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Cenozoic Tectonics, Magmatism, and Stratigraphy of the Snake River Plain–Yellowstone Region and Adjacent Areas themed issue

Yellowstone plume trigger for Basin and Range extension, and coeval emplacement of the Nevada–Columbia Basin magmatic belt

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ABSTRACT

Widespread extension began across the northern and central Basin and Range Province at 17–16 Ma, contemporaneous with magmatism along the Nevada–Columbia Basin magmatic belt, a linear zone of dikes and volcanic centers that extends for >1000 km, from southern Nevada to the Columbia Basin of eastern Washington. This belt was generated above an elongated sublithospheric melt zone associated with arrival of the Yellowstone mantle plume, with a north-south tabular shape attributed to plume ascent through a propagating fracture in the Juan de Fuca slab. Dike orientation along the magmatic belt suggests an extension direction of 245°–250°, but this trend lies oblique to the regional extension direction of 280°–300° during coeval and younger Basin and Range faulting, an ~45° difference. Field relationships suggest that this magmatic trend was not controlled by regional stress in the upper crust, but rather by magma overpressure from below and forceful dike injection with an orientation inherited from a deeper process in the sublithospheric mantle. The southern half of the elongated zone of mantle upwelling was emplaced beneath a cratonic lithosphere with an elevated surface derived from Late Cretaceous to mid-Tertiary crustal thickening. This high Nevadaplano was primed for collapse with high gravitational potential energy under the influence of regional stress, partly derived from boundary forces due to Pacific–North American plate interaction. Plume arrival at 17–16 Ma resulted in advective thermal weakening of the lithosphere, mantle traction, delamination, and added buoyancy to the northern and central Basin and Range. It was not the sole cause of Basin and Range extension, but rather the catalyst for extension of the Nevadaplano, which was already on the verge of regional collapse.

INTRODUCTION

The Basin and Range Province is one of the best exposed extensional areas in the world for studying the effects and causes of large-scale stretching of continental crust. The boundaries of this extended terrain were originally defined by the American physiographer N.M. Fenneman (1928, 1931), with subdivisions defined by later workers. In this paper, the province is subdivided into northern, central, and southern segments (Fig. 1), following the terminology of Jones et al. (1992) and Wernicke (1992).

The Cenozoic extensional history of the Basin and Range Province (Fig. 1) has been explained through a variety of models, which include: (1) broadly distributed shear of the plate interior driven by right-lateral motion of the Pacific plate (e.g., Atwater, 1970; Livaccari, 1979); (2) thickening of the crust sufficient to produce buoyancy-driven extensional strain (e.g., Sonder et al., 1987; Jones et al., 1998); or (3) subslab upwelling and lateral spreading of asthenosphere, derived either from a "slab window" behind the trailing edge of the Farallon plate (e.g., Dickinson and Snyder, 1979), or from the adiabatic rise of the Yellowstone mantle plume (e.g., Parsons et al., 1994; Saltus and Thompson, 1995; Pierce et al., 2002).

Sonder and Jones (1999) evaluated these and other models quantitatively by applying geological and geophysical constraints on the driving forces and the resisting forces inherent in each model. They subdivided driving forces into four groups: boundary forces associated with the relative motions of the Pacific, North American, and Juan de Fuca–Farallon plates, tractional forces applied to the base of the lithosphere, buoyancy forces associated with lithospheric density variations, and basal normal forces associated with mantle upwelling and/or gravitational instabilities. They concluded that boundary forces associated with plate interaction would produce neither the magnitude nor the rates of extension observed in the northern and central Basin and Range, and that, at best, these forces can only augment or modify the other forces necessary for continental extension. They also concluded that buoyancy forces have been the primary control for extension of the northern Basin and Range, but that boundary forces might have played a greater role in driving extension in the southern Basin and Range (Fig. 1).

This quantitative approach led Sonder and Jones (1999) to attribute most of the extension in the northern Basin and Range to gravitational potential energy, where crustal thickening and isostatic rise led to gravitational collapse of a high orogenic plateau (Dewey, 1988). Such a model is consistent with crustal shortening and the development of fold-and-thrust belts in central and eastern Nevada and western Utah during the late Mesozoic to early Tertiary Sevier-Laramide orogeny. By the end of the Late Cretaceous, this compressional regime had thickened the crust to an estimated 50–70 km (DeCelles, 2004; Best et al., 2013; Lechler et al., 2013), with a regional paleoelevation that was 3–4 km (e.g., Chase et al., 1998; Wolfe et al., 1998; Chamberlain et al., 2012). Gravitational potential energy variations appear to play the major role in extension today, as suggested by the modeling results of Flesch et al. (2000), Humphreys and Coblentz (2007), and Ghosh et al. (2013).

The effects of mantle upwelling (basal normal forces), however, are virtually indistinguishable from those produced by buoyancy

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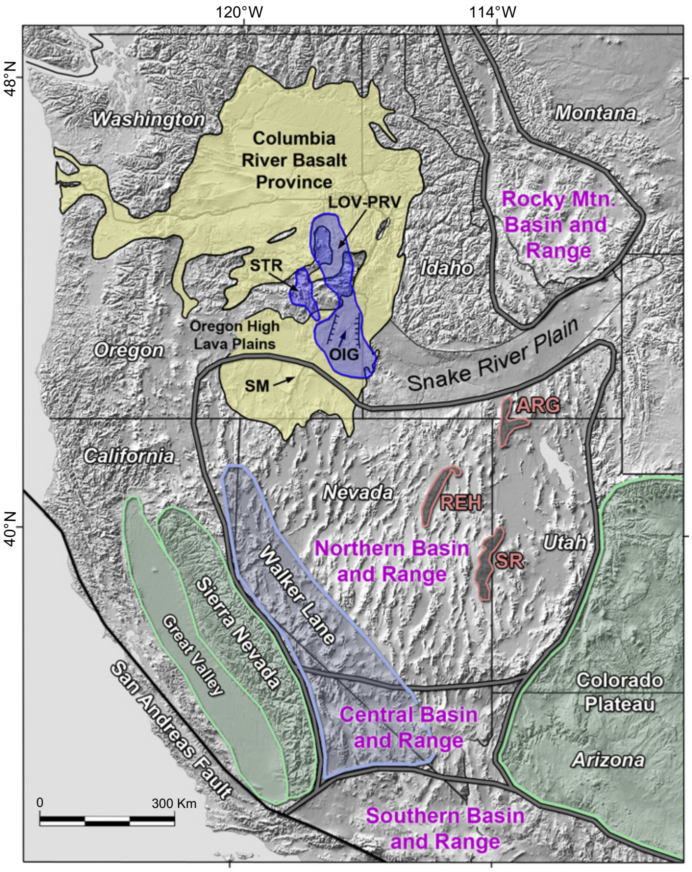

Figure 1. The Basin and Range Province, with northern, central, and southern segments modified from Jones et al. (1992) and Wernicke (1992). The northern and central Basin and Range are bounded to the west by the Sierra Nevada batholith and to the east by the Colorado Plateau. The northern Basin and Range incorporates most of Nevada and western Utah. Its northern boundary is transitional with the Oregon High Lava Plains and the southern extension of the Columbia River Basalt Province, which contains the initial eruptions of flood basalt in the vicinity of Steens Mountain (SM). It also extends to the northeast, south of the Snake River Plain. Coeval extension north of the Snake River Plain is sometimes referred to as the Rocky Mountain Basin and Range (Sonder and Jones, 1999). Middle Miocene to Pliocene volcanic fields lying north of the northern Basin and Range include the Strawberry volcanics (STR) and the Lake Owyhee–Powder River volcanic field (LOV-PRV). Metamorphic core complexes are located in the Snake Range (SR), the Albion–Raft River–Grouse Creek Ranges (ARG), and the Ruby Mountains–East Humboldt Range (REH). The Oregon-Idaho graben (OIG) is not typically considered part of the northern Basin and Range, although extension there was contemporaneous with mid-Miocene extension in north-central Nevada. The central Basin and Range incorporates the Colorado River extensional corridor of southern Nevada and adjacent California and Arizona. The southern Basin and Range lies southeast of the Sierra Nevada of southern California, extending eastward into the Sonora Desert of southern Arizona and southward beyond the confines of the map into Mexico.

forces associated with a thickened crust (Sonder and Jones, 1999). A vigorously upwelling mantle would increase driving forces not only by increasing horizontal asthenospheric flow, but also by warming the lithosphere, thus decreasing its strength, decreasing its densities, and increasing lithospheric buoyancy. Similar to crustal thickening, mantle upwelling can generate dynamic topographic uplift and high gravitational energy subject to normal faulting over a wide region (Houseman and England, 1986). Thermorheological and thermomechanical modeling suggests that a thickened crust may be unstable and may tend to collapse, but that major postorogenic extension happens only when the lithosphere is sufficiently weakened by a strong thermal perturbation in the mantle (Liu and Shen, 1998a; Liu, 2001).

Such studies lend credence to the idea that mantle upwelling associated with the Yellowstone hotspot might have provided the driving forces necessary for widespread Basin and Range extension (Pierce and Morgan, 1992; Parsons et al., 1994; Saltus and Thompson, 1995; Rowley and Dixon, 2001; Pierce et al., 2002). Traditional arguments against such a model have been partly based on the lack of supporting data from high-resolution seismic tomography and poorly constrained age correlations between the initiation of mantle upwelling and the timing of extension in the northern Basin and Range. There has also been a lack of interdisciplinary studies between workers with structural and tectonic interests in the Basin and Range Province and those with geophysical and petrological interests in mantle upwelling associated with the Columbia Plateau–Snake River Plain–Yellowstone magmatic system. In an attempt to help bridge this gap, this paper reviews a broad range of studies from both of these regions that appear to support a genetic relationship between mid-Miocene mantle upwelling of the Yellowstone plume and the contemporaneous onset of volcanism and crustal extension in the northern and central Basin and Range.

GEOLOGIC SETTING

Basin and Range extension was preceded by a protracted period of Mesozoic to Cenozoic (ca. 155–55 Ma) compression as the Farallon plate subducted beneath North America. Crustal shortening was associated with a complex history of deformation along the Luning-Fencemaker thrust belt of northwestern Nevada (pre–143 Ma), the Sevier fold-and-thrust belt of eastern Nevada and western Utah (ca. 165–80 Ma), and the broader Laramide belt of block uplifts and structural basins in eastern Utah and Colorado (ca. 70–40 Ma) (e.g., Wyld et al., 2003; Dickinson, 2013). By Late Cretaceous time, contraction in central and eastern Nevada had produced an overthickened crust (Coney, 1987), topographically expressed as a broad, low-relief, high-elevation plateau. This elevated paleosurface has been called the “Great Basin altiplano” by Best et al. (2009), or the “Nevadaplano” by DeCelles (2004), analogous to the present-day central Andean Altiplano of Bolivia and northern Argentina (Isacks, 1988).

After a lull in magmatic activity from ca. 90 to 45 Ma, magmatism resumed on this elevated plateau as a time-transgressive sweep of volcanic activity from ca. 45 to 18 Ma that propagated from north to south into southern Nevada (Fig. 2A; Best et al., 1989; Armstrong and Ward, 1991; Christiansen and Yeats, 1992). An evaluation of 4861 oxygen isotope analyses suggests that the volcanic sweep was coincident with topographic uplift that began in the north and swept southward across Nevada from ca. 40 to 23 Ma, further increasing the mean elevation of the Nevadaplano to ~4 km (Chamberlain et al., 2012). This period of uplift appears to have been

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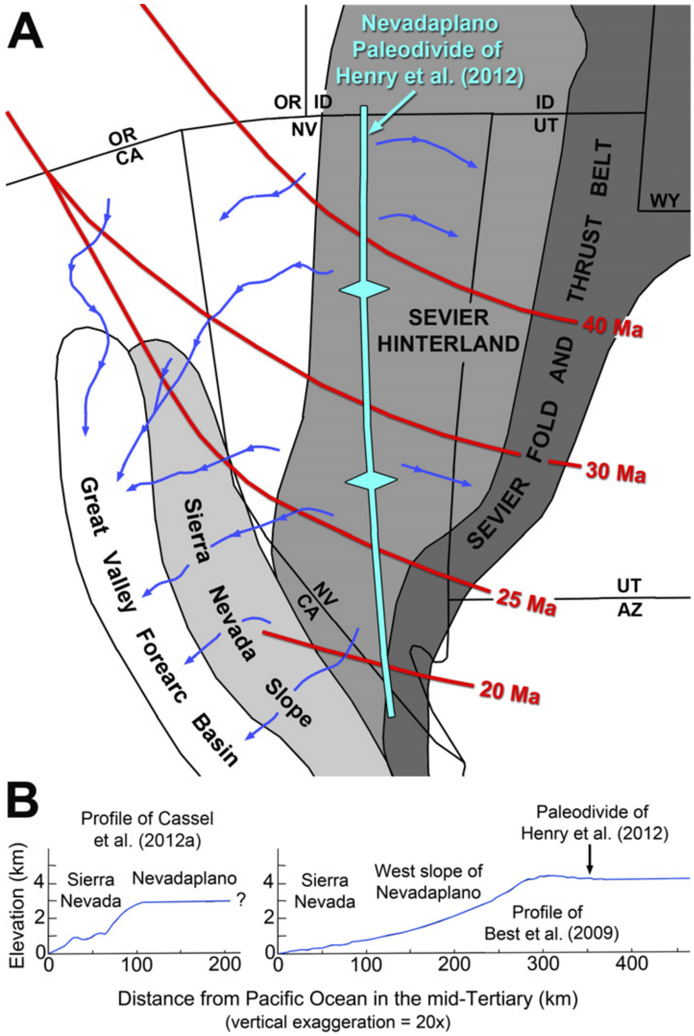

Figure 2. Mid-Tertiary volcanism and paleotopography across eastern California, Nevada, and western Utah prior to Neogene Basin and Range extension. (A) Map of tectonomagmatic relationships with state borders distorted due to palinspastic restoration at ca. 20 Ma. Red lines denote successive volcanic fronts of the mid-Tertiary volcanic sweep (from Dickinson, 2013). Paleodivide of the Nevadaplano is from Henry et al. (2012), separating eastward-draining and westward-draining paleovalleys of the Nevadaplano, marked by blue lines with arrows (modified from Henry, 2008; Henry et al., 2012; Dickinson, 2013). OR—Oregon, CA—California, NV—Nevada, ID—Idaho, UT—Utah, WY—Wyoming, AZ—Arizona. (B) Proposed paleotopographic profiles across the Sierra Nevada and Nevadaplano in the mid-Tertiary (modified from Henry et al., 2012). Although the geologic data suggest that the Sierra Nevada formed a lower western slope to the Nevadaplano, the nature of this slope differs in these two profiles. Cassel et al. (2012a) used stable isotope studies to suggest that the mid-Tertiary Sierra Nevada had an elevation and gradient similar to today, with the Nevadaplano forming a flat high-altitude plateau immediately east of the Sierra Nevada ramp. An alternative interpretation is given by Best et al. (2009), who suggested that the Nevadaplano had a near-constant western gradient to a flat-lying plateau located ~200 km east of the Sierra Nevada, but ~100 km east of the paleodivide of Henry et al. (2012).

associated with upwelling asthenosphere due to slab rollback (Coney and Reynolds, 1977; Severinghaus and Atwater, 1990; McQuarrie and Oskin, 2010; Dickinson, 2013) or to removal of the underlying Farallon slab (Humphreys, 1995), with further crustal thickening produced by partial melting during adiabatic mantle rise. It is also recorded in new thermochronometric and paleogeographic data, which demonstrate significant unroofing of the southwestern Colorado Plateau between 28 and 16 Ma (Flowers et al., 2008; Lamb et al., 2011; Lee et al., 2013).

The mid-Tertiary volcanic and topographic sweep was associated with large-volume eruptions of the "ignimbrite flare-up" (Coney, 1978), which generated >70,000 km^3 of felsic pyroclastic rocks from ca. 36 to 18 Ma (Best et al., 2013). Mapping of some of these massive ash-flow tuffs from Nevada into westward-draining paleovalleys demonstrates that the Sierra Nevada formed a lower western ramp to the Nevadaplano and that the structural and topographic boundary between the Sierra Nevada and the Basin and Range did not exist before 23 Ma (Fig. 2B; Henry et al., 2012). Multiple other lines of evidence, including $\delta^{18}O$ values in hydrated volcanic glass, support the conclusion that the Nevadaplano maintained a higher elevation than the Sierra Nevada at least into the Miocene (Cassel et al., 2012a, 2014).

TIMING OF EXTENSION: BEFORE 17 MA

Several workers have suggested that the Nevadaplano was subject to significant extension before and during the ca. 45–18 Ma volcanic sweep (Coney, 1980; Zoback et al., 1981; Gans and Miller, 1983; Gans, 1987; Christiansen, 1989; Gans et al., 1989; Livaccari, 1991). This idea is partly based on deformation associated with the Cordilleran metamorphic core complexes that was extrapolated to the entire Great Basin (Armstrong, 1982; Coney and Harms, 1984). In contrast, more recent studies suggest that there was only minor, localized extension in the Eocene and no significant extension in the Oligocene and early Miocene, but that major and widespread extension began in the middle Miocene at ca. 17–16 Ma (e.g., Best and Christiansen, 1991; Wernicke and Snow, 1998; Faulds et al., 2001; Henry, 2008; Wallace et al., 2008; Colgan and Henry, 2009; Colgan et al., 2010; Best et al., 2013).

There are only three notable core complexes in the northern Basin and Range: Albion–Raft River–Grouse Creek Ranges in northwestern Utah and southern Idaho, the Ruby Mountains–East Humboldt Range of northeastern Nevada, and the Snake Range of east-central Nevada, (Fig. 1). Although these core complexes show multiphase deformational histories, there is a lack of structural, volcanic, or sedimentary evidence in each for pre–middle Miocene extension at upper-crustal levels. Fission-track and (U/Th)/He dating in the southern part of the Ruby Range suggests that unroofing there did not occur until 17–16 Ma (Colgan and Metcalf, 2006; Colgan and Henry, 2009). Fission-track dating in the Snake Range demonstrates that the major period of extension there also began at 17 Ma, contemporaneous with mid-Miocene extension in the adjacent Shell Creek Range to the west and Kern Mountain Range to the south, where the weighted mean age for extension is 16.41 ± 0.25 Ma (Miller et al., 1999). Konstantinou et al. (2012) presented new geochronological and structural data showing that the Albion–Raft River–Grouse Creek metamorphic complex remained at midcrustal depths until it was exhumed by middle Miocene Basin and Range extension at ca. 14 Ma, contemporaneous with the synextensional Raft River Basin at ca. 13.5 Ma.

Beyond the confines of the metamorphic core complexes, there is little evidence for significant pre–middle Miocene extension in the northern and central Basin and Range. In northeast Nevada, Henry (2008) mapped a series of 45–40 Ma ash-flow tuffs that lie within Eocene paleovalleys that flowed off the Nevadaplano eastward into the Uinta Basin in Utah. The continuity of the paleovalleys, the ability of the ash-flow tuffs to flow long distances down the paleovalleys, and the mostly concordant contacts between the Eocene tuffs and Miocene deposits all suggest that little, if any, extensional faulting occurred between the Eocene and middle Miocene (Henry, 2008). In a similar fashion, the 28.9-Ma tuff of Campbell Creek and other tuffs advanced over vast stretches of northern Nevada but also descended down west-draining paleovalleys from northern Nevada into the Sierra Nevada (Henry et al., 2012; Henry and John, 2013). The lack of disruption in these paleodrainages indicates little or no faulting in western Nevada before 23 Ma, and before 29 Ma in northern Nevada as far east as the Ruby Mountains (Henry et al., 2012). The lack of wide, north-trending paleovalleys suggests that the type of east-west extension that produced the current Basin and Range topography was not present during the Oligocene.

In the broader ignimbrite province of south-central Nevada, substantial and widespread angular unconformities are lacking, along with a virtual absence of sedimentary deposits. Although some of the widespread ignimbrites record local extension near their eruption centers, there is no direct or indirect evidence of large-scale, regional extension associated with the ignimbrite flare-up from ca. 36 to 18 Ma (Best and Christiansen, 1991; Best et al., 2009, 2013).

In a 200-km-long, east-west transect across north-central Nevada (lat 40°30′N), Colgan and Henry (2009) noted the absence of 25–17 Ma sedimentary rocks, which indicates the lack of depocenters that should develop during crustal extension. Further south (lat 39°N), in an ~600-km-long, east-west transect across Nevada and western Utah, McQuarrie and Wernicke (2005, their Table 1) noted very minor extension between 35 and 25 Ma, but with most extension occurring after 18 Ma. Best and Christiansen (1991) pointed out that there are a number of areas in the northern Basin and Range where episodic deformation has taken place with locally significant extension before ca. 35 Ma and after ca. 17 Ma (e.g., Proffett, 1977; Proffett and Dilles, 1984; Taylor et al., 1989), but that there is no evidence for continuous or regionally significant extension during the ignimbrite flare-up. Farther south, in the central Basin and Range, several workers have documented the absence of significant extension prior to ca. 17 Ma (Wernicke and Snow, 1998; Snow and Wernicke, 2000; Faulds et al., 2001; McQuarrie and Wernicke, 2005).

Evidence from a variety of sources does not support the concept of widespread Eocene to early Miocene extension in the northern and central Basin and Range. The specific areas where pre–17 Ma extension has been documented (e.g., Proffett, 1977; Smith, 1992; Dilles and Gans, 1995; McGrew et al., 2000) appear to be geographically isolated, in some cases episodic, and unrelated to a long-lived event of regional scope. These observations are consistent with the lack of $\delta^{18}O$ evidence for regional collapse of the Nevadaplano, and the contrary evidence for a north-to-south migration of uplift in Nevada during the mid-Tertiary volcanic sweep (Horton et al., 2004; Cassel et al., 2012b; Chamberlain et al., 2012).

TIMING OF EXTENSION: AFTER 17 MA

In north-central Nevada, evidence for broadly based extension after 17 Ma comes from thermochronology and basin analysis of 12 middle Miocene basins south of Winnemucca, Battle Mountain, and Wells, Nevada (between lat 40°N and 41°N and long 115°W and 118°W; Colgan et al., 2008; Wallace et al., 2008). Colgan and Henry (2009) noted that the precise onset of extension in these basins is difficult to determine because of the error uncertainties inherent in apatite fission-track ages and from individual (U-Th)/He ages, but that significantly more precise ages come from sanidine $^{40}Ar/^{39}Ar$ dates of interbedded tuffs. They pointed out that there is always some part of the middle Miocene stratigraphy that lies below the lowest dated sample, so that radiometric dates must be considered minimal ages for the initiation of extension. The oldest $^{40}Ar/^{39}Ar$ date is ca. 16.3 Ma from the Carlin Basin (Wallace et al., 2008), with other dates as old as 15.8 and 15.9 Ma from basins in the central Shoshone Range south of Battle Mountain (Colgan and Henry, 2009).

Based on these dates, together with low-temperature thermochronology and structural/stratigraphic considerations, Colgan and Henry (2009) suggested that the onset of extension in north-central Nevada began at ca. 17–16 Ma and continued with waning intensity until 12–10 Ma. Current data from a wide variety of sources (summarized in McQuarrie and Wernicke, 2005) show that the greatest period of extension in the southern half of the northern Basin and Range also began after ca. 17 Ma, contemporaneous with the advent of abundant sedimentary deposition that indicates significant erosion (Best et al., 2013). Stockli (2005) summarized data from low-temperature thermochronology across the central part of the northern Basin and Range (near lat 40°N) that indicate the rapid cooling of fault blocks from a presumed period of extension that began at ca. 18–15 Ma. In the central Basin and Range, nearly all documented extension began at ca. 16 Ma, with more rapid motion in the early part of the deformational history, from ca. 16 to 10 Ma (Wernicke and Snow, 1998). Quigley et al. (2010) used apatite fission-track ages to show that rapid cooling and extension began at ca. 17 Ma in the transition zone between the Basin and Range Province and the Colorado Plateau. Many studies on individual faults in the central Basin and Range have documented rapid slip that began ca. 17–15 Ma (e.g., Fitzgerald et al., 1991; Gans and Bohrson, 1998; Miller et al., 1999; Reiners et al., 2000; Faulds et al., 2001; Carter et al., 2006; Fitzgerald et al., 2009). Minor extension between 16.5 and 15 Ma is also recorded in the northern Nevada rift (John et al., 2000), and in the Goosey Lake depression of the Santa Rosa–Calico volcanic field of northern Nevada (Vikre, 2007; Brueseke and Hart, 2008).

The youngest (<12 Ma) extension in the northern Basin and Range was of lower magnitude and concentrated largely at its eastern, western, and northernmost margins (e.g., Gilbert and Reynolds, 1973; Dilles and Gans, 1995; Armstrong et al., 2003; Stockli et al., 2003). In northwestern Nevada, south of Winnemucca, this younger distinct period of faulting is superimposed on the older middle Miocene distribution of extended terrains (Fosdick and Colgan, 2008). Farther north in southern Oregon and in the northwesternmost corner of Nevada, however, there is a lack of evidence for significant extension between 17 and 12 Ma, with only the younger phase of extension (<12 Ma) evident in the geologic record (Colgan et al., 2004, 2006; Meigs et al., 2009; Scarberry et al., 2010). A localized exception is the Oregon-Idaho graben in east-central Oregon (Fig. 1), where minor subsidence and coeval magmatism began between 15.3 and 14.3 Ma, and extensional faulting is recorded between 14.3 and 12.6 Ma

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(Cummings et al., 2000). In the central Basin and Range, most younger extension (<12 Ma) has been associated with the northwest movement of the cohesive Sierra Nevada–Great Valley block (Fig. 1), which has required a significant component of right-lateral transtension (McQuarrie and Wernicke, 2005).

In summary, there appears to be good evidence for a prolonged period of relative tectonic quiescence during continued uplift (i.e., the $\delta^{18}O$ evidence of Chamberlain et al., 2012) on the Nevadaplano from ca. 45 to 17 Ma, but with moderate, localized, and in places episodic, extension evident in the rock record. The evidence is equally consistent with a major and widespread period of extension that began abruptly at ca. 17–16 Ma and continued with waning intensity to 12–10 Ma to produce NNE-trending depositional basins and parallel fault blocks across a wide expanse of the northern and central Basin and Range. Extension continued after 12–10 Ma, but this was of lesser magnitude and focused on the western, eastern, and northernmost margins of the province.

MIDDLE MIOCENE VOLCANISM AND THE NEVADA–COLUMBIA BASIN MAGMATIC BELT

Workers have long noted that the interior of the western United States experienced an abrupt mid-Miocene transition in the style of volcanism, where a southward sweep of calc-alkaline intermediate to rhyolitic volcanism without basalt (ca. 36–18 Ma) was followed by a later period (ca. 17 to present) of tholeiitic to mildly alkaline basalt or bimodal volcanism (e.g., Christiansen and McKee, 1978; Christiansen et al., 2002). There was some calc-alkaline intermediate volcanism after 17 Ma, but this was specifically related to the mid-Miocene magmatism associated with the ancestral southern Cascades volcanic arc along the eastern edge of the Sierra Nevada and in the Walker Lane belt (Fig. 3; e.g., Busby et al., 2008; Cousens et al., 2008; Busby and Putirka, 2009; Colgan et al., 2011).

During the peak of extension, from ca. 17 to 10 Ma, broad regions of the northern and central Basin and Range (excluding the Walker Lane belt) were amagmatic, with significant mafic to bimodal volcanism localized in specific areas, mostly in the north-central part of the northern Basin and Range. These areas include the northern Nevada rift system (Fig. 3), where bimodal, mildly alkalic basalt/andesite and rhyolite erupted from ca. 16.5 to 15 Ma (Zoback et al., 1994; John et al., 2000), and in the earliest calderas at the western end of the Yellowstone–Snake River Plain hotspot track, where massive rhyolitic eruptions were generated from ca. 16.5 to 15.5 Ma (e.g., Brueseke et al., 2008, 2014; Coble and Mahood, 2012). This time interval also corresponds with the main-phase eruption of Columbia River flood basalt, from 16.7 to 15 Ma, in an area lying north of the Basin and Range Province (Barry et al., 2013; Camp et al., 2013; Reidel et al., 2013a). Over this short time interval, dike swarms developed along all major eruption sites associated with flood-basalt volcanism, first erupting from the Steens dike system in southeastern Oregon at ca. 16.7 Ma, and then migrating rapidly to the north into the Monument dike swarm of north-central Oregon at ca. 15.8 Ma and into the massive Chief Joseph dike swarm of northeastern Oregon and southeastern Washington, where the main phase of the Columbia River flood basalts erupted from ca. 16.6 to 15.6 Ma (Fig. 3; Camp and Ross, 2004; Barry et al., 2013; Reidel et al., 2013a).

The northern Nevada rift zone is partly delineated by numerous basaltic dikes along its length. The rift itself coincides with a prominent aeromagnetic anomaly that appears to be one part of a larger system of mafic intrusions (Blakely and Jachens, 1991). Glen and Ponce (2002) and Ponce and Glen (2002) used gravity inversion and aeromagnetic filtering techniques to identify at least three, and possibly six, large-scale, north-trending and arcuate anomalies across an east-west distance of ~150 km in northern Nevada, with the most prominent anomaly coinciding with the northern Nevada rift.

These geophysical anomalies delineate a system of large, linear intrusions, 4–7 km wide, at ~12–15 km depth (Ponce and Glen, 2002, 2008). One of these deep intrusions strikes into the area occupied by the Steens Mountain dike swarm, consistent with it being a large keel dike, or linear magma chamber for the surface dikes (Fig. 4). Filtering techniques allowed Glen and Ponce (2002) to map two of these anomalies northward into the Oregon-Idaho graben of eastern Oregon (Fig. 1). The prominent anomaly beneath the northern Nevada rift appears to extend much farther to the south, through the center of the northern Basin and Range to the Arizona border (Fig. 4); however, it has a weaker signature in southern Nevada and lacks evidence of surface volcanism (Blakely and Jachens, 1991). Still farther south, Anderson et al. (1994) extended the northern Nevada rift into the Colorado River extensional corridor and Lake Mead region, where mid-Miocene magmatism is extensive (e.g., Metcalf, 2004; Faulds et al., 2001).

The combined field and geophysical data demonstrate that a NNW–trending, 100–300-km-wide belt of ca. 16.7–15 Ma intrusions and volcanic centers extends over a distance >1000 km. From south to north, this belt includes (1) the southern segment of the easternmost aeromagnetic anomaly in southern Nevada, (2) the northern Nevada rift and associated wider belt of aeromagnetic anomalies to the west, (3) the region of 16.5–15.5 Ma rhyolitic calderas at the western end of the Yellowstone–Snake River Plain hotspot track, and farther north into the Oregon-Idaho graben (Lake Owyhee volcanic field) to Castle Rock, Oregon (Streck et al., 2011; Ferns and McClaughry, 2013), (4) the Steens Mountain dike swarm, (5) the Monument dike swarm, and (6) the extensive main-phase dikes of the Chief Joseph swarm that extends into the Columbia Basin of eastern Washington (Figs. 4 and 5). In addition to this axial belt, there are a few outliers of contemporaneous basaltic and/or rhyolitic magmatism exposed over a wider area in north-central Oregon, southernmost Nevada, and northeastern Nevada (Fig. 5A).

Zoback and Thompson (1978) were the first to suggest that the mid-Miocene northern Nevada rift was contemporaneous with the Columbia River basalt dikes and together formed an extensive zone of intercontinental rifting, thus prompting Pierce and Morgan (1992) to refer to this belt as the Oregon-Nevada rift zone. This term leads to a false impression, however, because there is little evidence for significant 17–15 Ma rifting or continental extension associated with the belt. The northern Nevada rift itself has been described as uncharacteristic of most intercontinental rift zones, in that it is dominated by magmatic rocks without significant graben-filling clastic rocks, and with only minor documented extension (John et al., 2000; Colgan, 2013). Because the entire belt is more accurately defined as a zone of magmatic intrusion dominated by shallow dike swarms, midcrustal intrusions, linear vent systems, and localized calderas, we refer to it herein as the "Nevada–Columbia Basin magmatic belt."

PLUME ORIGIN FOR THE YELLOWSTONE HOTSPOT

The origin for the Columbia River flood basalts and the Yellowstone–Snake River Plain hotspot track has been debated for decades (e.g., Smith et al., 1974; Craig et al., 1978; Pierce and Morgan, 1992; Christiansen et al., 2002; Yuan and Dueker, 2005; Waite et al., 2006; Hooper et al., 2007; Fouch, 2012). A plume origin accords well with traditional models that attribute flood basalt volcanism to melting above a mantle plume head and hotspot volcanism to melting above its connecting plume tail (Hill et al., 1992; Campbell and Davies, 2006). Such a model is consistent with the short duration (1.1 m.y.) and high eruption rate (~0.178 km^3/yr) of the main-phase Columbia River flood basalts from 16.7 to 15.6 Ma, and with the well-

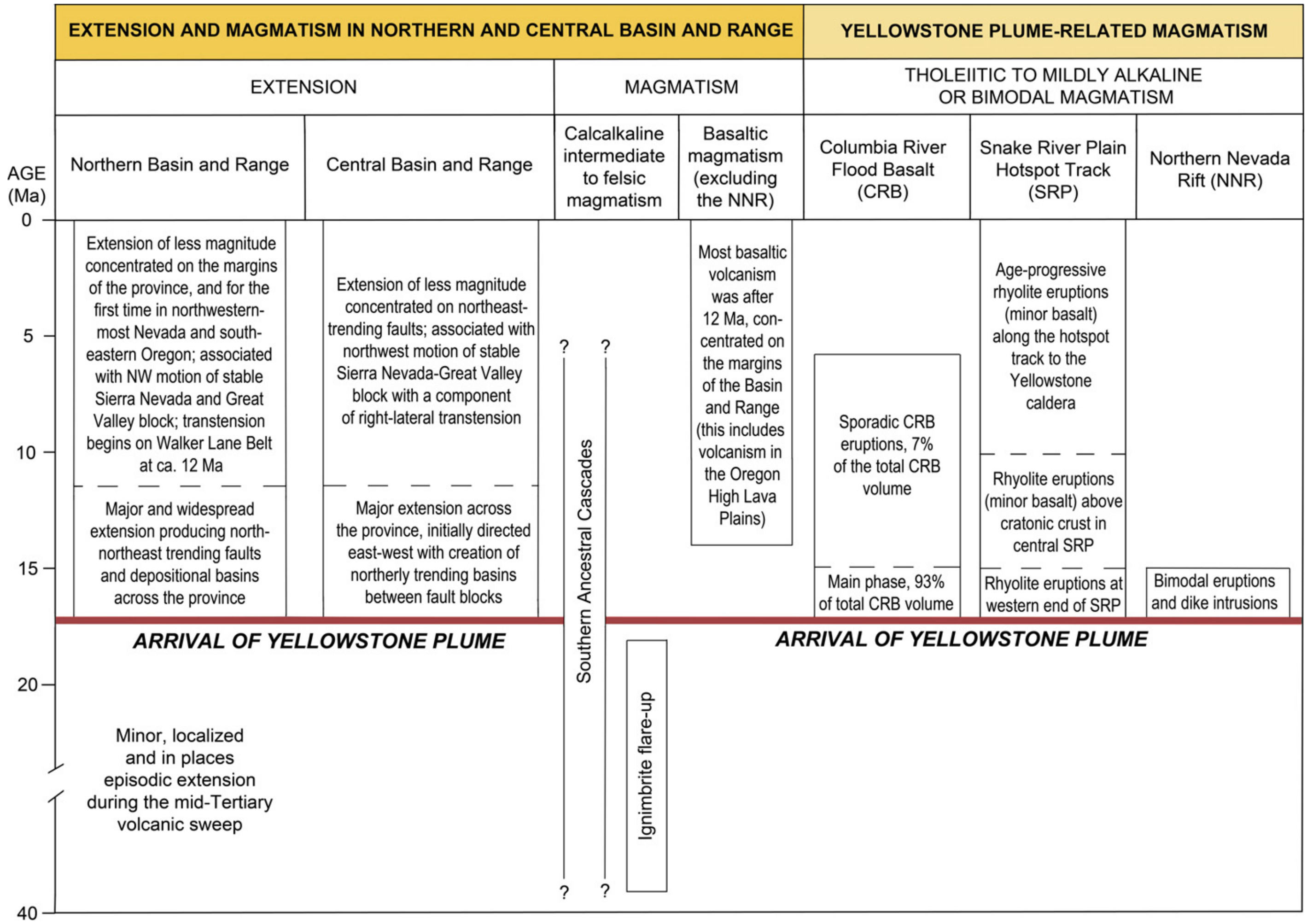

Figure 3. Late Eocene (40 Ma) to recent age correlations for extension and magmatism in the northern and central Basin and Range and plume-related magmatism in the Columbia River flood basalt province (CRB), the Snake River Plain hotspot track (SRP), and the northern Nevada rift (NNR). Not included in this figure is volcanism associated with the Lake Owyhee, Powder River, and Strawberry volcanic fields (Fig. 1; Ferns and McClaughry, 2013). These middle Miocene to Pliocene volcanic rocks are largely calc-alkaline and typically overlie the Columbia River flood basalt stratigraphy, but a few units in the Strawberry volcanics are contemporaneous with the main-phase eruptions of the Columbia River Basalt Group (Steiner and Streck, 2013). See text for more detailed discussion on the relative timing and magnitude of extension and magmatism.

defined northeastward migration of rhyolite volcanism along the Yellowstone–Snake River Plain hotspot track, in the same direction and at the same rate as plate migration, once the plume tail became well established at ca. 12–10 Ma (e.g., Pierce and Morgan, 1992, 2009; Camp and Ross, 2004; Shervais and Hanan, 2008). On the other hand, plume models have a more difficult time explaining the origin of the Oregon High Lava Plain (Fig. 1), which is defined by a similar, but less prominent age migration of rhyolite volcanism from ca. 12 Ma to recent, but in the opposite direction as the Yellowstone–Snake River Plain hotspot track (Ford et al., 2013).

A major objection to a plume origin for the Yellowstone hotspot had been the failure of seismic studies to identify a mantle anomaly beneath Yellowstone to a depth >200 km (e.g., Christiansen et al., 2002). Recent high-resolution tomography, however, now has resolved a deep, S-shaped anomaly described either as a "sheet of upwelling" that extends into the mantle transition zone through a gap in the remnant Farallon plate (Sigloch et al., 2008; James et al., 2011) or as a mantle plume that extends through the transition zone into lower mantle to ~1000 km depth (Obrebski et al., 2010, 2011; Humphreys and Schmandt, 2011; Schmandt et al., 2012; Tian and Zhao, 2012; Darold and Humphreys, 2013).

Camp (2013) listed 12 constraints for genetic models, all of which are consistent with a mantle plume origin, and many of which are inconsistent with nonplume interpretations. In addition to the deep-mantle origin delineated by recent seismic data, the more salient of these constraints are: (1) trace-element and isotopic support for a plume-like component shared by all formations of the main-phase Columbia River Basalt Group (e.g., Hooper and Hawkesworth, 1993; Wolff et al., 2008; Wolff and Ramos, 2013), (2) high ^{3}He/^{4}He values, indicating a lower-mantle source for basaltic rocks on the Snake River Plain and Yellowstone Plateau (Graham et al., 2009), and in the ca. 16 Ma Imnaha Basalt on the Columbia Plateau (Dodson et al., 1997), and (3) the upward deflection of the 660 km mantle discontinuity where it intersects the Yellowstone low-velocity anomaly (Schmandt et al., 2012), as predicted by a thermal plume rising through the mantle transition zone (Bina and Helffrich, 1994).

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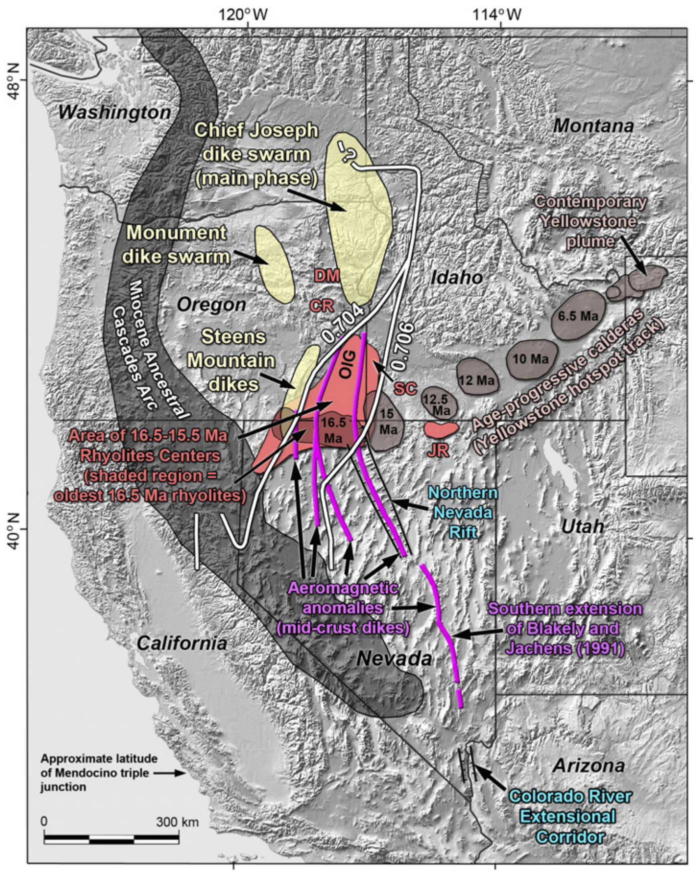

Figure 4. Magmatic features coincident with Basin and Range extension. Basement character is delineated by the 0.704 and 0.706 isopleths of initial $^{87}Sr/^{86}Sr$, modified from Pierce and Morgan (2009) and in southeast Oregon from Shervais and Hanan (2008). These lines of constant Sr-isotope ratios define (1) the boundaries between a thin lithosphere of accreted oceanic terranes north and west of the 0.704 line, (2) Precambrian continental lithosphere south and east of the 0.706 line, and (3) transitional lithosphere between the two. Magmatic features shown were all active during the early stages of mid-Miocene plume emplacement and onset of Basin and Range extension at ca. 16.5 Ma. The area delineated for the Chief Joseph dike swarm includes only the main-phase dikes active from 16.6 to 15.6 Ma; later dikes in the swarm extend farther to the west, east, and north, feeding more sporadic intrusions from ca. 15.6 to 5.5 Ma (e.g., Barry et al., 2013; Reidel et al., 2013a). The aeromagnetic anomalies marking the sites of midcrustal mafic intrusions in Nevada are delineated from Blakely and Jachens (1991) and Glen and Ponce (2002). The oldest rhyolites at 16.5 Ma occur at the western end of the Yellowstone–Snake River Plain hotspot track in the Oregon-Nevada border region. The remaining rhyolites extend as far east as the Jarbidge rhyolite (JR), with $^{40}Ar/^{39}Ar$ ages between 16.1 and 15.0 Ma (Brueseke et al., 2014), and as far north as Dooley Mountain (DM) with a K-Ar age of 14.2 Ma (Evans, 1992). The rhyolites extending northward through the Oregon-Idaho graben (OIG) are younger than 16 Ma (Ferns and McClaughry, 2013), with two possible exceptions based on K/Ar ages. One is a K/Ar age of 16.6 Ma for the Silver City rhyolite (SC; Panze, 1975), although more recent $^{40}Ar/^{39}Ar$ ages are between 16.3 and 15.6 Ma (Hasten et al., 2011). The other is a K/Ar age of 17.3 Ma for a rhyolite lava flow at the base of the Strawberry Volcanics (Fig. 1; Robyn, 1977; Steiner and Streck, 2013). The rhyolite at Castle Rock, Oregon (CR), is the presumed source region for the 15.9–15.4-Ma Dinner Creek tuff (Streck et al., 2011).

The deep-seated nature of the Yellowstone plume suggests that it is a long-lived feature, consistent with the progression of increasingly older rhyolite and basaltic ages along the hotspot track to the southwest (Fig. 4; Pierce and Morgan, 1992, 2009). The exact center of initial impingement is controversial and may be as far north as east-central Oregon (e.g., Glen and Ponce, 2002; Shervais and Hanan, 2008), although others feel that a more reasonable site is in the Oregon-Nevada border region where the oldest basalts (Steens basalt at 16.7 Ma) and the oldest rhyolites (16.5 Ma) are present near the western end of the hotspot track (Fig. 4; e.g., Pierce and Morgan, 1992, 2009; Camp, 2013).

RELATIONSHIP OF MAGMATISM AND EXTENSION

The initiation of Basin and Range extension and the contemporaneous onset of Columbia River flood basalt eruption at 17–16 Ma suggest that they may be genetically related. Coming from the perspective of most workers on the Basin and Range (e.g., Valentine and Perry, 2007; Putirka and Platt, 2012), it may seem reasonable to conclude that magmatism in the northern Nevada rift was the passive response to continental extension, and that a similar scenario can be drawn for coeval Columbia River Basalt volcanism. Dickinson (1997), for example, has suggested that the passive rise of mantle beneath the Columbia River basalt source region was related to the same intracontinental extension event that produced crustal stretching of the Basin and Range, both resulting from broadly based shear stress of the plate interior as a fully integrated transform system was established at 19–17 Ma.

On the other hand, it is difficult to reconcile any model of passive mantle upwelling with the very small amount of extension associated with the flood basalt eruptions. Although ~85% of the flood basalt volume (~175,000 km^3) erupted from the Chief Joseph dike swarm (Fig. 4), the magnitude of extension in this region is <<1%, only enough to accommodate the width of the combined dikes (Taubeneck, 1970). In the mid-Miocene, the Chief Joseph dike swarm represents the area of least regional strain and greatest eruptive volume, but the Basin and Range represents the area of greatest regional strain with only minor volcanic products. To the first order, this observation is inconsistent with magmatism driven by the passive rise of mantle into an extended lithosphere (White and McKenzie, 1989). We concur with most workers on the Columbia River flood basalts (e.g., Hooper, 1990; Camp and Ross, 2004; Hooper et al., 2007; Camp, 2013; Reidel et al., 2013b), who propose that magmatism did not result from the passive rise of shallow mantle, but instead from active upwelling and adiabatic melting of a deep-mantle source. We believe that there are clear exceptions, however, along the periphery of Basin and Range extension in the Mojave Desert and along the Walker Lane belt of western Nevada and California (McQuarrie and Oskin, 2010; Putirka and Platt, 2012; Busby et al., 2013a).

The initiation of large-scale Basin and Range extension at 17–16 Ma was contemporaneous with three large-scale magmatic events: (1) early volcanism in the incipient northern Nevada rift and related midcrustal intrusions at 16.5 Ma, (2) initial rhyolite volcanism at the southwestern end of the Yellowstone–Snake River Plain hotspot track at 16.5 Ma, and (3) the

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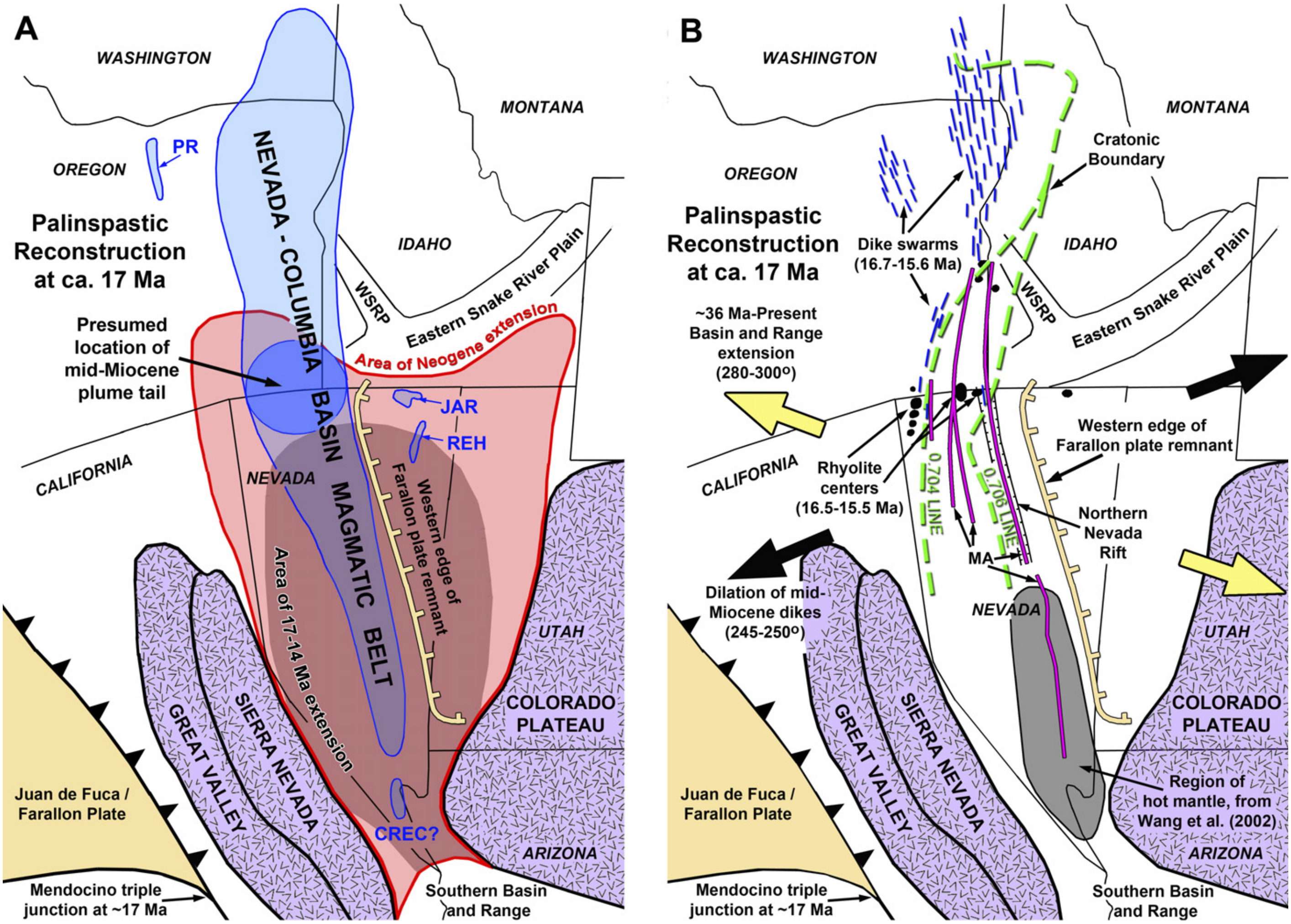

Figure 5. Mid-Miocene magmatism and extension in the western United States palinspastically reconstructed to ca. 17 Ma. Reconstruction includes eastward closure of ~250–300 km of the Sierra Nevada–Great Valley block relative to the Colorado Plateau (McQuarrie and Wernicke, 2005). State borders and the approximate location of the Mendocino triple junction are from Atwater and Stock (1998). WSRP—western Snake River Plain. Random "V" pattern—stable continental blocks separated by mid-Miocene Basin and Range extension. Magmatism along the Nevada–Columbia Basin magmatic belt is bounded on the east by an eastward-dipping fragment of the Farallon plate beneath eastern Nevada and western Utah at a depth between 100 and 600 km, as reconstructed from high-resolution seismic imagery (Burdick et al., 2008; Schmandt and Humphreys, 2010; Sigloch, 2011; Tian et al., 2011; Liu and Stegman, 2012; Tian and Zhao, 2012). We suggest that this feature was torn away from the subducting slab along a propagating rupture lying above the ascending Yellowstone plume at ca. 17 Ma. Mantle upwelling through the plate resulted in the rapid emplacement of magmatic intrusions, vigorous traction of the asthenosphere, and advective heating and thermal weakening of the lithosphere, all of which aided in the initiation of contemporaneous mid-Miocene extension. (A) Location of the ca. 16.7–15 Ma Nevada–Columbia Basin magmatic belt and its relationship to the area of 17–14 Ma Basin and Range extension (modified after Colgan and Henry, 2009) and the larger area of Neogene extension. Contemporaneous magmatism is present in outliers of the magmatic belt that include the source region for ca. 15.8 Ma Prineville Basalt (PR) in north-central Oregon (Reidel et al., 2013b), the ca. 16.1–15 Ma Jarbidge rhyolite (JAR) and nearby 16.5 Ma Seventy-Six Basalt in northeastern Nevada (Rahl et al., 2002; Brueseke et al., 2014), and mid-Miocene dikes in the Ruby and East Humboldt Mountains (REH) together with adjacent basaltic lavas in the ca. 16–10 Ma Humboldt Formation of northeastern Nevada (Hudec, 1990; Colgan, 2013). A more speculative outlier to the south includes mid-Miocene plutonic and volcanic rocks in the Colorado River extensional corridor (CREC; e.g., Anderson et al., 1994; Faulds et al., 2001; Metcalf, 2004). (B) Magmatism along the ca. 16.7–15 Ma Nevada–Columbia Basin magmatic belt includes midcrustal intrusions delineated by aeromagnetic anomalies (MA) of Blakely and Jachens (1991) and Glen and Ponce (2002), ca. 16.5–15 Ma bimodal volcanic rocks along the northern Nevada "rift," ca. 16.5–15.5 Ma rhyolite centers at the western end of the Snake River Plain hotspot track and in the Lake Owyhee volcanic field west of the western Snake River Plain, and the ca. 16.7–15.6 Ma dike swarms associated with the main phase of Columbia River flood basalt volcanism. The southern portion of the Nevada–Columbia Basin magmatic belt corresponds with the N-S region of hot mantle upwelling interpreted from young (<10 Ma) basalts by Wang et al. (2002). The southern half of the belt transects the region of known ca. 17–14 Ma Basin and Range extension. Large yellow arrows denote the long-lived orientation of extensional stress from ca. 36 Ma to present; large black arrows denote the short-lived orientation of dike dilation associated with emplacement of the Nevada–Columbia Basin magmatic belt from ca. 16.5 to 15 Ma.

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initial eruptions of the Columbia River Basalt Group (Steens Basalt) in southeastern Oregon at 16.7 Ma (Fig. 4). All three of these magmatic events began near the Oregon-Nevada border region, in the same general area where plate-motion studies and mantle flow rate estimates place the seismically defined Yellowstone mantle plume at 16.5 Ma (Smith et al., 2009; Camp, 2013). If the newly resolved plume (Obrebski et al., 2010, 2011; Schmandt et al., 2012; Tian and Zhao, 2012) arrived beneath the tristate region at ca. 17–16 Ma, then logically it should have played a major role in the initiation of volcanism and extension of the same age.

Regional Middle Miocene Stress Regime

Based on the consistent orientation of the northern Nevada rift and related dikes, Zoback et al. (1981, 1994) suggested that the least principal stress direction across the Basin and Range province was 245°–250° from ca. 17 to 10 Ma. They further suggested that a change in plate boundary conditions after 10 Ma rotated the extension direction by ~45° to 290°–300°, thus producing the northeast trend that defines the current Basin and Range structure and topography. This long-held view, however, has come into question as new data have become available from a variety of recent structural, stratigraphic, and thermochronologic investigations in northern Nevada, summarized in Colgan (2013). These studies show that the region surrounding the northern Nevada rift was actively extending as the rift formed, generating consistent NE-trending structures that reflect an extension direction of 280°–300° (Colgan, 2013). Several studies show that this stress orientation was present before, during, and after the abrupt onset of widespread crustal stretching that began at ca. 17 Ma (Wernicke and Snow, 1998; McQuarrie and Wernicke, 2005; Colgan, 2013).

The magnitude of regional middle Miocene strain associated with the NE-trending structures decreases from south to north, with the greatest percentage of strain in the central Basin and Range (200% since 17 Ma; McQuarrie and Wernicke, 2005), decreasing northward into north-central Nevada (40% from 17 to 10 Ma; Colgan and Henry, 2009), with little or no strain in the Nevada-Oregon border region and farther north (Taubeneck, 1970; Colgan et al., 2004, 2006; Scarberry et al., 2010). The paucity of strain farther north is consistent with recent global positioning system (GPS) data, demonstrating a clockwise rotation of western Oregon and Washington, with low rates of strain and random stress directions in eastern Oregon and Washington (McCaffrey et al., 2007, 2013; Payne et al., 2012). The contemporary rotation rates are similar to those inferred from paleomagnetic declination anomalies in the Columbia River flood basalts (e.g., Magill et al., 1982; Sheriff, 1984; Wells and Heller, 1988; Wells and McCaffrey, 2013), which suggest that the deficiency of tectonic stress in eastern Oregon and Washington has existed at least since the middle Miocene.

Localized Mid-Miocene Stress Associated with the Nevada–Columbia Basin Magmatic Belt

The prevailing NNE structural trends that define the mid-Miocene regional stress field appear to be contemporaneous with NNW magmatic trends that define the Nevada–Columbia Basin magmatic belt (Fig. 5). The lone exception to this magmatic orientation is in the Oregon-Nevada border region, where the strike of the large, aeromagnetically defined, midcrustal keel dikes and the Steens Mountain feeder dikes change to a NNE orientation. Although consistent with the regional stress orientation (Fig. 5), this broad zone of dike emplacement could have also been facilitated by the intrusion of magma into preexisting structures associated with the thrusted transition zone between the Sr 0.704 and 0.706 isopleth lines (Fig. 4; Reidel et al., 2013b; Camp et al., 2014).

Ponce and Glen (2008) used geophysical data to suggest that the location and trend of the northern Nevada “rift” itself might have been controlled by a preexisting structure. Colgan (2013), however, pointed out that dikes along the “rift” do not appear to intrude preexisting faults. Such structures may have provided local control, but it is difficult to argue that they were the primary control on dike orientation along the entire length of the Nevada–Columbia Basin magmatic belt.

Although mid-Miocene extension in the northern and central Basin and Range was considerable, there is a remarkably small amount of extension associated with the Nevada–Columbia Basin magmatic belt. Colgan (2013) estimated that the northern Nevada “rift” accommodated <0.5% of total extension across the northern Basin and Range, and there is no recorded extension associated with the Chief Joseph and Monument dike swarms. We believe that the high concentration of dikes in these areas of little or no extension is inconsistent with magma rise into extension-related fractures and more consistent with a mechanism of forceful dike injection (Rubin, 1995; see Magmatic Stress section).

The combined observations suggest that the interior North American plate was subjected to two partly contemporaneous stress regimes in the middle Miocene (Fig. 5): (1) a regional, largely amagmatic stress field (excluding Walker Lane) with maximum extension at ~280°–300° responsible for nearly all the gravitationally driven crustal stretching across the central and northern Basin and Range, but with low magmatic output, and with a diminishing magnitude of extension to the north, and (2) a localized magmatic stress field driven from the bottom up, with a 245°–250° orientation of very minor compressional stress associated with dike dilation (Rubin, 1995), but with a general increase in magmatic output to the north. The regional stress regime appears to be long-lived, from ca. 36 Ma to the present, but bottom-up magmatic stress was short-lived (mostly from 16.7 to 15 Ma) and specific to the abrupt arrival of basaltic to bimodal intrusions of the Nevada–Columbia Basin magmatic belt. Magmatism along the belt appears to be the result of active mantle upwelling, with its orientation controlled by a fundamentally deeper process in the sublithospheric mantle.

Magmatic Stress

The relationship among mantle upwelling, continental extension, and dike emplacement has been evaluated in a variety of analytical models. Such studies demonstrate that (1) lithospheric heating by mantle upwelling and related magma production can promote extensional stresses that are significantly lower than those associated with areas of amagmatic stretching and normal faulting (Buck, 2004), and (2) the stress needed to open dikes can be substantially less than that needed for tectonically controlled normal faulting (Buck, 2006; Qin and Buck, 2008).

Dike propagation is partly dependent on magma overpressure, which is the difference between total magma pressure and the lithostatic stress (Gudmundsson, 2006). If magma overpressure is high enough, dikes can produce their own fractures as they rise by forceful injection (Rubin, 1995; Traversa et al., 2010), with inelastic deformation in the near-tip stress field forming closely spaced dike-parallel joints (Delaney et al., 1986), breccia (Johnson and Pollard, 1973), and/or faulted and folded strata (Pollard et al., 1975); however, the majority of the deformation is elastic due to dilation perpendicular to the dike plane (Fig. 6; Rubin, 1995). Magmatic stress provides an alternative explanation for the northern Nevada “rift” in that it might represent the surface manifestation of forceful dike injection and near-tip deformation, with an ultimate source associated with the large, geophysically defined mafic intrusion lying beneath the magmatic lineament. Seismic data suggest that these large intrusions extend downward at least into the ductile lower crust (Potter

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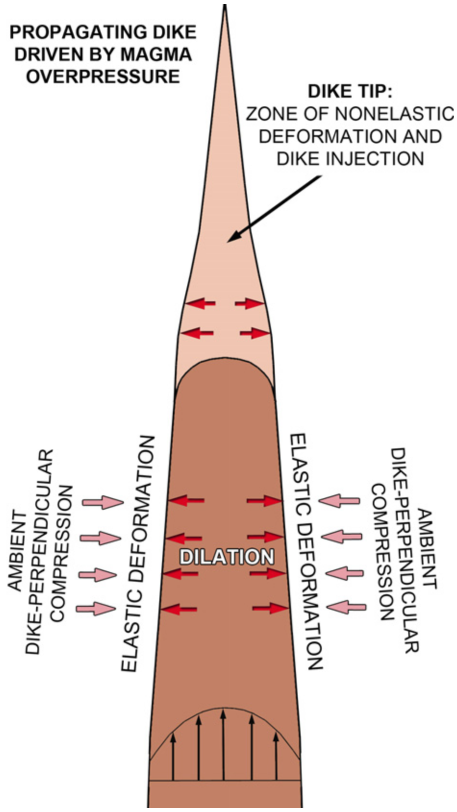

Figure 6. Propagation of a dike with excess magmatic pressure, modified from Rubin (1995). Deformation of the host rock during dike ascent is largely elastic where internal pressure generates ambient compressional stress perpendicular to the dike plane. Inelastic deformation is concentrated in the dike tip, where dikes widen their channels during ascent and/or where lateral transport occurs. Here, deformation is manifested in breccia, closely spaced dike-parallel joints, and/or folded strata. Length of the dike tip is typically ~0.01 times the length of the dike (Rubin, 1995).

et al., 1987), consistent with the establishment of their NNW trends at a depth unaffected by regional stress in the brittle upper crust.

Other studies demonstrate that vertical sheetlike bodies of ascending magma will rise through the lower crust by ductile flow, but as they enter the brittle upper crust, they tend to reorient to the regional stress regime, with the strike of the dikes aligning perpendicular to the minimum principal stress axis (σ_3; e.g., Lister and Kerr, 1991; Parsons and Thompson, 1991). Whether or not they succeed depends on a variety of factors, including dike propagation rate, distance traveled, magma overpressure, and magma viscosity (Emerman and Marrett, 1990; Mériaux and Lister, 2002). High propagation rates can be obtained from low-viscosity basaltic magma with high overpressures. For example, high magma overpressures can be expected near shallow-level magma chambers, where basaltic dikes emanate outward in radial patterns that lie oblique to both regional stress axes and preexisting structures (e.g., Mériaux and Lister, 2002). One such example is at Spanish Peaks, Colorado, where dike orientation remains radial near central intrusive bodies, but at greater distances where magma overpressure is reduced, the dikes curve to more uniform orientations controlled by the regional stress (Muller and Pollard, 1977). On the other hand, Mériaux and Lister (2002) showed that dikes that maintain high magma overpressure will preserve orientations that are little affected by the regional stress field. Emerman and Marrett (1990) noted that dike propagation rate is also important because dikes that do not reorient themselves quickly enough will have orientations that might be only weakly related to the state of regional stress.

Magma flux rate, magma overpressure, and the dike propagation rate will be considerably higher in the extensive dike swarms associated with flood basalt volcanism, where 10^5–10^7 km^3 of basalt can erupt over time spans of 1–3 m.y. (e.g., Bryan and Ferrari, 2013). During the peak eruption of Columbia River flood basalt, high magma flux and overpressure are implied by the high average volume of individual flows (~1350 km^3) generated in only a few months to a few years (Reidel et al., 2013a). We suggest that this condition was prevalent during rapid dike emplacement along the entire length of the Nevada–Columbia Basin magmatic belt, but with only minor surface expression above cratonic lithosphere in central Nevada.

Walker (1993) noted that under conditions of high thermal-energy supply rate, magma pathways may become established between the mantle source and the surface, permitting long-sustained eruptions typical of flood basalt provinces. The Columbia River flood basalts, however, appear to have been modified to variable degrees in magma chambers beneath the magmatic belt (Carlson et al., 1981; Carlson, 1984; Caprarelli and Reidel, 2004; Durand and Sen, 2004; Ramos et al., 2013; Wolff and Ramos, 2013). Thermomechanical modeling, combined with the evaluation of cumulate bodies underlying large igneous provinces, suggests that the modification of flood basalt provinces takes place in large sill-like structures near the Moho, where olivine and pyroxene begin to crystallize, with further crystallization of plagioclase in dikes and sills that develop at various crustal levels (Cox, 1980; Karlstrom and Richards, 2011). Whereas diking in large felsic magma chambers in the upper crust is inhibited by prolonged heating of country rock, leading to ductile deformation that retards overpressurization (e.g., Jellinek and DePaolo, 2003; Gregg et al., 2013), the magma flux of flood basalt eruptions is high enough to induce elastic deformation and dike ascent (Karlstrom and Richards, 2011). We suggest that the orientation of the flood basalt reservoirs was acquired at sub-crustal depth, with dikes maintaining high overpressures that dominated or overprinted the shallow stress regime. The high effusion rate of the lavas is consistent with open-system processes, where the rate of eruption was largely balanced by the constant influx of new magma, thus maintaining high magma flux as modified magmas continued to ascend to the surface.

The mid-Miocene arrival of the Yellowstone plume was accompanied by a massive but short-lived magmatic event, from ca. 16.7 to 15 Ma, concentrated along a north-south zone coincident with dike injection. This abrupt intrusive event was driven by a bottom-up process into an upper crust that was under a regional extensional stress, but with little strain before 17–16 Ma. We suggest that a substantial magma flux allowed dikes to maintain the same orientation acquired during their ascent through the mantle lithosphere and lower crust (dike dilation at 245°–250°). Rapid dike emplacement beneath the Nevada–Columbia Basin magmatic belt was therefore largely unaffected by preexisting structures and regional stress (extension at ~280°–300°), with the latter diminishing rapidly into eastern Oregon and Washington (McQuarrie and Wernicke, 2005; Payne et al., 2012; McCaffrey et al., 2013).

GEODYNAMIC MODEL

Initiation of the Yellowstone hotspot has been attributed to the ascent of a plume head that first impinged on the lithosphere in the mid-Miocene (e.g., Pierce and Morgan, 1992, 2009; Pierce et al., 2002; Camp and Ross, 2004; Shervais and Hanan, 2008). Others workers have suggested instead that the Yellowstone hotspot is a long-lived feature that resided offshore to produce the basaltic Siletzia terrane, a chain of oceanic islands that was obducted onto the continental margin at 50–45 Ma (Duncan, 1982; Wells et al., 1984, 2014; Murphy et al., 1998; McCrory and Wilson, 2013). Under this scenario, the plume source (i.e., tail) would be trapped beneath the Farallon plate before reappearing at ca. 17 Ma to produce the Columbia River flood basalts and the Yellowstone–Snake River Plain hotspot track (Duncan, 1982). A starting plume head is therefore unnecessary since the initial high-volume flood basalts could be attributed to a feeding plume tail that was able to accumulate

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significant mass beneath the subducting slab from ca. 45 to 17 Ma (Geist and Richards, 1993; Obrebski et al., 2010). Recent plate-motion studies show that a long-lived Yellowstone hotspot is a viable model consistent with the timing of Siletzia accretion and re-emergence of the hotspot near the source region of mid-Miocene volcanism considered here (McCrory and Wilson, 2013; Wells et al., 2014). Either model is possible, but for the purposes of this paper, we will use the term "plume head" to describe the accumulation of hot mantle that first arrived beneath the Nevada–Columbia Basin magmatic belt at 17–16 Ma.

Several seismic studies have identified significant gaps or tears in the Farallon–Juan de Fuca plate that have been attributed to different processes. Xue and Allen (2007), Obrebski et al. (2010), and Coble and Mahood (2012) suggested that tearing was the result of slab interaction with the rising Yellowstone plume, consistent with the rate of subduction and the depth of slab termination. In a plume rupture scenario, plume material would rise through the ruptured slab, where it would spread largely to the north beneath the thinner oceanic lithosphere of the accreted terranes lying west of the thicker Precambrian craton in Idaho (Fig. 5).

An alternative explanation that is more consistent with linearity of magmatism along the entire length of the Nevada–Columbia Basin magmatic belt comes from the model simulations of Liu and Stegman (2012). These simulations are based partly on tomographic models that have identified a quasi-linear N-S slab gap in the Farallon plate at ~650 km beneath eastern Utah at longitude ~110°W, with a broken eastern remnant of the Farallon slab beneath Colorado at a depth of 700–800 km (Fig. 7A; Burdick et al., 2008; Schmandt and Humphreys, 2010; Tian et al., 2011; Tian and Zhao, 2012; Sigloch, 2011). Their model simulations produce a mid-slab tear beneath southeastern Oregon at around 17 Ma that propagates rapidly to the north and south from 17 to 15 Ma. By ca. 15 Ma, the tear extends over a linear NNW trend of 900 km, largely coincident with the area overlain by the Nevada–Columbia Basin magmatic belt. The initial tear in their simulations is attributed to dynamic pressure at a hinge in the subducting slab, and therefore would not require slab interaction with the Yellowstone mantle plume (Fig. 7B). On the other hand, location of this initial rupture is in the same broad area where several workers place the Yellowstone plume at ca. 16.5 Ma (Pierce and Moragn, 1992, 2009; Pierce et al., 2002; Camp and Ross, 2004; Smith et al., 2009).

There are both positive and negative attributes to the plume-rupture and hinge-rupture models. Although there is strong geological and seismic evidence for plume arrival at ca. 17 Ma, it is difficult to reconcile plume-triggered rupture of a strong, coherent slab with the weakness of mantle plumes, which have exceedingly low viscosities (~100 times lower than upper mantle; Stegman et al., 2006) and low yield strengths (Betts et al., 2012). In comparison, the hinge-rupture model of Liu and Stegman (2012) appears to require an early Tertiary period of flat subduction west of the initial hinge tear beneath southeastern Oregon (Fig. 7B), but geologic evidence suggests that flat subduction in southeast Oregon and northern Nevada ceased at ca. 35 Ma during the early Tertiary volcanic sweep (Fig. 2). This volcanic migration has been widely attributed to a period of slab rollback (e.g., Coney and Reynolds, 1977; Severinghaus and Atwater, 1990; McQuarrie and Oskin, 2010; Dickinson, 2013), which may have begun during the accretion of the Siletzia terrane in Washington and Oregon at ca. 50 Ma (Schmandt and Humphreys, 2011). This, in turn, is consistent with the westward migration

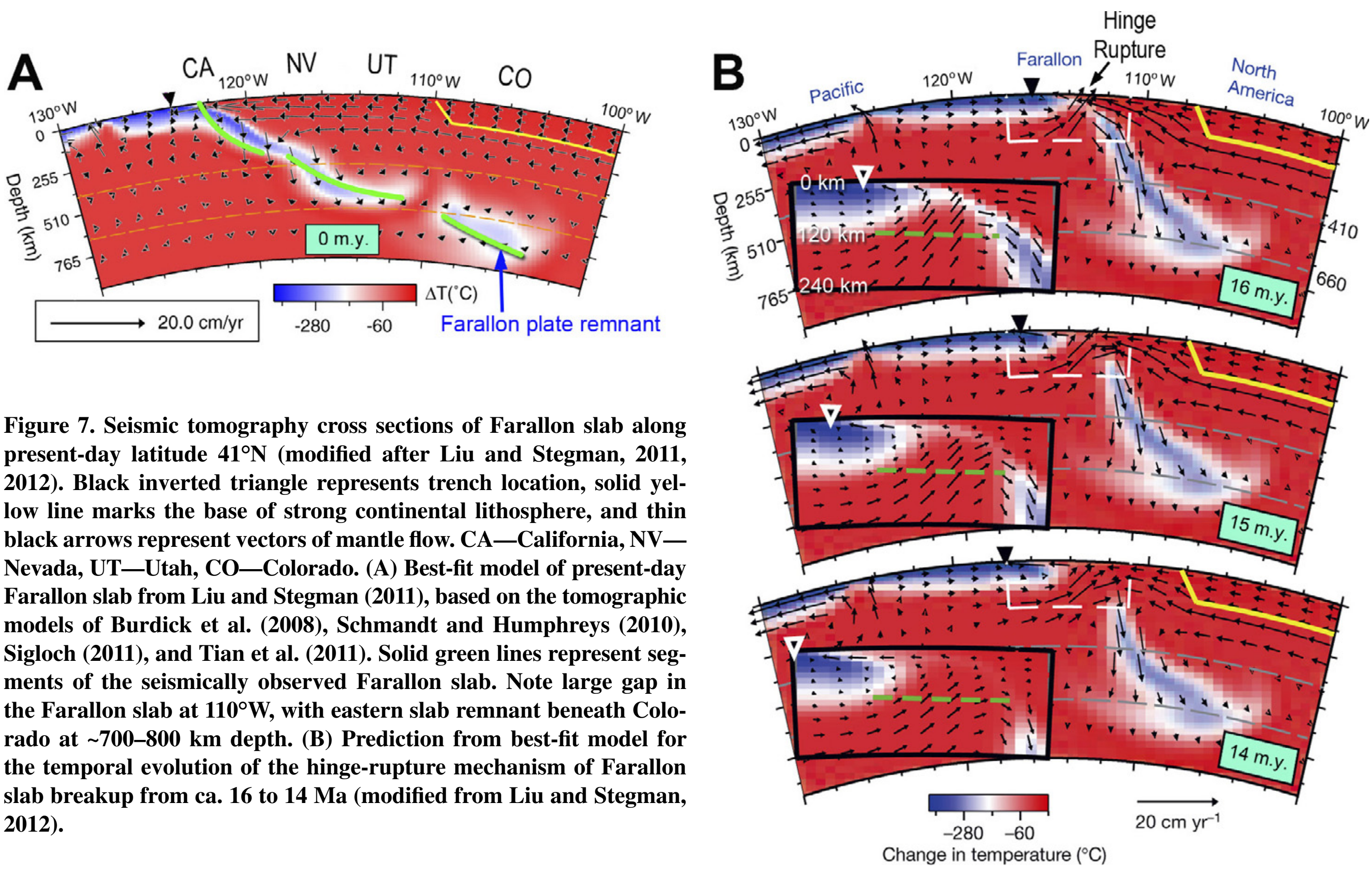

Figure 7. Seismic tomography cross sections of Farallon slab along present-day latitude 41°N (modified after Liu and Stegman, 2011, 2012). Black inverted triangle represents trench location, solid yellow line marks the base of strong continental lithosphere, and thin black arrows represent vectors of mantle flow. CA—California, NV—Nevada, UT—Utah, CO—Colorado. (A) Best-fit model of present-day Farallon slab from Liu and Stegman (2011), based on the tomographic models of Burdick et al. (2008), Schmandt and Humphreys (2010), Sigloch (2011), and Tian et al. (2011). Solid green lines represent segments of the seismically observed Farallon slab. Note large gap in the Farallon slab at 110°W, with eastern slab remnant beneath Colorado at ~700–800 km depth. (B) Prediction from best-fit model for the temporal evolution of the hinge-rupture mechanism of Farallon slab breakup from ca. 16 to 14 Ma (modified from Liu and Stegman, 2012).

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of volcanism associated with the ancestral southern Cascades arc in the California-Nevada border region (Fig. 4; du Bray et al., 2013).

We believe that the field, seismic, and geochemical data can be more reasonably explained in a model that combines, in part, the conclusions of Xue and Allen (2007), Obrebski et al. (2010), Coble and Mahood (2012), and Liu and Stegman (2012). Although plumes may lack the mechanical strength necessary to rupture a coherent slab, their thermal energy and high magma flux have the ability to uplift slabs while simultaneously decreasing slab strength by thermal diffusion (e.g., Macera et al., 2008; Betts et al., 2012). Coble and Mahood (2012) noted that early Tertiary rollback of the Farallon slab may have been followed by plume arrival and rapid uplift of the slab in the mid-Tertiary. They supported this argument by noting a hiatus in back-arc volcanism that began at ca. 23 Ma and ended with the outbreak of flood basalt volcanism at ca. 16.7 Ma, presumably from the cessation of corner flow resulting from uplift. Such a model would likely generate an abrupt downward bend in the slab on the eastern side of plume uplift (e.g., Coble and Mahood, 2012, their fig. 3D), similar to the hinge proposed by Liu and Stegman (2012), but weakened further by thermal diffusion from plume impingement. Rupturing of this weakened hinge, and rapid propagation of slab tearing to the north and south would then proceed in accordance with the model simulations of Liu and Stegman (2012), but it would be accompanied by the rapid rise of buoyant plume material along the length of slab tearing.

Whatever the cause of rupture, the Nevada–Columbia Basin magmatic belt appears to be associated with the adiabatic rise of hot mantle forced into a more-or-less tabular shape as it was squeezed through the torn slab. The deformed plume head propagated northward beneath thin lithosphere of accreted oceanic terranes, but it was also emplaced beneath thicker lithosphere of the Precambrian craton (Fig. 8). Its impingement beneath the North American plate was associated with the abrupt onset of partial melting and the initiation of massive dike injection into the lithospheric mantle.

Although the orientation of magmatism was controlled by sublithospheric processes, the lithospheric mantle and crust played a significant role in controlling the style of near-surface magmatism, extension, and uplift. Where the Nevada–Columbia Basin magmatic belt intruded into Precambrian lithosphere containing continental and/or transitional crust (i.e., south and east of the 0.704 Sr isopleth), the large, aeromagnetically defined intrusions remained in the mid-to-lower crust, with a few smaller dikes extending through the upper crust to feed sparse, small-volume eruptions in the northern Nevada "rift"; however, where the belt crossed into the thinner oceanic lithosphere (west and north of the 0.704 line; Fig. 4), numerous mafic dikes ascended rapidly to the surface to feed the voluminous, main-phase eruptions of the Columbia River flood basalts (Fig. 8; Reidel et al., 2013a).

The significantly greater volume of volcanism north of the 0.704 line can be attributed to two main factors. (1) Buoyant upwelling mantle will always spread preferentially "uphill" into thinner lithosphere (Thompson and Gibson, 1991; Sleep, 1997; Ebinger and Sleep, 1998), consistent with the Yellowstone plume spreading preferentially beneath the thinner lithosphere lying north of the 0.704 line. Melting at this shallower depth should have generated a significantly greater volume of magma by adiabatic decompressional melting, than would be expected beneath the thicker lithosphere of the northern Basin and Range, where a smaller volume of magma would be generated at greater depths (White, 1993). (2) Glazner and Ussler (1989) noted that subtle variations in the density difference between rising basalt magma and the crust will determine the level of neutral buoyancy and the height at which the column of magma may rise by hydrostatic pressure. The higher density of the thin oceanic crust north of the 0.704 line therefore favors the eruption of basaltic magmas, in contrast to the thicker, lighter crust beneath the northern Basin and Range to the south (Fig. 8).

The Nevada–Columbia Basin magmatic belt represents the short-lived axial expression of surface volcanism from a deformed plume head that may have spread laterally before becoming attached to the North American plate in the mid-Miocene, and eventually detached from the Yellowstone plume tail by plate motion. Pierce and Morgan (2009) showed that this detachment occurred over a transition interval that began at ca. 14 Ma, with complete separation producing a well-defined hotspot track after 10 Ma. The difference between the lack of significant volcanism in the Great Basin and abundant volcanism above similar Precambrian lithosphere in the Snake River Plain is attributed to the constant flux of the Yellowstone plume tail, which is thought to be ~200 °C hotter than mantle at the periphery of the plume spreading (e.g., Campbell and Davies, 2006).

MANTLE SOURCE COMPONENTS AND EVIDENCE FOR HOT MANTLE UPWELLING AT 17–16 MA

The main-phase eruptions of the Columbia River Basalt Group, from 16.7 to 15.6 Ma, comprise the Steens, Imnaha, and Grand Ronde Basalts, all of which share a common isotopic component described by several workers as a plume source (Brandon and Goles, 1988, 1995; Hooper and Hawkesworth, 1993; Camp and Hanan, 2008; Wolff et al., 2008; Wolff and Ramos, 2013). The least diluted of this plume component is Imnaha Basalt, which has trace-element signatures of oceanic-island basalt (OIB), similar to those in Hawaii and Iceland (Hooper and Hawkesworth, 1993; Bryce and DePaolo, 2004), and moderately high $^{3}He/^{4}He$ ratios, as one would expect from a deep mantle source (Dodson et al., 1997).

Major-element, trace-element, and He-isotope systematics for Snake River Plain basalts are also consistent with a deep-mantle source (e.g., Craig et al., 1978; Hughes et al., 2002; Graham et al., 2009), but their Sr and Nd isotope ratios appear to be intermediate between depleted mantle and continental crust or lithospheric values (Leeman et al., 1985; Hughes et al., 2002). To address this apparent discrepancy, Hanan et al. (2008) and Jean et al. (2014) used Pb, Sr, and Nd data to demonstrate that the isotopic ratios in Snake River Plain basalt are consistent with a plume source that has been modified by interaction with ancient Precambrian lithosphere underlying the hotspot track.

The source characteristics of younger than 17-Ma volcanic rocks from the northern and central Basin and Range differ in being highly variable and not as readily defined. Ignoring the arc-related calc-alkaline rocks along the California-Nevada border (du Bray et al., 2013), studies on alkalic and tholeiitic basalts in the rest of the province have focused on variations in depth of melting, with several workers suggesting that they were derived from one or a combination of lithospheric and/or normal-temperature asthenospheric sources (e.g., Farmer et al., 1989; Bradshaw et al., 1993; Rogers et al., 1995; DePaolo and Daley, 2000). Only small numbers of $^{3}He/^{4}He$ analyses have been published on Basin and Range basalts (Reid and Graham, 1996; Dodson et al., 1998), and these analyses appear to be more consistent with a source from the subcontinental lithosphere. This does not preclude, however, the existence of a mantle plume, which could have provided the thermal energy necessary to melt previously metasomatized lithospheric mantle (e.g., McKenzie, 1989). Other workers have noted that Basin and Range alkali basalts have OIB chemical signatures that are consistent with a mantle plume origin (e.g., Fitton et al., 1991; Feuerbach et al., 1993; Wang et al., 2002).

Few of the Basin and Range basalts meet the chemical requirements of a primary magma. Unlike the Columbia River flood basalts that erupted above oceanic lithosphere, Basin and Range basalts overlie Precambrian continental

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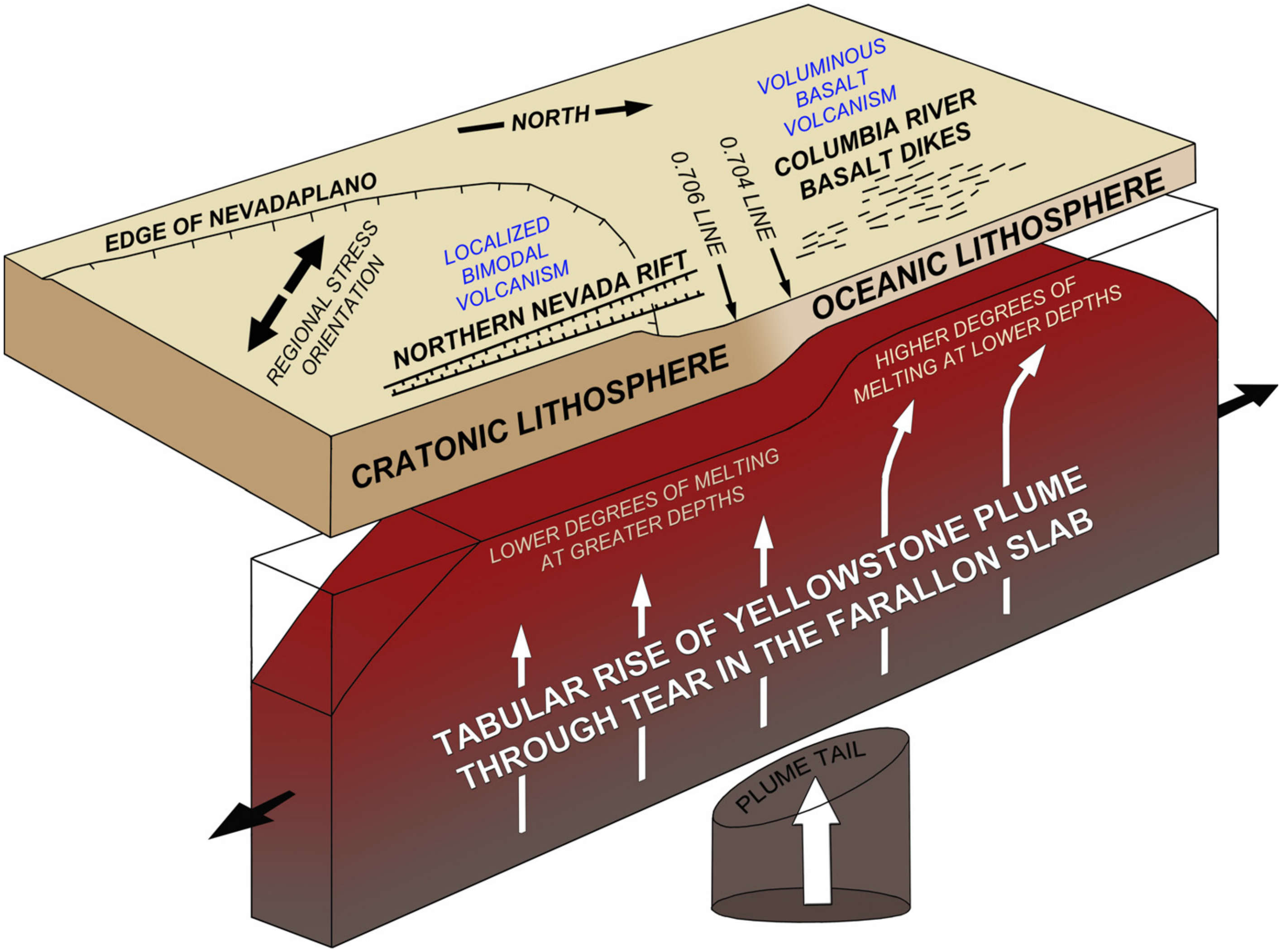

Figure 8. Effect of a tabular body of plume material emplacement beneath cratonic and oceanic lithosphere at ca. 16 Ma. The diagram is aligned parallel to the Nevada–Columbia Basin magmatic belt. The ascending tabular mass of plume material preferentially rises beneath the thinner oceanic lithosphere of accreted terranes north of the 0.704 Sr line. Melting at this shallower depth would have led to a higher degree of partial melting and more voluminous basaltic magma than would be generated beneath the thicker cratonic lithosphere to south and east of the 0.704 line. The relatively high density of thin oceanic crust was favorable to the ascent of basalt magmas, thus generating voluminous flood basalt volcanism. The thicker, lower-density cratonic crust south of the 0.706 line was unfavorable to the rise of basaltic magmas (Glazner and Ussler, 1989), thus generating a much smaller volume of bimodal volcanic products and midcrustal intrusions. The NNW direction of dike intrusion along the length of the Nevada–Columbia Basin magmatic belt was not controlled by regional stress, but by magmatic stress above the tabular zone of mantle melting. Regional stress south of the 0.706 line is the result of high gravitational potential energy of the Nevadaplano and boundary forces associated with Pacific–North American plate interaction. Plume emplacement resulted in thermal weakening of the cratonic lithosphere, added buoyancy, and mantle tractional forces that were superimposed upon the regional stress regime, thus providing the catalyst for Basin and Range extension. These plume-related forces were also in effect north of the 0.706 line, in an area of relatively low gravitational potential energy and far removed from plate boundary forces. Thus, there was little or no extension associated with the northern part of the Nevada–Columbia Basin magmatic belt between ca. 17 and 12 Ma; however, the later period of Basin and Range extension would generate fault-block structures in northwestern Nevada and southeastern Oregon beginning ca. 12–10 Ma.

crust and show considerable evidence of being modified by crustal contamination and/or magma mixing with continental crustal melts (Glazner et al., 1991; Glazner and Farmer, 1992; Farmer et al., 1995). Such hybrid magmas will inherit trace-element and isotopic ratios from older, more-evolved crust, thus making it more difficult to identify the original mantle source component. In this sense, many of these basalts are similar to Snake River Plain basalts, which show similar trace-element and isotopic values (Wang et al., 2002; Hanan et al., 2008; Jean et al., 2014).

Wang et al. (2002) took a different approach by creating a mantle melting profile across the central and northern Basin and Range, using ~1000 new and published analyses on basalts younger than 10 Ma. Their profile is based on a quantitative assessment of major-element and rare earth data, using a melting model to determine the depth range and degree of partial melting. They found that the shape of the melting region across the Basin and Range mimics the shape of the asthenosphere-lithosphere boundary, with nearly all melts derived from the underlying asthenosphere. Their data demonstrate that melting depths beneath the western Great Basin (eastern Sierra Nevada–Mojave) are shallow (50–75 km), with basalt generated by adiabatic melting of normal-temperature asthenosphere; however, greater melting depths are evident beneath the northern and central Basin and Range (100–140 km), where a thicker mantle lithosphere requires melts to be produced by mantle potential temperatures that are 230 °C hotter than normal-temperature mantle that exists in the western Great Basin. Wang et al. (2002) concluded that deep melting and high mantle temperatures are consistent with a plume genesis. These basalts are similar in major-element composition to basalts in the Snake River Plain (Lum et al., 1989), with similar calculated depths and temperatures of melting (Wang et al., 2002).

The region of deep melting investigated by Wang et al. (2002) appears to have been intruded by an earlier episode of mafic magmatism in the mid-Miocene. This is consistent with the aeromagnetic anomaly that transects this area south of the northern Nevada rift, and with the occurrence of mid-Miocene magmatic rocks in and around the Colorado River extensional corridor (Fig. 5). Although the melting profile of Wang et al. (2002) is based on basalts that are younger than 10 Ma, we suggest that Basin and Range extension may have set the stage for volcanism in the late Cenozoic from decompressional melting of residual hot mantle that was first emplaced there in the mid-Miocene.

Additional evidence that the Nevada–Columbia Basin magmatic belt is currently underlain by anomalously hot mantle include: (1) the modeling results of Parsons et al. (1994) and Saltus and Thompson (1995), which demonstrate, independently, that the high altitude of the extended terrain in northern Nevada must be isostatically supported by a broad mass of hot, low-density mantle, consistent with the high heat-flow measurements in that region (Lachenbruch and Sass, 1978), (2) seismic evidence for low-velocity mantle beneath the Basin and Range province (e.g., Schmandt and Humphreys 2010), some of which has been attributed to elevated mantle temperatures combined with partial melting and/or hydrated mantle (e.g., Sine et al., 2008; Rau and Forsyth, 2011), (3) thermomechanical and geochemical modeling showing that flood basalt volcanism in a region of little or no extension requires the arrival of a hot mantle source (White and McKenzie, 1989, 1995), and (4) the recent seismic interpretations of a contemporary hot mantle plume beneath Yellowstone (Obrebski et al., 2010, 2011; Humphreys and Schmandt, 2011; Schmandt et al., 2012; Tian and Zhao, 2012; Darold and Humphreys, 2013), lying along an age-progressive hotspot track that places its arrival near the Oregon-Nevada-Idaho tristate area at 16.5 Ma.

Although we attribute the abrupt arrival of hot mantle beneath the Nevada–Columbia Basin magmatic belt to rapid rise of the elongated Yellowstone plume head, a broader region of plume spreading beyond the plume axis cannot be ruled out. A possible indicator of the westward encroachment of hot mantle comes from the apatite fission-track data of Surpless et al. (2002). They noted that a significant increase in geothermal gradients occurred in the transition zone between the Basin and Range and the Sierra Nevada in the mid-Miocene. They suggested that this event led to a decrease in crustal strength, resulting in the initiation of large-scale extension of this region at ca. 15–14 Ma.

LITHOSPHERE FOUNDERING AT THE PERIPHERY OF PLUME EMPLACEMENT

EarthScope's USArray provides an unprecedented data set of detailed seismic imagery that resolves a number of subvertical, high-velocity mantle structures (e.g., Burkett and Gurnis, 2013) interpreted by several workers as lithospheric downwellings beneath the edges of crustal blocks bounding the Basin and Range Province to the north, southwest, and southeast (Saleeby et al., 2003; Zandt et al., 2004; Frassetto et al., 2011; Levander et al., 2011; Obrebski et al., 2011; Darold and Humphreys, 2013). These large-scale features are similar to those generated in model experiments, where slab-like downwellings of lithospheric mantle and lower crust are produced at the edges of outwardly spreading mantle plumes (Fig. 9), with enhanced delamination occurring where plumes advance beneath crustal blocks of variable thickness and thermomechanical behavior (Burov et al., 2007; Guillou-Frottier et al., 2007, 2012; Cloetingh et al., 2013).

At the northern end of the Nevada–Columbia Basin magmatic belt in southeastern Washington, Camp and Hanan (2008) used field and chemical data to suggest that subvertical downthrusting of Mesozoic oceanic lithosphere may have occurred during plume impingement against the westward bend of the cratonic boundary north of the Chief Joseph dike swarm (Fig. 4). Darold and Humphreys (2013) presented an alternative interpretation, which posited that a fragment of the Farallon plate (Siletzia) was accreted to the continent in the early Tertiary, but it was removed in the mid-Miocene by vertical downthrusting

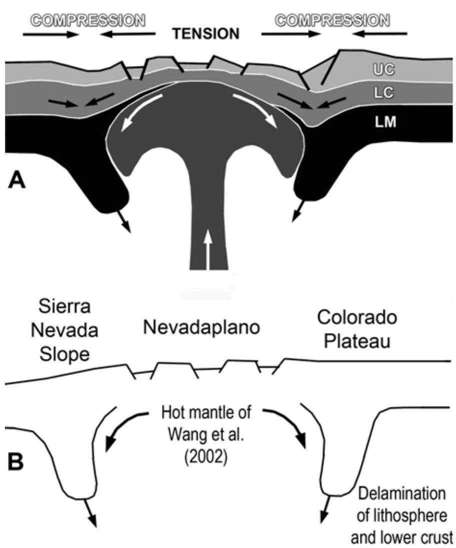

Figure 9. Mantle downwelling at the periphery of plume emplacement. (A) Schematic cross section based on numerical experiments showing the rheological effects of plume emplacement on lithosphere composed of stratified layers having variable ductile characteristics (i.e., UC—upper crust; LC—lower crust; LM—lithospheric mantle); modified after Cloetingh et al. (2013). The numerical models demonstrate that plume impact causes each layer to deform internally with its own characteristic wavelength, with the delamination of mantle lithosphere a common occurrence near the convecting edge of spreading plumes (e.g., Burov et al., 2007). This downwelling behavior and the production of slab-like instabilities are further enhanced where plumes spread outward against thicker crustal blocks and cratonic boundaries (Burov et al., 2007; Guillou-Frottier et al., 2007, 2012). (B) Schematic east-west cross section of proposed mantle downwelling beneath the Sierra Nevada batholith and the Colorado Plateau by mantle traction symmetrically associated with tabular rise and spreading of hot mantle beneath the southern portion of the Nevada–Columbia Basin magmatic belt. Such a model is consistent with seismic imagery (e.g., Frassetto et al., 2011; Levander et al., 2011), and with the numerical simulations of Liu and Shen (1998b), demonstrating that mantle upwelling beneath the region of Basin and Range extension could result in removal of ductile mantle and lower crust in adjacent terrains through a series of similar delamination events.

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associated with the northward-propagating Yellowstone plume. They cited recent seismic data that resolve this delaminated fragment as a vertical high-velocity slab ~350 km beneath the Chief Joseph dike swarm (Schmandt and Humphreys, 2011). They suggested that peeling away of the Siletzia fragment also enabled more localized foundering of high-density restite beneath the Cretaceous Wallowa batholith of northeastern Oregon, with the loss of this plutonic root facilitating magma rise and voluminous flood basalt volcanism at the site of the Chief Joseph dike swarm.

Delamination is also evident west of the Basin and Range beneath the southern Sierra Nevada mountain range, which lacks a significant crustal root and is largely supported today by buoyant asthenosphere beneath an abnormally thinned lithosphere (Jones and Saleeby, 2013). The onset of lithosphere removal appears to have been contemporaneous with late Cenozoic uplift of the High Sierra. Liu and Shen (1998b) attributed this event to mantle upwelling during the main phase of Basin and Range extension. They tested this idea with numerical simulations showing that ductile flow beneath the Sierra would produce a series of delamination events with crustal thinning comparable to the observed seismic data.

Today, most workers believe that delamination in the southern Sierra began in the late Miocene or early Pliocene due to the gravitational collapse of garnet-pyroxenite rocks at the base of the Sierra Nevada batholith (e.g., Saleeby et al., 2003, 2013; Jones et al., 2004, 2014; Zandt et al., 2004). Since these high-density roots to the batholith were generated from conductive cooling that began in the Cretaceous, it seems reasonable to ask why they would remain stable until the late Miocene. Saleeby et al. (2003) suggested that delamination must have been linked to convective forces associated with Basin and Range extension, which, as described here, accelerated in the mid-Miocene. We suggest that evidence for earlier delamination that began in the mid-Miocene might have been lost to later delamination events. Alternatively, mid-Miocene upwelling of hot mantle beneath the Sierra lithosphere would have decreased asthenosphere density and viscous strength (Elkins-Tanton, 2007), thus providing ideal conditions for delamination to take place in the late Miocene.

If delamination is the result of hot mantle upwelling beneath the Basin and Range, then one would expect to see similar downwellings on the eastern side of the province. Indeed, the recent USArray data demonstrate that portions of the lithospheric root of the Colorado Plateau are also found to be missing (Liu et al., 2011; Obrebski et al., 2011). Levander et al. (2011) resolved a vertical high-velocity anomaly beneath the west-central Colorado Plateau that extends more than 200 km in depth. They proposed that this body of mantle and lower crust was delaminated from the base of the plateau at the beginning of Pliocene uplift, and that it represents only one of a series of earlier delamination events. They argued that lithospheric foundering is the result of basaltic melts from upwelling mantle that infiltrated the base of the plateau, where they crystallized to high-density rocks, thus producing Rayleigh-Taylor instabilities. They did not recognize any process capable of such melting in the mid- to late Miocene and therefore proposed that upwelling was related to removal of the Farallon slab 20–30 m.y. ago. We suggest instead that such a source may have arrived in the mid-Miocene with ascent and outward spreading of the elongated Yellowstone plume head beneath the Basin and Range, with the main mechanism of delamination coming from mechanical downthrusting at the interface of the western Colorado Plateau.

There may have been other contributing factors to delamination, but the timing and deformational style of these events, in widely spaced regions around the periphery of the Basin and Range province, are consistent with model predictions of active mantle downwelling along well-defined crustal boundaries during plume emplacement (Liu and Shen, 1998b; Burov et al., 2007; Guillou-Frottier et al., 2007, 2012).

PALEOELEVATION CHANGES

Thermal buoyancy associated with the contemporary Yellowstone plume tail has generated an ~1000-km-wide geoid anomaly, the largest such feature in North America (Roman et al., 2004), in an area where estimates of gravitational potential energy attain their greatest values (Humphreys and Coblentz, 2007). Uplift associated with the arrival of the Yellowstone plume near the Oregon-Nevada border region hypothetically should have generated an anomaly and topographic swell at least as large, or in the case of a plume head, significantly larger (e.g., Sleep, 1990; Hill et al., 1992).

Mid-Miocene uplift along the hotspot track is evident in the U-Pb detrital zircon data of Beranek et al. (2006). They show that early paleodrainages at ca. 15 Ma were first directed radially away from a tumescent highland above the plume center in northern Nevada before reversing flow toward the Snake River Plain basin as the hotpot migrated to the northeast. Chamberlain et al. (2012) suggested that at least some part of low δ^{18}O values recorded in that region by Takeuchi and Larson (2005) and Kohn et al. (2002) can be attributed to ancient streams draining into eastern Oregon from the rising "bulge" of Beranek et al. (2006). Further evidence for an eastward migration of uplift along the hotspot track comes from a large number of climate proxies for uplift summarized in Pierce et al. (2002). Both uplift and volcanism were accompanied by Basin and Range normal faulting, which has also migrated to the northeast in tandem with the hotspot since ca. 10 Ma (Anders et al., 1989; Anders and Sleep, 1992; Anders, 1994).

Evidence for a period of uplift immediately preceding the initiation of flood basalt volcanism at ca. 16.7 Ma comes from deep canyons cut into the prebasalt surface along the Idaho–northeast Oregon border that were filled by the first flood basalt flows (Imnaha Basalt) to erupt in that area (Kleck, 1976; Hooper et al., 1984). Further uplift during flood basalt volcanism is demonstrated by the development of a comagmatic paleoslope in the flood basalt stratigraphy that reflects a south-to-north migration of uplift at ~2–3 mm/yr, in the same direction as the northward migration of flood basalt volcanism from southeastern Oregon to southeastern Washington (Camp, 1995).

In contrast, evidence for regional uplift in the early stages of plume emplacement in the northern Basin and Range is lacking. Chamberlain et al. (2012) did not observe a rapid decrease in δ^{18}O values at 17 Ma, which should record such an event, although they acknowledged that their oxygen isotope data over this critical period are limited to only one section near Elko, Nevada. On the other hand, increased oxygen isotope values record the mid-Miocene collapse of the Nevadaplano that began in the northern and central Basin and Range at ca. 17–16 Ma (Chamberlain et al., 2012, p. 240). This seems to be corroborated by paleoelevation estimates from leaf physiognomic studies showing that surface altitudes were ~1.5 km higher in the mid-Miocene than they are at present (Wolfe et al., 1997). The rapid decrease in surface elevations is consistent with the data presented here and elsewhere for a major period of crustal thinning from ca. 17 to 12 Ma (e.g., Wernicke and Snow, 1998; McQuarrie and Wernicke, 2005; Colgan and Henry, 2009; Quigley et al., 2010).

Chamberlain et al. (2012) suggested that the proxy data at ca. 17 Ma in Nevada are still largely unexplored. Nevertheless, taking the current data at face value suggests that the topographic response to plume emplacement in north-central Nevada is reflected in the abrupt initiation of a major period of extension leading to collapse of the Nevadaplano. In contrast, the topographic response to plume emplacement in northeastern Oregon, eastern Washington, and southwestern Idaho is reflected in the contemporaneous initiation of uplift with little associated crustal extension.

DRIVING FORCES AND PLATE KINEMATICS

In their quantitative analysis of driving and resisting forces for Basin and Range extension, Sonder and Jones (1999) concluded that boundary forces are not capable of driving extension throughout the northern and central Basin and Range, but that such extension is more likely the result of high gravitational potential energy derived from either buoyancy forces intrinsic to the lithosphere or basal-normal forces imposed on the lithosphere from below (i.e., plume impingement and/or delamination effects). The modeling results of Humphreys and Coblentz (2007) and Ghosh et al. (2013) are also in accordance with a dynamic topography with high gravitational potential energy being the primary control on extensional deformation in the Basin and Range today.

Liu (2001) agreed that buoyancy forces associated with a thickened crust can generate a topographic highland that is dynamically unstable and tends to collapse; however, he argued that major and widespread extension can only take place when the lithosphere is sufficiently weakened by a thermal perturbation. In light of the evidence for emplacement of a hot mantle source at ca. 17 Ma, it seems reasonable to conclude that the arrival of the Yellowstone plume played a significant role in the initiation of Basin and Range extension of the same age. Still, plate-margin kinematics and buoyancy forces from crustal thickening of the Nevadaplano cannot be ignored. Uplift of the Nevadaplano began with crustal shortening in the Mesozoic and early Tertiary (Coney, 1987; DeCelles, 2004) and was further uplifted to ~3–4 km during the mid-Tertiary volcanic sweep (Flowers et al., 2008; Chamberlain et al., 2012), with no evidence of collapse until ca. 17–16 Ma. This suggests that the Nevadaplano was at its highest elevation at the time of plume emplacement, consistent with a region that was primed for collapse.

Early attempts to describe the origin of Basin and Range extension were largely focused on a single cause, but it seems clear from more quantitative analysis that multiple forces are at work (e.g., Sonder and Jones, 1999; Flesch et al., 2000; Humphreys and Coblentz, 2007; Ghosh et al., 2013). Emplacement of the Yellowstone plume was not the sole cause of Basin and Range extension, but rather the catalyst that initiated widespread extension as additional forces were superimposed on a high plateau, under regional stress, that was already on the verge of collapse. The apparent lack of uplift in the northern Basin and Range at ca. 17 Ma (Chamberlain et al., 2012), combined with contrary evidence for the onset of extension and crustal stretching at ca. 17–16 Ma, can be attributed to a variety of factors. The high gravitational potential energy of the Nevadaplano, together with the rapid emplacement of magmatic intrusions, advective heating and thermal weakening of the lithosphere, additional buoyancy associated with plume emplacement, and vigorous traction of asthenospheric mantle associated with plume spreading and lithospheric foundering may have all played a role in triggering the abrupt and widespread collapse of the Nevadaplano.

The area of collapse corresponds with the area of Basin and Range faulting that began at ca. 17–16 Ma (Fig. 5). This area excludes most of northwestern Nevada and southeastern Oregon, where there is a lack of evidence for either uplift or collapse during plume emplacement. This excluded area is underlain by oceanic to transitional crust and did not undergo the extensive crustal thickening that is evident in the adjacent continental rocks beneath the Nevadaplano (Figs. 2 and 4). The current average altitude of the highly extended northern Basin and Range, at ~1.5 km above sea level, suggests that collapse of the Nevadaplano is not complete, due to the thermal buoyancy of the hot mantle that still resides beneath this region (e.g., Parsons et al., 1994; Saltus and Thompson, 1995).

Boundary forces due to Pacific–North American plate motion also played a significant role, which is partly expressed in the orientation of the regional stress field (extension at ~280°–300°) operating across the central and northern Basin and Range since ca. 36 Ma (Flesch et al., 2000; McQuarrie and Wernicke, 2005). Closer to the plate boundary, these forces were generating significant deformation that was well under way by 24–22 Ma, before plume-triggered extension began in the plate interior. This is evident in detachment faulting and tectonic exhumation in the California borderlands, Los Angeles Basin, Mojave Desert, and the Colorado River extensional corridor, contemporaneous with rotation of the western Transverse Ranges (Hornafius et al., 1986; Crouch and Suppe, 1993; Nicholson et al., 1994; Faulds et al., 2001; Dickinson, 2002; Miller, 2002). Boundary forces also played the major role in short-lived extension (ca. 28–14 Ma) of the southern Basin and Range of southeastern California, Arizona, and northern Mexico (Henry and Aranda-Gomez, 1992; Bohannon and Parsons, 1995). There is no evidence that a mantle plume was involved in extension or melting in these areas, but instead volcanism appears to have been associated with transtension and the passive rise of magma derived from a shallow, slab-window source (McQuarrie and Oskin, 2010). The effect of plate boundary forces is also evident in the westernmost portion of the northern and central Basin and Range, particularly along the Walker Lane belt, where right-lateral faulting began ca. 12 Ma (Faulds et al., 2005; Faulds and Henry, 2008; Busby, 2013; Busby et al., 2013a, 2013b).

During mid-Miocene extension (ca. 17–12 Ma), boundary forces were progressively weaker at greater distances from the plate boundary, while forces associated with high gravitational potential energy differences were more dominant (e.g., Sonder and Jones, 1999; Humphreys and Coblentz, 2007). The overall mid-Miocene regional stress regime (driven by the combination of plate boundary and gravitational forces) diminished in magnitude northward, with very low magnitudes and random orientations in northwestern Nevada, eastern Oregon, and eastern Washington (Flesch et al., 2000; Humphreys and Coblentz, 2007; McCaffrey et al., 2007, 2013; Payne et al., 2012), as demonstrated by the absence of NE-trending mid-Miocene extensional structures in those regions (e.g., Colgan et al., 2004; Scarberry et al., 2010). Here, north of the Nevadaplano, localized magmatic stress played a larger role in producing the generally consistent orientation of NNW–trending dikes (Steens dikes being the exception) by forceful dike injection during the rapid, main phase of plume-induced magmatism from 16.7 to 15 Ma. After ca. 12–10 Ma, the younger and weaker phase of Basin and Range extension expanded into northwestern Nevada and southeastern Oregon, contemporaneous with boundary-driven transtension along the Walker Lane belt, and into northeast Nevada and southern Idaho, contemporaneous with the northeast migration of volcanism along the Yellowstone–Snake River Plain hotspot track (Fig. 5; Anders et al., 1989; Pierce and Morgan, 1992, 2009).

Despite the presence of these driving forces, plate kinematics must provide the right conditions to allow extension to take place. Cousens et al. (2008) noted an abrupt pulse in volcanism along the Miocene ancestral Cascades arc along the California-Nevada border, followed by westward progression of radiometric ages from the mid-Miocene into the Pliocene. This is consistent with the initiation of mid-Miocene slab rollback, thus providing the space needed for Basin and Range extension and the northward translation of the Sierra Nevada and Great Valley block (e.g., Humphreys and Coblentz, 2007). Continued northward translation and Basin and Range extension have been aided by a zone of net area decrease in northwestern California (between 38°N and 42°N) that has been described as a "window of escape" (Flesch, 2007; Kreemer and Hammond, 2007), guiding the extension directions throughout the history of Basin and Range crustal stretching.

CONCLUSIONS

The genetic relationships among mid-Miocene arrival of the Yellowstone hotspot, Basin and Range extension, and coeval magmatism are based on a number of observations and interpretations that lead to the following conclusions.

(1) Mid-Miocene Basin and Range extension began abruptly at ca. 17–16 Ma with a widespread period of crustal stretching that continued to ca. 12–10 Ma, producing NNE–trending depositional basins and fault blocks across a wide expanse of the northern and central Basin and Range. Continued extension after 12–10 Ma was of lower magnitude and focused on the western, eastern, and northernmost margins of the province.

(2) Seismic imagery has resolved a conduit of hot, low-density mantle beneath Yellowstone National Park, which has been interpreted by several studies as a mantle plume tail with a source in the lower mantle. The mid-Miocene location of the plume was centered near the Oregon-Nevada-Idaho tristate region at 16.5 Ma, coincident in time and space to the initiation of large-scale Basin and Range extension at 17–16 Ma, the initiation of the earliest Columbia River flood basalts in southeastern Oregon at 16.7 Ma, the initiation of rhyolite eruptions at the western end of the Yellowstone–Snake River Plain hotspot track at 16.5 Ma, and the initiation of volcanism along the northern Nevada rift system at 16.5 Ma. These close temporal relationships suggest that the Yellowstone plume may have played a major role in the initiation of volcanism and extension of the same age.

(3) The short-lived (16.7–15 Ma) Nevada–Columbia Basin magmatic belt is delineated by NNW–trending intrusions and volcanic centers extending over a distance of >1000 km, from southern Nevada into the Columbia Basin of eastern Washington. The belt is defined entirely by magmatic structures with no evidence of significant extension in Oregon and Washington, and very little evidence of extension along the northern Nevada "rift." The center of the belt lies near the projected site of plume impingement at 16.5 Ma. Little or no extension along a belt that includes flood basalt volcanism is inconsistent with magmatism resulting from the passive rise of mantle into an extended lithosphere, and it is more consistent with a process of active mantle upwelling and voluminous adiabatic melting.

(4) The orientation of regional stress that exists in the northern and central Basin and Range today has been in existence since ca. 36 Ma. Northeast-trending faults and depositional basins produced during mid-Miocene extension correspond with a 280°–300° orientation of maximum extension. This orientation differs from the apparent 245°–250° extension direction for contemporaneous dikes along the Nevada–Columbia Basin magmatic belt. The ~45° difference in these magmatic trends, localized to areas with little or no extension, suggests that the dikes were emplaced by forceful injection. Their NNW trends were not controlled by regional stress in the upper crust, but instead by a deeper process in the sublithospheric mantle. The high magma flux, high magma overpressures, and rapid propagation enabled these dikes to maintain their original orientation without being realigned to the regional stress field.

(5) We attribute emplacement of the Nevada–Columbia Basin magmatic belt to rupture of a thermally weakened flexure in the Farallon slab as it was uplifted by a buoyant Yellowstone plume, consistent with recent seismic tomography and model simulations. As plume material squeezed through a N-S tear, the rapid rise of a tabular mass of hot mantle resulted in adiabatic partial melting and the abrupt onset of surface volcanism along the entire length of the magmatic belt at ca. 16.7–16.5 Ma.

(6) The Nevada–Columbia Basin magmatic belt may represent the axial surface manifestation of late Cenozoic plume convection under a broader region of the Basin and Range. Lateral spreading at the base of the lithosphere may have resulted in mantle traction and delamination at the edges of crustal blocks surrounding the province. Such areas are revealed by seismic data as mantle drips or foundering slabs located beneath northeastern Oregon adjacent to an E-W bend in the cratonic boundary, beneath the southern Sierra Nevada batholith, and beneath the stable Colorado Plateau. Removal of ductile mantle and lower crust at the edges of such crustal blocks is consistent with numerical models of plume emplacement.

(7) Early Tertiary uplift of the Nevadaplano continued during the mid-Tertiary volcanic sweep. This high plateau was therefore primed for collapse at the time of plume impingement in the mid-Miocene. The onset of extension was largely the result of gravitational collapse of the Nevadaplano, which was triggered by an abrupt thermomechanical perturbation of the lithosphere associated with plume emplacement. Today, the residual heat energy from this mid-Miocene perturbation is evident in heat-flow measurements, melt temperature calculations on late Cenozoic volcanic rocks, and buoyancy calculations. Boundary forces from North American–Pacific plate interaction also played an important role in controlling the 280°–300° extension direction in the northern and central Basin and Range, and in providing the dominant control on deformation in the southern Basin and Range, the California borderlands, Los Angeles Basin, Mojave Desert, and the Walker Lane belt.

(8) The magnitude of regional stress from boundary and buoyancy forces was greatly diminished north of the Nevadaplano. This area of northwestern Nevada, eastern Oregon, and eastern Washington was unaffected by Sevier-age crustal thickening and was far removed from Pacific–North American plate interaction. There was therefore a paucity of significant mid-Miocene (17–12 Ma) extension in northwestern Nevada, eastern Oregon, and eastern Washington. Here, flood basalt volcanism along the elongated plume axis increases in volume to the north. This greater volume is attributed to the thin nature of the oceanic lithosphere and the higher density of thin oceanic crust, both of which favor the ascent of basaltic magmas.

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