observations do not delimit the eastern extent of slab window-related volcanism. The Springerville volcanic field is east of the slab windows identified by Atwater and Stock (1998) and Dickinson (1997), but appears to be within the MTJ-affected region. However, development of the MTJ may have triggered lithosphere thinning east of the slab window. Activity at the Springerville volcanic field perhaps indicates that the approach of the Mendocino Fracture Zone was accompanied by instability in the subducting Farallon slab (Coney and Reynolds, 1977). Although Atwater and Stock (1998) suggested that freezing asthenosphere against the base of continental lithosphere may yield a very limited time span for any particular region of a slab window, our observations (Fig. 2) suggest that the life span may be ~25 Ma. MTJ arrival does not explain volcanic activity farther east, where volcanism began (after a  $>25$  Ma hiatus) at 40 Ma at  $32^\circ N$ . This timing appears to be well in advance of the arrival of the MTJ (Fig. 2).

### Compositional Evidence for an Evolving Slab Window

Volcanic rock composition variations are remarkably well correlated with MTJ arrival time and illustrate the spatial extent and time scale over which the Cordilleran slab window developed (Fig. 4). Ratios of  $La/Nb$  and  $Ba/Nb$ , for example, decrease dramatically ~17 Ma after the MTJ arrival (Figs. 4A, 4B), indicating a rather abrupt shift from a fluid-enriched source to a source that approaches asthenosphere-like depletions in LILEs. This temporal contrast occurs for all volcanic rocks in CA, NV, and AZ, regardless of rock composition. These changes parallel decreases in  $^{87}Sr/^{86}Sr$  and increases in  $^{143}Nd/^{144}Nd$ , which also approach asthenosphere-like values. It is interesting that these isotopic shifts occur much sooner after (5 Ma) MTJ arrival than the LILE/HFSE ratio shifts. In NV and AZ, the decrease is evident for all volcanic rock compositions, but in CA the isotopic shift is only clearly developed among mafic basalts ( $MgO > 6\%$ ; Figs. 4C, 4D). The additional complexity in CA suggests that more evolved rocks there have assimilated more high  $^{87}Sr/^{86}Sr$  and

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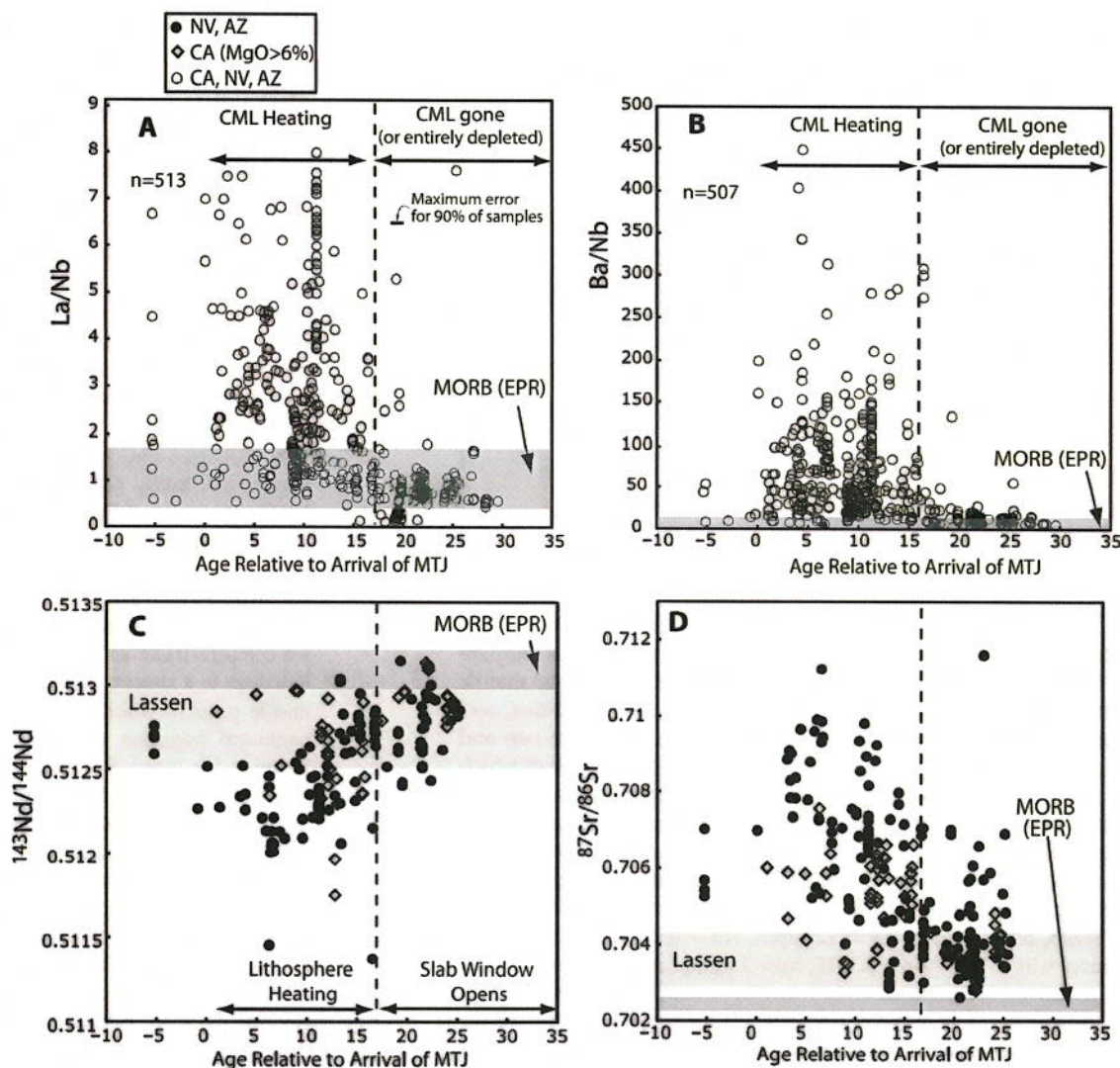


**Figure 4.** La/Nb and Ba/Nb ratios compared to ages of volcanic rocks relative to Mendocino Triple Junction (MTJ) arrival. (A) La/Nb. MORB mid-oceanic ridge basalt; CML—continental mantle lithosphere; EPR—East Pacific Rise. (B) Ba/Nb. La/Nb and Ba/Nb are indices of subduction-related fluid inputs into the mantle source of volcanic rocks. The dramatic decrease in these ratios at 17 Ma after the MTJ arrival indicates the time scale over which such fluids are driven out of CML. (C)  $^{143}\text{Nd}/^{144}\text{Nd}$ . (D)  $^{87}\text{Sr}/^{86}\text{Sr}$ . Both  $^{143}\text{Nd}/^{144}\text{Nd}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  trend toward, but do not uniformly reach, MORB-like ratios that characterize the convective asthenosphere. Because Nd is not easily mobilized by fluids, we surmise that these isotopic ratios reflect the gradual physical removal of CML, which appears to occur immediately after MTJ arrival, whereas the more sudden loss of La- and Ba-bearing components may signal a separate degradation of the fluid-enriched portions of the CML source.

low  $^{143}\text{Nd}/^{144}\text{Nd}$  crustal materials, such as found in the Sierra Nevada Batholith (Putirka et al., 2012). However, the contrast might also result from complexities in mantle sources to the west (CA) compared to the east (AZ and NV; Fig. 5). In CA, Basin and Range volcanic rocks are derived from at least one source with low La/Nb and low  $^{143}\text{Nd}/^{144}\text{Nd}$ , and perhaps at least two other sources, one of which is similar to that beneath NV and AZ, and the other having low  $^{143}\text{Nd}/^{144}\text{Nd}$  and high La/Nb.

#### Volcanism as Controlled by Crustal Extension

McQuarrie and Oskin (2010) suggested that volcanism follows the onset of crustal extension in any given area. To test this idea and to better quantify the relationship, we applied estimates for the timing of the onset and end of crustal extension from McQuarrie and Wernicke (2005, their Fig. 4) to the onset of volcanism using our reconstructed volcanic locations (Fig. 6A). The

data plot into two groups; a black vertical line demarcates the disappearance of a subduction-related signal within erupted volcanic rock compositions (Fig. 6A). Across the central Basin and Range, volcanism migrated westward with time, mostly following the onset of crustal extension. To the east, however, volcanism tends to pre-date extension by 3–5 Ma, while to the west extension predates volcanism by a very consistent average of  $2.0 \pm 0.8$  Ma, exclusive of the Grapevine Mountains vector (see McQuarrie

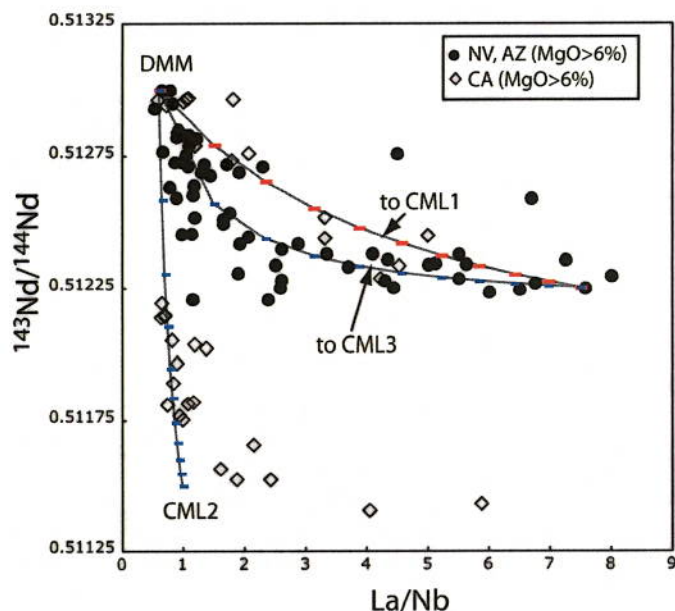


**Figure 5.** To illustrate the complexity of mantle sources, we compare  $^{143}\text{Nd}/^{144}\text{Nd}$  to La/Nb ratios. At least two different continental mantle lithosphere (CML) sources for basaltic rocks were identified, and Basin and Range basalts (>6% MgO) may have at least two and possibly three CML sources (see Putirka et al., 2012), each of which trends toward depleted mid-oceanic ridge basalt mantle (DMM) asthenosphere. The curves illustrate simple mixing (assuming no fractionation of La/Nb during partial melting) between a modified DMM source with  $^{143}\text{Nd}/^{144}\text{Nd} = 0.5130$ , Nd = 0.581 ppm, La = 0.09 ppm, Nb = 0.1485 ppm and a CML1 source with  $^{143}\text{Nd}/^{144}\text{Nd} = 0.51225$ , Nd = 2 ppm, La = 1.5 ppm, Nb = 0.2 ppm and a CML2 source with  $^{143}\text{Nd}/^{144}\text{Nd} = 0.5115$ , Nd = 2 ppm, La = 0.2 ppm, Nb = 0.2 ppm. CML3 is similar to CML1, but has a much higher Nd concentration (7 ppm), in order to better fit the apparent curvature of the Nevada (NV)-Arizona (AZ) basaltic trend. CA—California.

and Wernicke, 2005, table 2 therein). Extension in the Grapevine Mountains putatively predates volcanism by ~7.2 Ma, but that vector is based on a Cordilleran fold-thrust belt correlation that was convincingly disputed by Miller and Prave (2002) and Miller and Pavlis (2005). Consequently, if the magnitude of the displacement is much less, the initiation of extension might be much later than that portrayed.

These time-space patterns correlate well with compositional contrasts. In the east, where volcanism predates extension, volcanism is much more evolved, with a mean composition of dacite, whereas to the west, basaltic andesite is the mean erupted composition (Fig. 6A). This trend in  $\text{SiO}_2$  matches a steady westward decrease in  $^{87}\text{Sr}/^{86}\text{Sr}$  isotopic ratios that is especially evident among andesitic samples (Fig. 6B); only basalts from the unusual Deep Springs locality in eastern CA (Beard and Glazner, 1995) deviate from the trend. These

compositional trends indicate (1) that basaltic precursors in the eastern part of the Basin and Range were more likely to have undergone crustal assimilation and (2) that in the early stages of crustal extension (to the east, near the Colorado Plateau), basalt was less likely to be erupted than more differentiated, less dense counterparts. We posit that in the initial phases of slab window development, extension rates may have been low, and so perhaps impeded the rapid transit of basaltic magmas through thick crust. The inferred time lag between volcanism and the onset of extension in the east (3–5 Ma) is smaller than the time lag of 5–10 Ma inferred by Liu and Furlong (1994) for thermal softening of the lithosphere. The difference between the two measures may well reflect an episode of unerupted basalt underplating of the crust. Alternatively, Parsons et al. (1992) suggested that when extension rates are sufficiently low, vertical dike intrusions increase the horizontal prin-

cipal stress and thereby decrease ambient deviatoric stresses, leading to the intrusion of sills rather than dikes, especially at the base of weak crust. These models are not mutually exclusive. For example, it is possible that (1) during the initial phases of extension (15–20 Ma), basaltic sills partially melted preexisting crust at the western margin of the Colorado Plateau, leading to the eruption of silicic magmas, and subsequently (2), thermally softened crust facilitated extension (which migrated west, as did the locus of volcanic activity), and magmas became more mafic once the stress regime favored eruption of mantle-derived basaltic magmas through vertical conduits.

A final space-time pattern tests the model of Putirka and Busby (2007), wherein the crack interaction model of Takada (1994) was applied to explain the synchronicity of high  $\text{K}_2\text{O}$  eruptions and transtensional normal faulting near the Sierra Nevada Range front at  $38^\circ\text{N}$ . To test whether this relationship is generally applicable, we compare Basin and Range rocks with high K-indices in a temporal context in Figure 7, the middle panel of which (Fig. 7B) illustrates the northward migration of volcanism on which Equation 1 is based. The vast majority of Basin and Range volcanics have a positive K-index (higher than the Cascades), but still mostly <1.9. The locations of all rocks with K-index > 1.9 are not randomly distributed, but are concentrated in three areas: the Yucca Mountain–Southern Nevada volcanic field (Noble et al., 1984; Broxton et al., 1993; Valentine and Perry, 2007), the Colorado River Extensional Corridor (CREC; Howard and John, 1987), and in a narrow band that trends from northeastern NV south to Kane Springs Wash, where it meets the CREC.

In two of these three areas, high K-index volcanism is spatially coincident with areas of high total strain and temporally coincident with high rates of extension. At Yucca Mountain, nearly all rocks with a high K-index (>1.9) were erupted between 11 and 13 Ma, during peak extensional strain between 12.7 and 11.6 Ma (Fig. 7A; Fridrich et al., 1998, 1999). In the CREC, high K-index volcanic rocks define a northward-younging trend (Fig. 7B) that correlates with northward migration of high extensional strain. The earliest phase of CREC volcanism, in the Chocolate Mountains ( $33^\circ\text{N}$ ;  $37^\circ\text{N}$ – $21.5^\circ\text{N}$ ), corresponds to the onset of rapid exhumation and extension in this region (Jacobsen et al., 2002). High K-index rocks erupted between 23 and 18 Ma are coeval and cospatial with accelerated exhumation of the Whipple Mountain metamorphic core complex (Anderson et al., 1988), and the onset of detachment faulting and basin development (at  $\sim 34.5^\circ\text{N}$ , near the Colorado River; Beratan and Nielsen,

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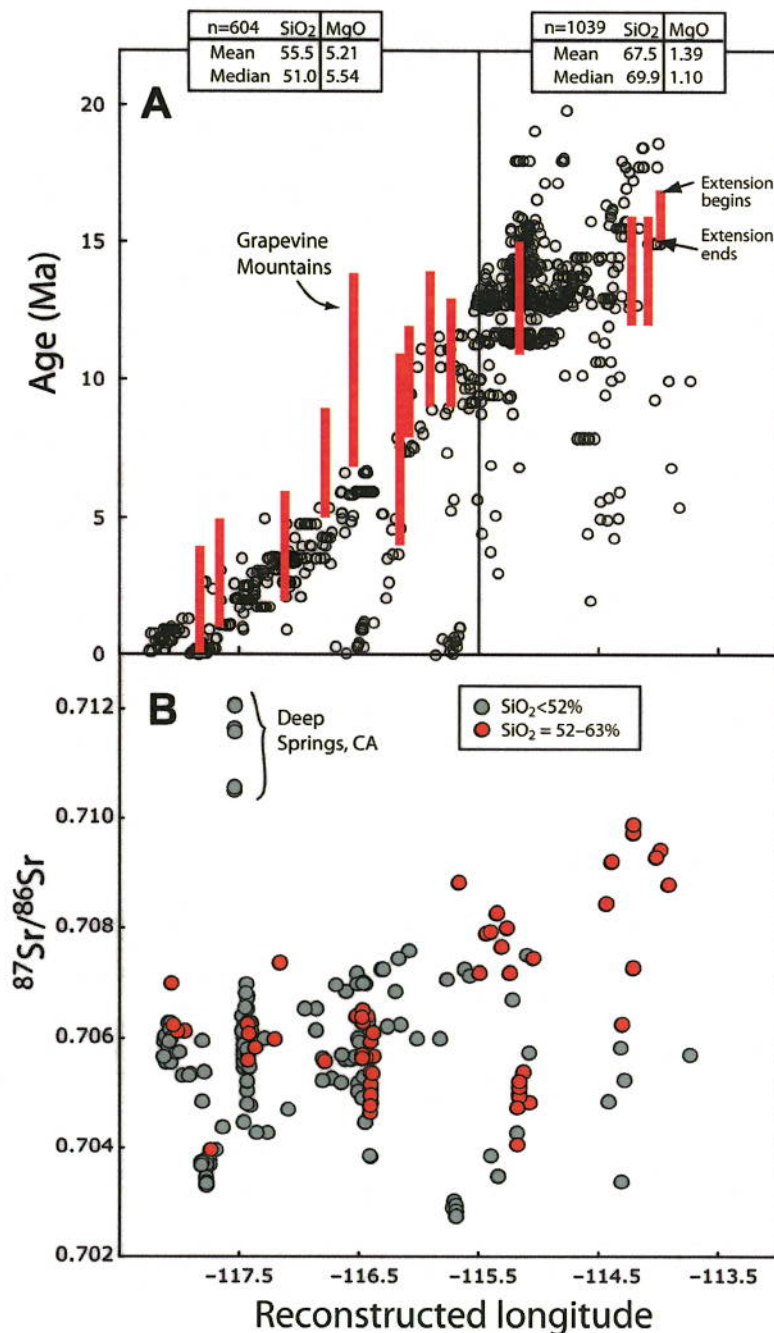


Figure 6. (A) Reconstructed longitudes are used to compare volcanism with the onset of extensional strain. To the east, volcanism precedes extension by as much as 5 Ma, and is highly felsic. To the west, extension leads volcanism by 2 Ma, and volcanism produced more mafic compositions. The black vertical line separates these two regimes, and inset panels show the mean  $\text{SiO}_2$  and  $\text{MgO}$  values for volcanic rocks on either side of the dividing line. (B) Comparison of the  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios for these same volcanic rocks. Basaltic rocks ( $\text{SiO}_2 < 52\%$ ) show no clear variation of  $^{87}\text{Sr}/^{86}\text{Sr}$  from east to west, indicating that spatial compositional variations evident in A are not controlled by the mantle source region. In contrast,  $^{87}\text{Sr}/^{86}\text{Sr}$  for andesites and dacites clearly increase to the east, which indicates that to the east and in the initial phases of extension, magmas tend to pond within the crust and assimilate older continental wall-rock crustal materials. CA—California.

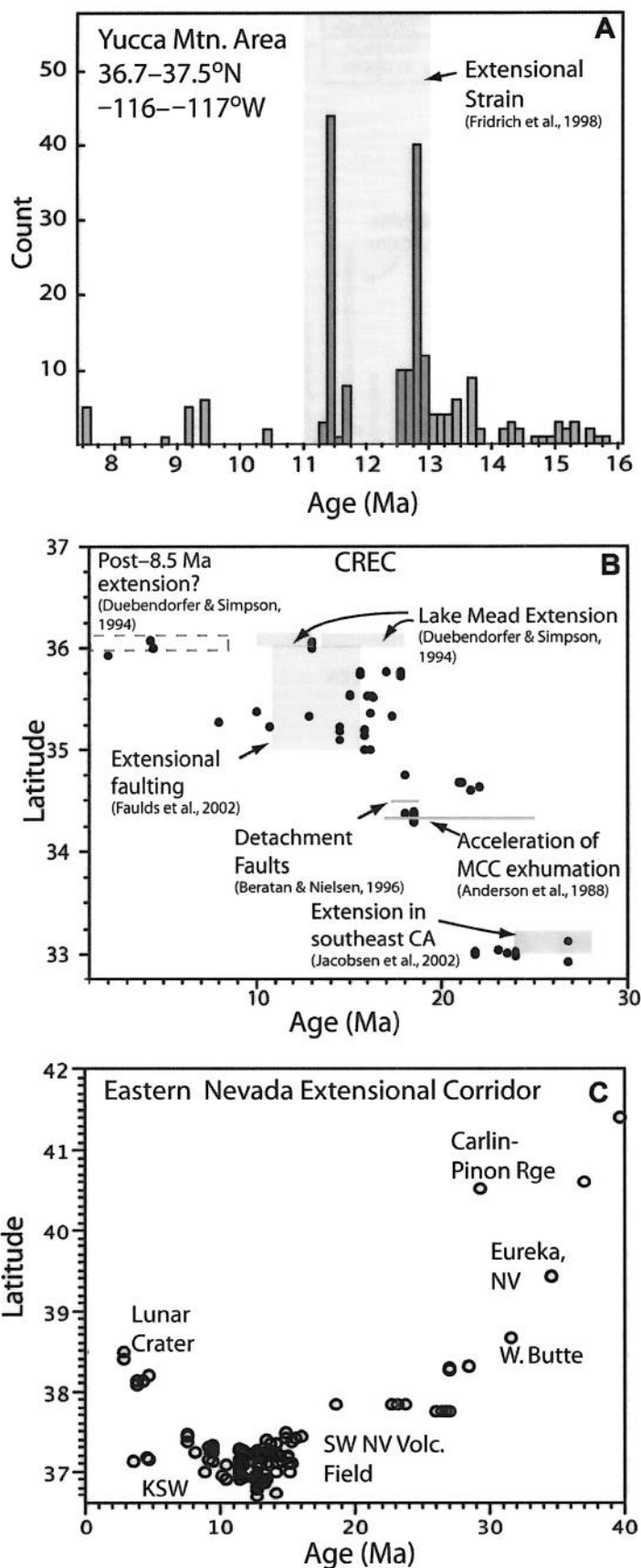
1996) (Fig. 7B). Younger clusters of high-K index volcanics in the CREC, erupted between 18 and 12 Ma, coincide with extensional faulting that Faults et al. (2002) ascribed to a period of major extension between 16.5 and 11 Ma (between 35 and 36°N).

In the Lake Mead region, Duebendorfer and Simpson (1994) (Fig. 7B) documented post-8.5 Ma normal fault activity near Lake Mead that might correlate with the youngest high-K volcanism at the northern end of the CREC. The distribution of high-K-index volcanic rocks near the NV-Idaho border also migrates southward toward the CREC, from Windous Butte (38.6°N, at 31 Ma; Brooks and Snee, 1996), to Indian Peak (38.27°N, at 27 Ma; Best et al., 1989) then south to the southwestern NV volcanic field immediately north of the Yucca Mountain area (37.8°N, at 27–18 Ma; Warren et al., 2000), and culminating at Kane Springs Wash (37.25°N, at 14 Ma; Novak and Mahood, 1986) (Fig. 7C). Strain rates along most of this trend are unknown, but at the southern terminus of the region (at Kane Springs Wash), Scott et al. (1995) found that the time of peak extension rate, at 15–13.5 Ma, coincided with the eruption of peralkaline (i.e., high K-index) rocks at 14.4 Ma. These space-time coincidences indicate that K-index may be a powerful tool for gauging the timing of extensional faulting where stratigraphic or structural information is insufficient.

## Active or Passive Mantle Upwelling?

Whether the initial phases of volcanism at the western margin of the Colorado Plateau, which precede extension, are first driven by active mantle thermal upwelling, or are truly a passive response to slab window opening is not clearly resolved by our space-time-composition comparisons. In Putirka et al. (2007), estimates of mantle potential temperatures were used to discriminate between basaltic magmas generated by mantle plumes (caused by active upwelling of thermal plumes) or passive upwelling of asthenosphere due to lithosphere thinning. Rocks with sufficiently high  $\text{MgO}$  are rare in the Basin and Range, so the methods employed in Putirka et al. (2007) are not easily applicable. However, partial melting depths serve as a proxy for melting temperatures, as applied by Wang et al. (2002) to infer the presence of a mantle plume beneath the Basin and Range. Wang et al. (2002) concluded that partial melting depths increase from eastern CA to the interior of NV, and hypothesized that mantle temperatures should increase correspondingly, signaling a mantle plume situated beneath central NV. In order to evaluate this hypothesis, we extended their profile to the east (as far as the Raton-

**Figure 7.** (A) Histogram of age dates for high K-index ( $>1.9$ ) samples; gray field shows timing of extensional strain, based on field studies of Fridrich et al. (1998). (B) Age dates for volcanic rocks with K-index  $>1.9$  from the Colorado River Extensional Corridor (CREC; Fig. 1) are plotted against latitude. Gray fields mark the timing and spatial extent of high crustal extension rate episodes, based on field studies (citations in figure). In both cases, the timing of high K-index (Equation 2) volcanism coincides with the timing of maximum extensional strain. CA—California. (C) Age dates of volcanic rocks with K-index  $>1.9$  from eastern Nevada (NV) are plotted against latitude for an area we refer to as the Eastern Nevada Extensional Corridor (ENEC; Fig. 1). Field data are insufficient to define the timing of extensional strain in the ENEC; we posit that the southward-migrating trend portrayed in this figure likely reflects migration of high extensional strain similar to that characteristic of the CREC. KSW—Kane Springs Wash; SW NV Volc.—southwest Nevada volcanic field. A–C test the model of Putirka and Busby (2007), which suggests that high  $K_2O$  volcanism is associated with episodes of high extension strain.



Clayton volcanic field in NM) and compared  $SiO_2$  contents for volcanic rocks with 8%–10% MgO for all volcanic rocks (within CA, NV, AZ, UT, CO, and NM) erupted between long 122°W and 103°W (Fig. 8A). This analysis indicates that  $SiO_2$  contents decrease across the entire Cordillera, from a mean of 49.3%  $SiO_2$  west of 121°W to a mean of 46%  $SiO_2$  east of 104°W. These observations are corroborated by qualitative changes in Sm/Yb ratios, which increase with greater depths of melting due to the increased influence of garnet in the mantle source region (Fig. 8C). If high melting pressures were driven by high melting temperatures due to a mantle plume, the plume would then be centered east of the Raton-Clayton volcanic field, i.e., somewhere in Kansas. This seems implausible, which motivates us to search for an alternative control on melting pressures.

Decreasing  $SiO_2$  contents along the west-to-east transect are consonant with a model analogous to that proposed by Haase (1996) for oceanic islands. In Figure 9A, we show how a thermal lithosphere with variable thickness may control magma compositions in the U.S. Cordillera. Our qualitative cross section (Fig. 9A) is based on geophysical results (Goes and van der Lee, 2002) that show a thermal lithosphere that

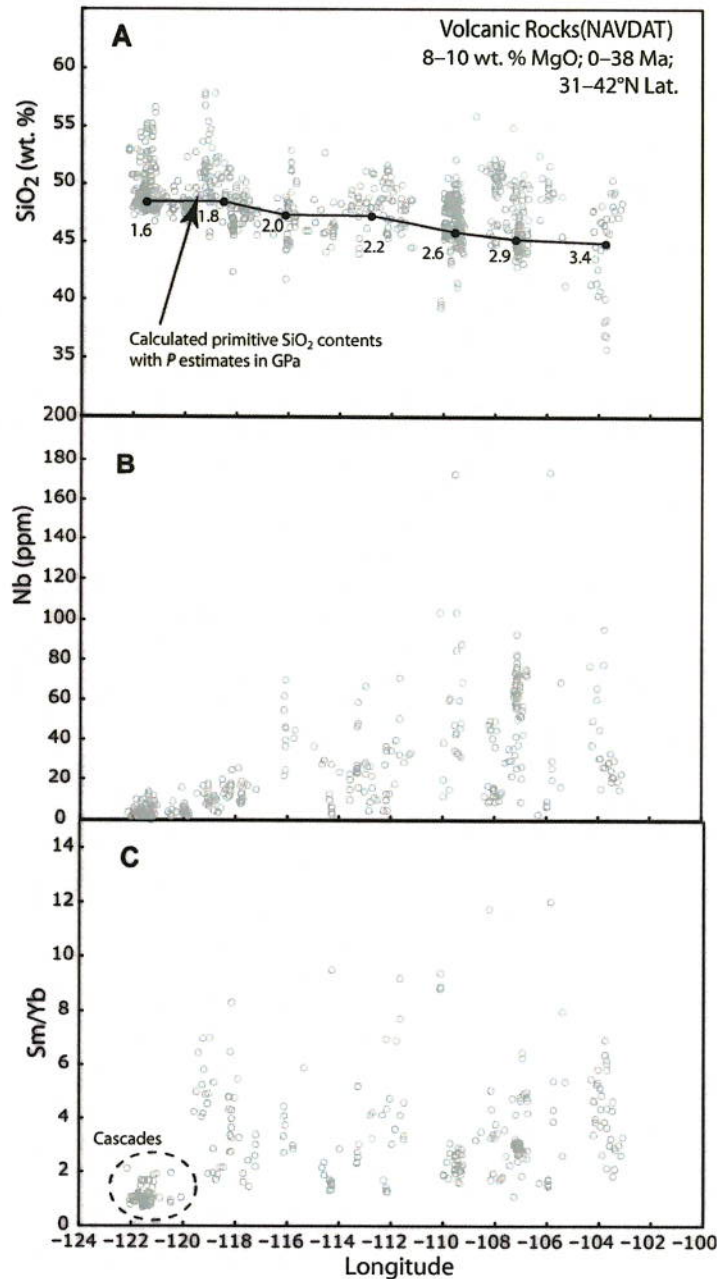
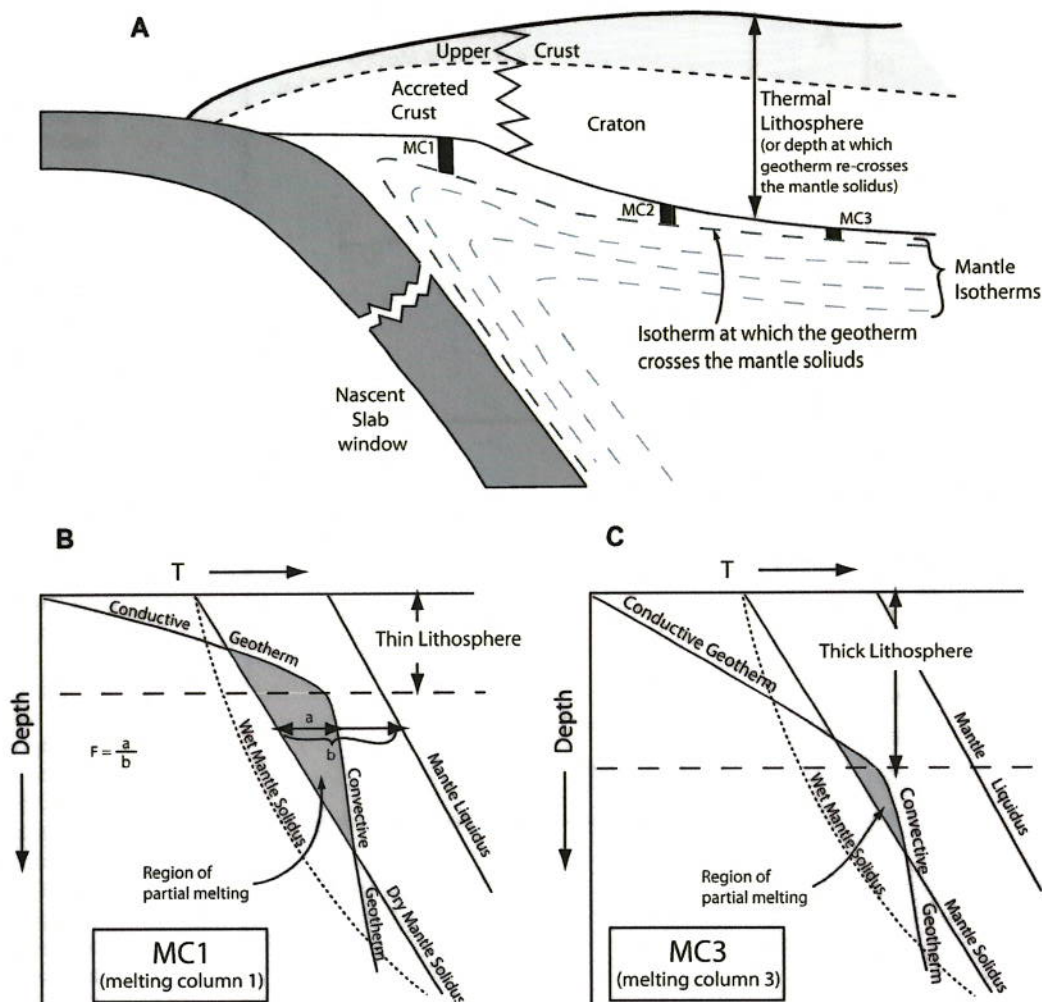


**Figure 8.** Longitudinal variations for Cordilleran volcanic rocks younger than 60 Ma and with  $8\% \leq \text{MgO} \leq 10\%$ , from NAVDAT (North American Volcanic and Intrusive Rock Database; [www.navdat.org](http://www.navdat.org); California, Nevada, Arizona, Utah, Colorado, and New Mexico;  $n = 1033$ ). (A) SiO<sub>2</sub>. (B) Nb. (C) Sm/Yb. Decreasing SiO<sub>2</sub> abundances along the west to east transect imply increasing partial melting depths, as do increasing maximum Sm/Yb ratios. Estimates in A of pressures (*P*) of melt extraction, which increase to the east, were calculated using models of Putirka (2008) (see Putirka et al., 2012). Increasing Nb abundances (and also increasing Sm/Yb, as these ratios are sensitive not just to pressure but also to melt fraction) along the west to east transect indicate smaller degrees of partial melting to the east. The step increase in Sm/Yb ratios from the Cascades to the Basin and Range likely reflects an eastward transition into a mantle lithosphere source (see Putirka et al., 2012, for an analysis of this transition).

thickens from west to east. A key aspect of this model is that the thermal lithosphere controls the top of the melting column. If the mantle is of constant composition and water content, then initial melting depths (the bottom of the melting columns, MC1, MC2, and MC3 in Fig. 9A) will be controlled by the depth of a given isotherm, e.g., the 1300 °C isotherm; that isotherm should reach shallower depths where the lithosphere is thinner (as beneath accreted crust at a continental margin, Fig. 9A), or beneath a slab window, but fall to greater depths where the lithosphere is thick (as beneath colder cratonic crust and mantle). Of course, the mantle is not of constant composition, but as shown in Putirka et al. (2012), subduction-related enrichments in the Basin and Range and Cascades do not vary spatially to any great degree. In any case, the key control in Figure 9A is the fact that a thicker thermal lithosphere means that colder isotherms are depressed downward, and so the top of a melting column beneath cratonic mantle will occur at greater depths (MC3), while beneath thin lithosphere, melting columns will reach much shallower depths (MC1). In this model, melting columns are thus shorter and deeper beneath thick lithosphere (e.g., MC3, compared to MC1). This does not mean that melting cannot occur elsewhere in the lithosphere; the Cordilleran mantle lithosphere is almost certainly spatially and compositionally complex. For example, from the Cascades, there is an eastward step increase in Sm/Yb ratios, which likely indicates an eastward transition into a mantle lithosphere source, perhaps one that contains pyroxenite (see Putirka et al., 2012, for a discussion of the transition). Variations in the thermal lithosphere can both induce the spatial patterns shown in Figure 8 and at the same time avoid the problem of placing a plume beneath the plains of the midwestern U.S. In Figure 9A, we show the case of subduction (or nascent slab window), where isotherms are perturbed by a subducted slab; however, the key effect on the melting columns is hypothesized to be produced by a thermal lithosphere that varies in thickness, so the qualitative results are the same if the slab were to break, detach, and lead to a slab window.

The effect on melt compositions of the model of Figure 9A is qualitatively illustrated in Figures 9B, 9C. Beneath thicker lithosphere (MC3), melt fractions are less because the area bounded by the geotherm and the mantle solidus (whether wet or dry) is reduced (Figs. 9B, 9C). The resulting partial melts will be enriched in incompatible elements (e.g., high K; Fig. 2, or high Nb, and Sm/Yb; Figs. 8B, 8C), and will have a higher pressure signal (lower SiO<sub>2</sub> and higher Sm/Yb), because the mean depth of melting will be greater and the mean *F* will be lower.



Several space-time-composition patterns indicate that Basin and Range volcanism is a passive response to tectonic processes. First, arrival of the MTJ provides a clear temporal trigger for the onset of volcanism south of 36.75°N (Fig. 2). Between 5 and 20 Ma after MTJ arrival, isotopic and trace element patterns also shift from continental mantle lithosphere toward asthenosphere compositions, signaling the time scale of lithosphere degradation and removal (Fig. 4) (see also Putirka et al., 2012). The role of passive tectonics relative to mantle

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processes is further supported by spatial trends of variable  $\text{SiO}_2$  contents in basaltic rocks; decreasing  $\text{SiO}_2$  contents across the entire Cordillera indicate that partial melting depths are controlled by lithosphere thickness rather than a mantle plume. Additionally, the effects of slab window processes are discernible as far east as eastern NV and AZ. Reconstructed volcanic rock locations in the central Basin and Range also largely confirm the view of McQuarrie and Oskin (2010) that volcanism follows the onset of extensional deformation (we calculate the lag to be ~2 Ma; Fig. 6A). High- $\text{K}_2\text{O}$  volcanism is especially sensitive to extension and is characteristic of significantly extended regions at times dominated by high extension rate (Fig. 7).

However, neither near-surface extension nor volcanism ubiquitously leads the geologic evolution of any particular place within the Basin and Range. Adjacent to the Colorado Plateau, where central Basin and Range extension was initiated, volcanic rocks are silicic, contaminated by crustal assimilation (high  $^{87}\text{Sr}/^{86}\text{Sr}$ ), and their eruption preceded extension by 3–5 Ma. Consequently, during the early phases of continental extension, stress states might have favored sill formation, rather than dike injection (Parsons et al., 1992), and thermal softening (Liu and Furlong, 1994) may have been required to initiate subsequent, more rapid crustal extension and basaltic volcanism. Later, and to the west, however, extension precedes volcanism by 2 Ma and becomes much more mafic in composition. This 2 Ma lag may have allowed development of a favorable stress regime within the crust that fostered magma transport from a deep mantle source to the surface (Takada, 1994; Putirka and Busby, 2007); this process is perhaps best indicated by eruption of low-degree partial melts (alkalic or high  $\text{K}_2\text{O}$  magmas) in regions of high extension during episodes dominated by high extension rate.

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