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THE MIOCENE GREAT BASIN OF WESTERN NORTH AMERICA AS AN EXTENDING BACK-ARC REGION *

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ABSTRACT

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A narrow continental volcanic arc has lain near the western margin of the North American continent since mid-Miocene time. Its south end began retreating northward 18 m.y. ago. The active remnant of this andesitic arc today is the Cascade Range of Oregon, Washington, and British Columbia.

A large part of the Great Basin and region to its north was situated throughout this period first wholly, then partially, in a back-arc position relative to the arc. This region had, and still has, most of the characteristics of back-arc regions elsewhere in the world—a thinned, faulted and extended crust and lithosphere, limited development of a geophysical bilateral symmetry (a mid-to-late-Miocene symmetry axis is marked by narrow swarms of basaltic dikes with associated positive linear magnetic anomalies that are parallel to the contemporary trench), high heat flow, low Bouguer gravity and seismic wave velocity values, notable seismic attenuation in the mantle, and a fault plane solution record dominated by normal-fault mechanisms.

This back-arc region differs somewhat from most others (those composed of oceanic crust) because of its continental setting and recent geologic history. Despite its name, the Great Basin stands high, nearly 2 km above sea level. This high elevation owes, in part, to the fact that extension and heating of the continental lithosphere are still underway, though not as a back-arc phenomenon. The present lithospheric stretching results from oblique extension near the transform boundary of the North American plate, the San Andreas fault system. The latest stretching has probably maintained the physical state of the lithosphere first developed during back-arc spreading.

INTRODUCTION

History of investigation

More than a century of debate concerning the structure and origin of the Basin and Range province has passed, most of it focused on the Great Basin, both because of longer and more detailed geological and geophysical scrutiny there, and because

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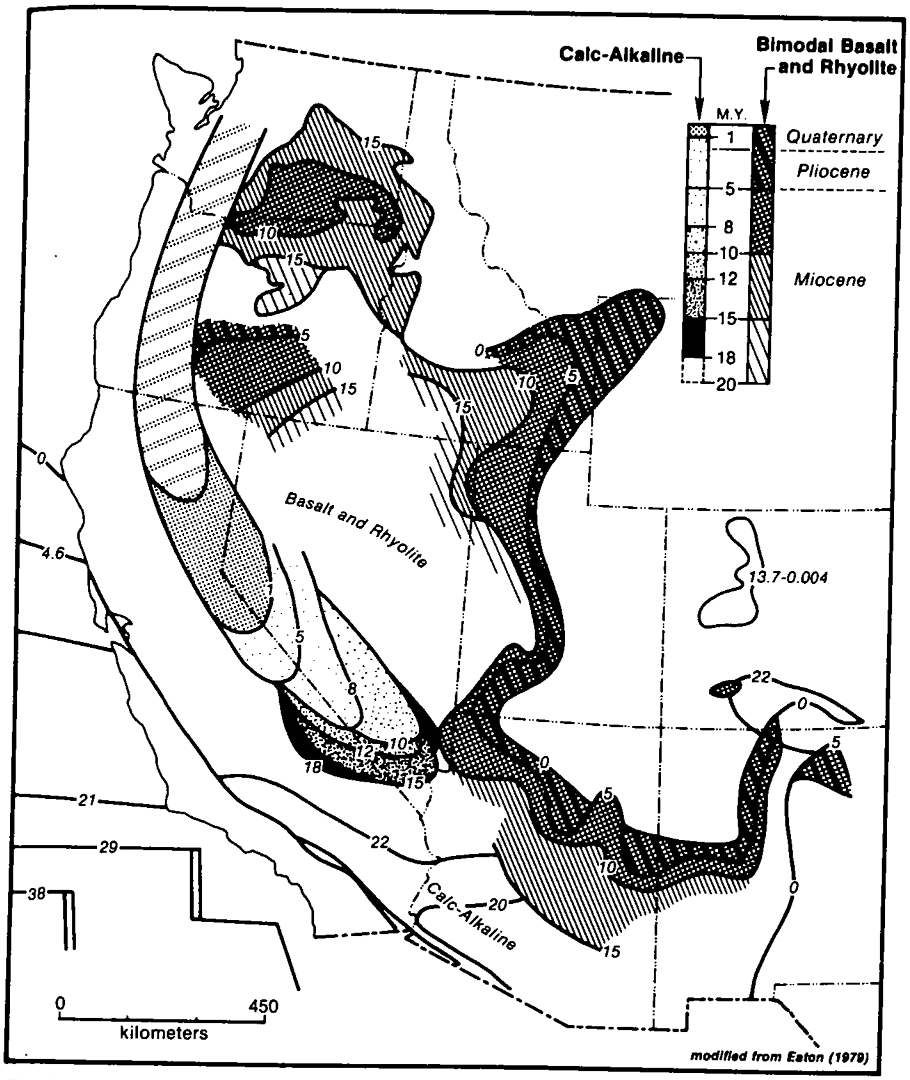
The best documented record of early termination of calc-alkaline volcanism in the Great Basin is found on its east side, near Marysville, Utah. It occurred there in the period 22–24 m.y. For the rest of the region the record is less clear for the next 5 m.y., but by 17–18 m.y. ago, a well-defined, classically proportioned and continuous andesitic volcanic arc lay along the western margin of the Great Basin, its axis



modified from Eaton (1979)

Fig. 2. Spatial distribution of calc-alkaline intermediate and bimodal (basaltic and high silica rhyolitic) volcanic rocks in the western United States, 18 m.y. ago to the present. The calc-alkaline rocks define a continental volcanic arc, the northern three-fifths of which constitute today's active Cascade Range volcanic chain. While the south end of this arc was retreating northward, bimodal activity migrated outward only a short distance near the margins of the Basin and Range province (from Cross and Pilger, 1978; and Eaton, 1982).

aligned along the California–Nevada border in the south and along the locus of the Cascade range in the north (Fig. 2). Behind it lay a region of bimodal volcanism (Fig. 2), some of it, e.g. the middle Miocene Columbia River basalts, of a tholeitic plateau or flood-like nature. The region behind this arc was characterized then as now by the emplacement of distributed normal faulting and by swarms of basaltic or basaltic andesite dikes. The dikes that occupy the *axis* of the region display sharp, stripe-like magnetic anomalies. Together, this is the evidence suggesting that the Miocene Great Basin was an actively spreading back-arc region.

If lithospheric thinning underway in Miocene time has continued to the present, changing only in rate, direction and aggregate degree, the basic physical state of the crust and lithosphere may have been, at least to a crude first approximation, similar to what it is today. Following this argument, it will be instructive to examine the present characteristics of the province, for they are generally in accord with those of actively (or recently) spreading back-arc regions. They thus tend to support the fossil evidence for the Miocene and younger back-arc character of the region. One may anticipate at the outset of such an examination that because the region has evolved as *continental* crust and lithosphere, its attributes will display some degree of contrast within those of oceanic crust and lithosphere deformed in the same manner.

PHYSICAL CHARACTERISTICS OF THE GREAT BASIN

Geophysical symmetry

The Great Basin displays a subtly developed geophysical and topographical bilateral symmetry (Eaton, 1976; Eaton et al., 1978; Eaton, 1982). It is unlike that of oceanic spreading ridges or back-arc basins in that its elements are those of long wavelength (100 km and greater) gravity anomalies rather than short wavelength (5–50 km) magnetic anomalies. The axis of this bilateral symmetry trends nearly N–S (Eaton et al., 1978, figs. 3–8), but in the center of the region, at the mid-latitudes of Nevada, there is a large gravity anomaly pair cocked counterclockwise to the general axis of regional symmetry, its pattern not unlike that of the wings of a butterfly. The axis of this pair trends NNW, parallel to the arc and trench (Fig. 3). It appears to reflect an early stretching of the lithosphere. It is overprinted by a younger, high-frequency gravity grain that does not participate in the symmetry.

Neither this older symmetry, nor that of the Great Basin as a whole, is reflected in a symmetry of the surface or near-surface geology. The surface geology is the end product of a very long and complex crustal history. It thus displays a significant contrast with the generally much younger, monolithologic, and essentially homogeneous oceanic crust of rather simple history, a crust displaying detailed patterns of magnetic stripes developed at its spreading ridges. Such magnetic stripes are much more poorly developed in back-arc basins than at the ridges and this fact may also explain their apparent scarcity in the Great Basin.

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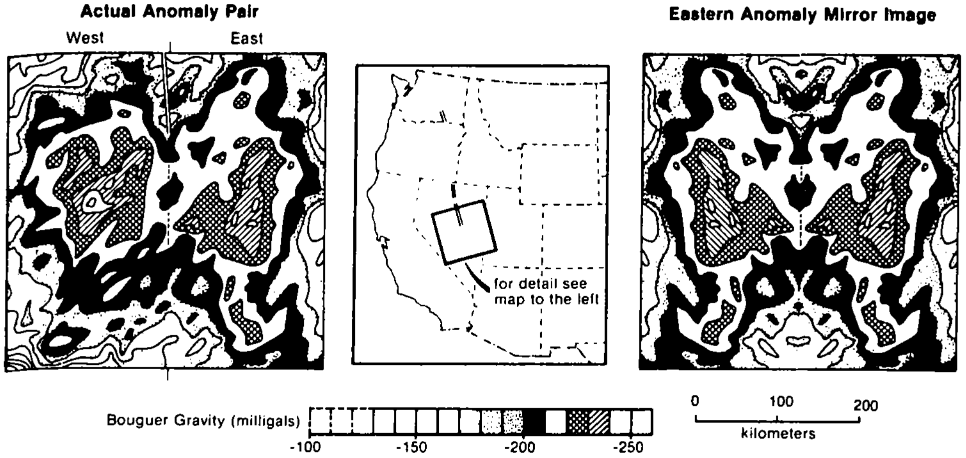

Fig. 3. Gravity symmetry in the central Great Basin. Also shown on the index map (center) by heavy parallel lines are the axes of discontinuous, linear magnetic highs. The latter are associated with exposed basaltic dike swarms along the axis of Miocene back-arc spreading. The southernmost magnetic lineation also appears on the actual gravity map. From Eaton et al. (1978) and Zietz (1982).

Since the strength of a magnetic field arising from a simple source configuration decays approximately as the third power of the distance to its source, while the gravity field decays as the square of that distance, magnetic data tend to emphasize shallower features. The magnetic field over an area in which the geology reflects the products of events spanning a range from Precambrian to Holocene time will be highly complex in terms of pattern development, as well as pattern discrimination and recognition. Despite this fact, there are two areas in the Miocene back-arc region of the western U.S. that display prominent stripe-like, positive magnetic anomalies not unlike those developed at oceanic ridges. One of these is in northern Nevada and the other, in southeastern Washington (see heavy parallel lines in Fig. 3). Exposed swarms of basaltic or basaltic andesite dikes of middle Miocene age are associated with both. Sub-parallel dikes of like nature are seen elsewhere in the region, but do not have similar magnetic expression. The dike swarms and associated magnetic anomalies described here are distinguished by their sharp and linearly continuous character. They are co-linear with the axis of gravity symmetry in central Nevada. They reflect an aborted continental extensional rift that was parallel to the contemporary arc. It developed several hundreds of kilometers from the trench of that time. Because the magma that nourished it was mantle-derived, the lithosphere must have parted locally, but only to a degree equal to their aggregate width.

The crust of the Great Basin has not been "oceanized", but it developed some of the characteristics of what is generally referred to as "transitional crust," that which we sense geophysically at a passive margin, the boundary between continental and oceanic crust after lithospheric parting.

Heat flow

High heat flow is a prime characteristic of back-arc basins. Watanabe et al. (1977) proposed a general history for newly created back-arc basins, one in which the basin lithosphere cools toward a steady state value of approximately 2.2 HFU and, with a thin lithosphere, a maintenance of that value for several tens of millions of years. These investigators reported modal heat flow values of 2.5 HFU for the Plio-Pleistocene Okinawa trough and Mio-Pliocene North Fiji basin and 2.0 HFU for the Oligo-Miocene Shikoku and Parece Vela basins.

Average heat flow in the Great Basin today is 2.20 HFU. The mode is 2.15 HFU (see Fig. 4A). Direct comparison of these values with those of marginal basins must be made with caution, however, for radiogenic heat production in a potash-rich continental crust has the potential for elevating observed heat flow values above those typical of oceanic crust. To get around this problem, a comparison of *reduced* heat flow values is advised. Reduced heat flow in the Great Basin (average value, 1.66 HFU) is greater than that of stable North America by as much as 50–100% (Lachenbruch and Sass, 1978) and the warmer subregions within or near it, like those at Battle Mountain, Nevada, at Long Valley, California and in and near the Rio Grande rift in southern New Mexico, display reduced heat flow values that exceed those of the stable craton by as much as 300%. It thus appears that the anomalously high heat flow of the Basin and Range province is more likely a reflection of crustal extension, groundwater convection, magmatic intrusion, and possibly some local thermal refraction at basin-range boundaries, than of abnormally high crustal heat production.

Although the range of observed heat flow values in the Great Basin is high (from 0.4 to 4.8 HFU) and heat flow is sharply variable on a local scale owing to convection and refraction, several attempts have been made to define regional patterns and even to smooth and contour the data (Blackwell, 1978; Eaton et al., 1978; Swanberg, in Sass et al., 1981). The results of the most comprehensive of these efforts, one based on the silica geothermometry of groundwaters from several thousands of wells, as opposed to thermal gradient measurements made in fewer than 150 wells studied thus far, is shown in Fig. 4. One sees that a large central area of the Miocene and younger back-arc region under discussion is characterized by a heat flow of 2.5 HFU and that this value drops outward in all directions, falling to as low as 1.5 HFU on the Colorado Plateaus to the east.

Most investigators (Blackwell, 1969; Sass et al., 1971; Roy et al., 1972; Lachenbruch and Sass, 1977, 1978) agree that the anomalous heat loss in the Great Basin is probably due to some form of mass transfer in the lithosphere (e.g., ductile stretching, magmatic intrusion, volcanism or hydrothermal convection), but they differ with one another in terms of the relative importance they attach to each of these contrasting mechanisms. For example, Blackwell (1978) suggested that hydrothermal convection and plutonism-volcanism play the prime role in energy transfer

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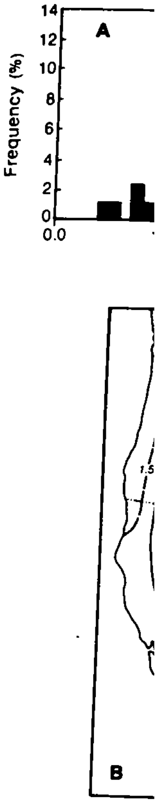

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where thermotectonic activity is younger than 17 m.y. This is the period of that style of extensional deformation referred to as basin-range faulting. He further suggested that where major thermal events are older than 17 m.y., conduction may be the dominant heat transfer mechanism, with hydrothermal convection being significant primarily only on a local scale.

Maps of the surface distribution of volcanic rocks for different time periods in the Mesozoic and Cenozoic Eras in the Great Basin (Stewart and Carlson, 1976; Stewart, 1980; Eaton, 1982, fig. 5) reveal more vigorous and widespread magmatism in the period 17–6 m.y. (middle and late Miocene time) than in the time since 6 m.y. ago (Pliocene and Quaternary time). Following Blackwell's (1978) reasoning, these relations suggest a higher convective heat loss through magmatism while the Great

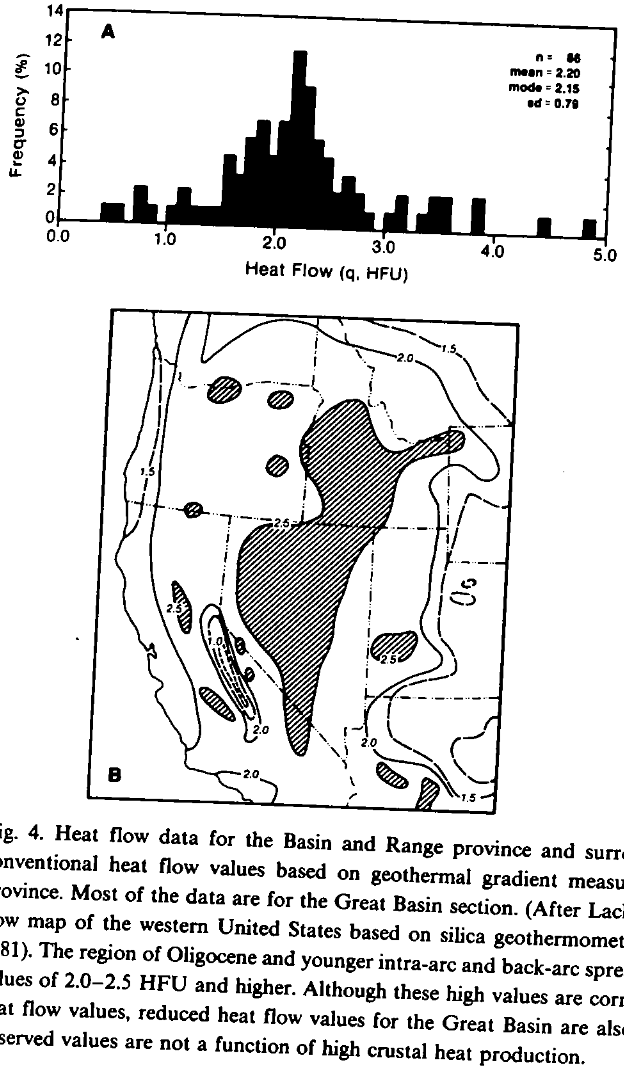


Fig. 4. Heat flow data for the Basin and Range province and surrounding regions. A. Histogram of conventional heat flow values based on geothermal gradient measurements in the Basin and Range province. Most of the data are for the Great Basin section. (After Lachenbruch and Sass, 1978.) B. Heat flow map of the western United States based on silica geothermometry (after Swanberg, in Sass et al., 1981). The region of Oligocene and younger intra-arc and back-arc spreading is characterized by heat flow values of 2.0–2.5 HFU and higher. Although these high values are correlative with conventional *observed* heat flow values, reduced heat flow values for the Great Basin are also high, reflecting the fact that the observed values are not a function of high crustal heat production.

unlikely that upper mantle velocities and heat flow would have been substantially lower then.

Seismic velocities in the crust also seem to be anomalous. It has been noted elsewhere that the Great Basin crust is characterized nearly everywhere at levels no deeper than one-third to one-half its total thickness by a zone of low compressional-wave velocities and high electrical conductivity (Eaton, 1980). Comparison of thirteen velocity profiles for the Great Basin crust with a reference profile in the interior of the Colorado Plateaus (data extracted from Prodehl, 1979) reveals that the upper, middle, and lower crust of the Great Basin have the following velocity properties relative to those of the Colorado Plateaus: (1) the *upper* crust of the Great Basin displays velocities that are generally less than (in a few localities they are equal to) that of the upper crust of the Colorado Plateaus; (2) the *middle* crust has velocities consistently lower than those of the Colorado Plateaus; and (3) the *lower* crust displays a variety of values, the majority of which are lower than (but some of which are greater than, and some equal to) those of the plateaus.

A mean of the velocity values calculated for rays traversing the crust perpendicular to its layering reveals that for seismic refraction lines originating at 28 shot-points in the Great Basin, the average crustal velocity is 6.13 km/s. By contrast, those lines originating at 9 shotpoints *outside* the Great Basin (in the Colorado Plateaus, Middle Rocky Mountains, Snake River Plain and Sierra Nevada range) show an average value of 6.37 km/s (see $\bar{u}$ values in Prodehl, 1979, tables 2–53). That the lower average seismic velocity in the crust of the Great Basin does not stem from its sedimentary layers is confirmed by corrected values of reduced travel times at critical distances utilizing P_g travel times (Prodehl, 1979, p. 43).

The observed crustal velocity contrast between the Great Basin and its surroundings, 6.13 vs. 6.37 km/s, should give rise to a density contrast of approximately $-0.04/\mathrm{g\ cm}^{-3}$. A density contrast of like sign probably also exists in the mantle, but as Eaton et al. (1978) have pointed out, a significant part of the -150 to -230 mGal gravity low that characterizes the Great Basin originates in the shallow crust.

Average Bouguer values over the Great Basin are nearly 100 mGal lower than those for most of the rest of the Basin and Range province. This intraprovince contrast is apparently due to lower Great Basin density values at all levels (upper, middle and lower crust and upper mantle alike), for although Eaton et al. (1978) noted the requirement of assigning part of the gravity anomaly source to the uppermost 15 km of the crust, Prodehl (1979, pp. 24–25) observed that velocities in the uppermost mantle of the Basin and Range Province are consistently higher than 8.0 km/s south of the southern edge of the Great Basin and consistently lower than 8.0 km/s within the Great Basin itself. Because the region to the south has been inactive tectonically for much of the past 5–10 m.y. and has an appreciably lower elevation (Eaton, 1979), this probably signifies early thermal subsidence, as well as advanced erosion and the development of through-going drainage unimpeded by continued uplift of the blocklike ranges.

Basin was a back-arc region, than in the period since then. Perhaps the total heat loss witnessed today represents a sum of the convective heat loss associated with present day extensional tectonism and conductive heat loss associated with Miocene lithospheric thinning and spreading. Whatever the case may be, it seems hard to escape the conclusion that the Great Basin was a region of high heat flow during Miocene time and that it thus had one of the dominant characteristics of active back-arc regions.

Seismic structure of the crust and lithosphere

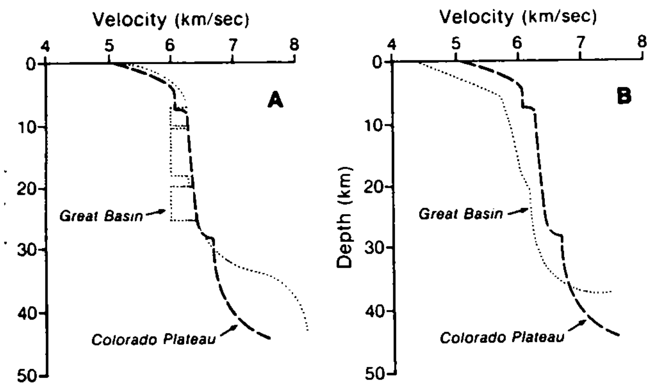
The crust of the Great Basin is everywhere less than 40 km thick. Locally, it is as thin as 23 km and, in at least one locality, it may be thinner than 20 km. Such thicknesses contrast sharply with those of the Sierra Nevada range on the west and the Colorado Plateaus on the east. There, thickness values everywhere exceed 40 km and are more generally found to range from 40 to 50 km (Thompson and Burke, 1974; Smith 1978, fig. 6-2).

The thickness of the Great Basin *lithosphere* has been estimated to be 65 km (Thompson and Burke, 1974). This is less than the worldwide average thickness for continental lithosphere. It is roughly in proportion to the anomalous thickness of the Great Basin crust relative to that of normal, continental cratonic crust. The asthenosphere thus protrudes upward beneath the Great Basin.

Seismic velocities

In addition to anomalous crustal thickness and high heat flow, upper mantle velocities beneath the Great Basin are anomalously low (7.4–7.9 km/s) relative to those beneath the craton (8.0–8.2 km/s; Pakiser and Zietz, 1965; Herrin, 1969; Prodehl, 1970, 1979; Thompson and Burke, 1974; and Smith,1978). Such P_n values are rarely found beneath continental crust (Japan and the Kurile Islands seem to be an exception). Such values are suggestive of elevated upper mantle temperatures.

A recent compilation of P_n data by Black and Braile (1982) revealed strong inverse correlations between P_n velocities and heat flow and between P_n velocities and estimated temperatures at the crust–mantle boundary. The empirical relationships derived by them fall well within a measured range of values based on experimental studies of the effect of temperature on compressional wave velocity of ultramafic rocks in the laboratory. An apparent correlation between heat flow on the one hand and crustal "age" and lithospheric thickness on the other, led Black and Braile (1982) to suggest that increases in lithospheric thickness and P_n velocity are mutually related processes, the result of cooling of the lithosphere following major thermotectonic events. Inasmuch as crustal extension is still underway in the Great Basin, we cannot attribute observed low P_n values with any certainty to the back-arc phase of spreading, but because the earlier episodes of extensional faulting appear to have been somewhat more vigorous than those at present (Eaton, 1982), it seems



g. 5. Crustal velocity sections for the Great Basin and a reference section in the Colorado Plateaus. Two pes of vertical profiles characterize the Great Basin: (A) one in which there is a distinct low-velocity yer in the Great Basin crust and (B) a continuously downward-increasing velocity profile that generally imics the profile for the Colorado Plateaus, except for uniformly lower velocities and a thinner crust. rom Prodehl, 1979).

Two velocity profiles from the Great Basin are shown in comparison to a eference profile for the Colorado Plateau in Fig. 5. The data are based on the nterpretations of Prodehl (1979). One profile represents a line running from the evada Test Site, near Mercury, Nevada, to Kingman, Arizona, the other, an ast–west line running from San Luis Obispo, California to Chinle, Arizona. The irst profile (Fig. 5A) displays a low-velocity layer from a depth of 7–25 km, ounded above and below by strong velocity gradients. The other (Fig. 5B) reveals a rofile similar in general form to that for the Colorado Plateaus, but one where reat Basin velocities are lower at all levels above the base of the crust than those of he plateaus. In both cases, the Great Basin crust is seen to be thinner than that of he plateaus.

The lower values of crustal seismic velocities in the Great Basin are probably ore the result of mechanical and thermal effects, than of an inherent compositional ontrast. Relatively elevated temperatures, a high degree of faulting and fracturing nd elevated pore pressures (leading to reduced effective pressures) may all contribute to a lowering of seismic velocities. A moderately high fracture porosity would reduce both seismic velocity and bulk density.

Macrofractures lower both V_p and V_s and increase the ratio V_p/V_s in granitic rocks, in situ (Moos and Zoback, 1983). There is a correlation between regions with macrofractures and anomalously low observed seismic velocities. Velocity reductions of these sorts are observed over a wide range of frequencies (10 Hz to 20 KHz), and the presence of macrofractures is seen to be a controlling influence on the velocity–depth function in granitic rocks.

Just how much of the observed velocity difference relates to conditions that obtained at the time of back-arc spreading and how much followed from the most (those associated with the growing transform boundary) is

impossible to determine. It does not seem unreasonable to suggest, however, that the general characteristics of a thin, low-velocity crust and lithosphere so typical of back-arc regions elsewhere in the world probably first developed in the Great Basin region during the initial phases of intra-arc and back-arc spreading and have simply been renewed and, therefore, maintained to the present.

HISTORY OF PLATE INTERACTION

Extensional history of the region

The history of extensional deformation in the Great Basin has been much described (Eaton et al., 1978; Eaton, 1979, 1982; Zoback and Thompson, 1978; Zoback et al., 1981). It began as intra-arc spreading associated with intermediate, calc-alkaline volcanism and continued first as back-arc spreading and then as transform-related oblique extension, the two later phases being associated with bimodal volcanism. The direction of spreading changed notably as the transform boundary (the San Andreas fault system) grew north-northwestward along the western margin of the continent, exerting greater and greater influence on tectonic activity inland.

Trajectories of minimum principal stress from two of the chapters of this history (back-arc and back-transform) are shown in Fig. 6. Initial extension was southwestward (Fig. 6A), normal to the trench, the extension being related to subduction of the Farallon plate. Later (after 10 m.y. ago), it was reoriented west-northwest, in consonance with the change from back-arc to oblique transform-related extension associated with northwest passage of the Pacific plate.

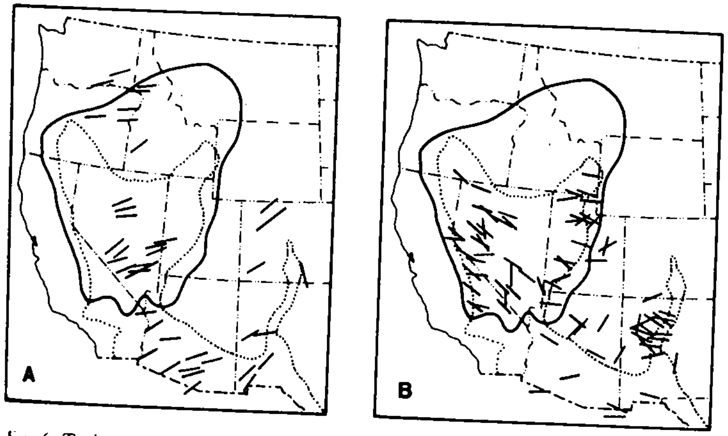


Fig. 6. Trajectories of minimum principal stress in the western United States (after Zoback et al., 1981, and Eaton, 1982). A. Mid-Cenozoic time, B. Late Cenozoic time. The first reflects the stress field associated with intra-arc and back-arc spreading and the second, the stress field associated with transform motion at the western margin of the N

Farallon–North American plate convergence

Because it is the back-arc phase of extension with which we are concerned here, it may be instructive to examine the chronologic relations of plate interactions over a span of time that includes this phase. Two recent publications have addressed this issue, the paper of Carlson (1982) and an abstract of Engebretson et al. (1982). Figure 7, taken from their working models, shows that from late Mesozoic through early Cenozoic time (specifically, from 135 to 43 m.y. ago), the rate of convergence between the Farallon and North America plates generally increased, rising from roughly 25 mm/yr to a maximum of 170 mm/yr. As noted earlier, the first well documented extension in the Cordillera took place 51 m.y. ago, in northern Washington. This was a time of near-peak convergence rates. The extension was probably transtensional in nature, as convergence was oblique. At 42–43 m.y., the rate of relative convergence between the two plates dropped sharply to less than 100 mm/yr, and, according to Engebretson et al. (1982), it dropped still farther at 37 m.y., to 65 mm/yr. It was roughly 37 m.y. ago when the earliest, well documented extension in the Great Basin began. The onset of this extension followed by only 5 m.y. the end of transcurrent movement on the proto-San Andreas fault and corresponded to the beginning of a period of normal, as opposed to oblique, plate convergence (Carlson, 1982). Intra-arc and back-arc extension were thus initiated at a time of rapidly declining rates of plate convergence. Plate–plate coupling had probably begun to lose its efficacy.

The model of Engebretson and others also suggests (the details are not provided in Fig. 7; one must examine their original figure) that the rate of relative motion between the North American plate and the hot spot reference frame dropped at 37 m.y. ago to levels it had known only prior to 125 m.y. ago. It suggests, in addition that the age of the Farallon plate being subducted beneath North America at the latitude of the southern Great Basin had reached a near-minimum, 5–7 m.y., old. It must have been relatively thin and of low density owing to its youth and still-high temperature. It is doubtful that it had much rigidity throughout that reach close to the trench and perhaps it was this characteristic as much as any that reduced the strength of plate–plate coupling.

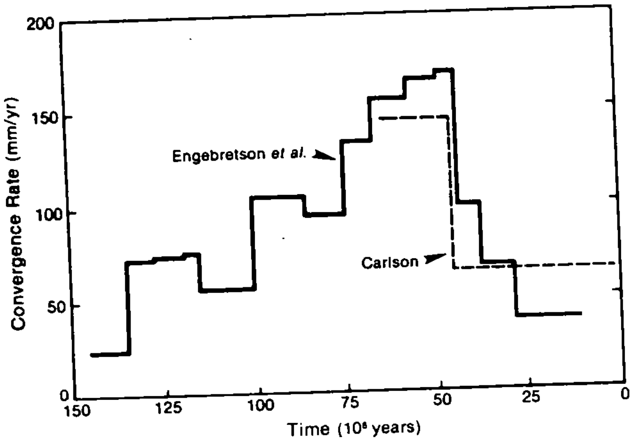


Fig. 7. Estimated plate convergence rates between the Farallon and North American plates vs. time (after Carlson, 1982 and Engebretson et al., 1982). Initial extension in the Great Basin began around 37 m.y. at

Figure 7 reveals no sign of a new convergence rate phenomena associated with events in the critical period 22–18 m.y., the time in which the calc-alkaline volcanic arc shrank geographically from a cluster of widely distributed, broad volcanic centers to the well-proportioned arc seen in Fig. 2. Nor are there any events of note during this period in the record of North American-hot spot reference frame convergence. The abrupt narrowing and increased coherence of the continental volcanic arc would thus appear to be a product of something other than change in the absolute or relative motion of the North American plate. Perhaps it reflects a stabilization of the distance of the spreading ocean ridge from the trench, a phenomenon that would have the effect of stabilizing the age of the oceanic lithosphere at the trench and, as a result, a fixation of the zone of magma generation in the subducting slab, was responsible.

Evidence from the arc

In attempting to quantify evidence for transgression and regression of subduction-related calc-alkaline magmatism in the southwestern United States, Coney and Reynolds (1977) made the assumption that the magmas had been generated at a more or less fixed depth on (or in) the subducting Farallon plate. In their model, the distance of magmatic activity from the trench is used to calculate the dip of the down-going plate. They interpreted the record as reflecting a flattening of dip from 80 to 50 m.y., a period of decreasing age of the oceanic lithosphere at the trench and consequently an increasing buoyancy for the downgoing slab. It was followed by a steepening after 40 m.y. The time–distance plot of part of Coney and Reynolds' compilation is shown in Fig. 8A, where the data have been replotted at a different scale and enclosed by a new bounding envelope. Also shown are similar data for a large region to the south (Fig. 8B). Both data sets as rendered here are limited to the last 50 m.y. of the geologic record, the time from latest Laramide orogeny to the present. The horizontal lines drawn across both diagrams identify the period 22–18 m.y., the time of development of the narrowed continental volcanic arc on the west side of the Great Basin. Over a wide range of latitudes south of the Great Basin, a west-southwestward component of seeming regression of the eastern margin of the volcanic arc was underway. At the latitudes of northern Mexico (Fig. 8B), it was a phase of very rapid regression in the period 22–18 m.y. and at the latitudes of the southwestern United States (Fig. 8A), the trenchward retreat of the eastern margin of the arc appears to have been accelerating rapidly. Farther north f the Great Basin, a far more complex pattern of change was evolving. From 44 to 2 m.y. ago the south edge of a broad calc-alkaline arc stretching across the entire width of the Great Basin was advancing southward (Cross and Pilger, 1978). Although its southwestern edge migrated west-southwestward from 41 to 22 m.y. ago, its eastern edge, that part equivalent to the bounding envelope on the right of the data shown in Figs. 8A and B, remained firmly fixed at the eastern margin of the Great Basin. There was no systematic or continuous westward or west-southwestward arc regression in the Great Basin prior to 22 m.y., rather, a very abrupt and discontinuous shift of the arc's eastern margin between 22 and 18 m.y. This shift was accompanied by a relative lull in volcanic activity. Later, motion of the trailing edge of the narrowed arc consisted of north-northwestward migration of its southern terminus in consonance with migration of the Mendocino triple junction (Snyder et al., 1976). It was during this later time (post 18 m.y.) that extensional spreading in the Great Basin was, by simple definition alone, back-arc spreading. As noted earlier, the direction of this extension was west-southwest, normal to the arc and trench.

The apparent westward retreat of the eastern margin of that part of the arc south of the Great Basin and represented by the time–distance plots in Fig. 8 has been interpreted as reflecting a progressive steepening of the downgoing Farallon plate. It followed a significant shoaling with which the compressional stresses of the Sevier–Laramide orogenies were associated (Coney and Reynolds, 1977; Dickinson and Snyder, 1978; Lipman, 1980; Eaton, 1982).

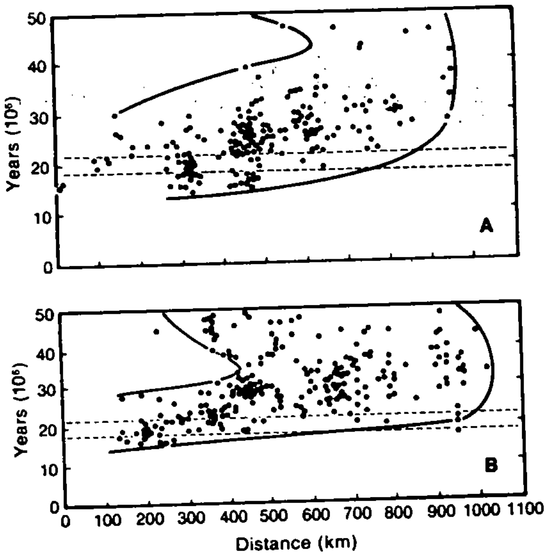


Fig. 8. Distance of calc-alkaline magmatism from the western margin of the North American plate vs. time; (A) at the latitudes of the southwestern United States and (B) in northwest Mexico (after Coney and Reynolds, 1977 and Damon et al., 1981). Continental arc volcanism was still advancing eastward (inland) during the period 50–40 m.y. and the state of stress in the North American plate was compressional. From 40 to approximately 22 m.y. ago it began a seeming slow regression westward. After 22 m.y. ago the regression was rapid and, at the latitudes of Mexico, extinction of arc volcanism was nearly instantaneous

Re-examination of these data, especially the more recently published compilation of Fig. 8B, suggests that this interpretation may be incorrect. The average slope of the "regressive" limb of the bounding curve in Fig. 8B suggests an extremely rapid retreat. If, as Coney and Reynolds (1977) maintained, this represents a westward retreat of a zone of magma generation at a depth of 150 km or so on the Farallon plate, then the angular rate of downward rotation of the plate would have had a value improbably high for a young and buoyant Farallon plate under the influence of a gravity-driven torque, even at anomalously elevated temperatures in the asthenosphere.

If the supposed regression in Fig. 8 is *not* an expression of a steepening of the subducting Farallon plate, what might it reflect? The data for Mexico (Fig. 8B) suggest that the waning phase of calc-alkaline volcanism was nearly synchronous across the region and that it "flamed-out" nearly everywhere at approximately the same time. An alternative interpretation of these data is that they reflect a singular convective overturn of the mantle, a buoyant rise and horizontal spreading of a thermal diapir of the sort defined by the model of Toksöz and Hsui (1978). Such a diapir might spread laterally beneath the base of the lithosphere, driving lithospheric extension in the process, finally dying at the same time nearly everywhere as it cooled by conductive and convective heat loss.

A diapir of this sort is created only after a long period of subduction, e.g. 25 m.y. or more, but the history of continuous subduction of the Farallon plate beneath western North America in Mesozoic and Cenozoic time is far more than adequate, regardless of whether convergence was oblique or normal. The problem with this hypothesis is that by 40 m.y. ago the dip of the Farallon plate apparently was much lower than the one modelled by Toksöz and Hsui (1978) and, in consequence, the diapir, if one was actually created in this manner by hydrodynamically-forced convection in the asthenosphere, should have risen much farther inland. This alternative, then, does not appear to help us with the dilemma posed by the observational data. Perhaps the source region of the downgoing slab became fully depleted or spent, incapable of further production of intermediate magmas, if that was the locus of their origin.

The abrupt westward displacement and narrowing of the volcanic arc at the latitudes of the Great Basin might be thought to represent some form of arc jump. Perhaps it was associated with a new phase of subduction, one involving a different part of the Farallon plate, one plunging downward at a steeper dip, the cause unexplained. If this were the case, however, it is unreflected in the record of relative plate convergence (Fig. 7). More troubling, it would have necessarily involved a young, buoyant downgoing slab, a seemingly unlikely candidate for steep subduction. We are thus left with a paradox. The observational data happened appear to be unassailable. The arc narrowed between 22 and 18 m.y., transforming the region of the Great Basin from one of intra-arc spreading to one of back-arc spreading. The change was accompanied by a relative lull in volcanism in the Great Basin, after which bimodal volcanism began.

Much of the crust of the western part of the Great Basin and the region west of it is believed to represent accreted fragments of other plates, some far-travelled from their locus of origin. Their transportation and docking, much of which preceded the extensional episodes described here, does not appear to have exerted an effect on the extensional structures or its related volcanism.

SUMMARY

Sometime between 22 and 18 m.y. ago, the Great Basin and the region to its north came to lie, at first fully, and then only partially, behind a narrow calc-alkaline volcanic arc. Extensional spreading took place in a west-southwest direction, normal to the arc and trench. The south end of this arc then began to migrate steadily north-northwestward, exposing part of the former back-arc region to the "broadside" of a growing transform fault system at the western edge of the continent. The south end of the arc had moved but a short distance northwestward by 10 m.y. ago, lying at that time not far south of the latitude of 37°N, in southern Nevada. After 10 m.y. ago, however, it began to move northwestward more rapidly. From a period that began sometime between 10 and 6.5 m.y. ago and continued to the present, two highly significant tectonic phenomena occurred:

(1) Extensional faulting waned at all latitudes south of 36°N, behind the San Andreas fault, i.e., in the Sonoran Desert section of the Basin and Range province (in southeastern California and southern Arizona). This tends to cast doubt on the efficacy of the Atwater model at these latitudes.

(2) The direction of extension in the Great Basin changed from west-southwest to west-northwest, apparently reflecting onset of the influence of the Atwater model but *only* north of the junction between the Garlock and San Andreas faults.

The structural style commonly referred to in the literature as basin-range faulting apparently began to develop 17 m.y. or so ago, but the present day structural grain in the Great Basin does not reflect that faulting. It is the product of the younger of two episodes of back-arc and back-transform extension, one in which the spreading direction was west-northwest. As basin-range faulting developed in the Great Basin, simple lithospheric rifting of the crust farther north elicited a voluminous outpouring of basalt in the Columbia River plateau subregion. The entire region, Great Basin and Columbia Plateau alike, lay behind a calc-alkaline volcanic arc at that time, and hence its tectonism was an expression of continental back-arc spreading.

Several of the characteristics created during this episode of spreading are pre-
in the geologic record. Some appear to be preserved also in the geophysical

record. The latter have probably undergone some degree of modification during late time, though. Because later events were of a qualitatively similar nature, however, i is suggested that they did not alter the geophysical record in any fundamental way but may, instead, have actually enhanced it. Thus we see preserved in the geophysical record evidence of crustal spreading about a middle Miocene axis of symmetry This axis is marked by a major swarm of basaltic dikes and, in the south, by a moderate degree of gravity symmetry. The crust and lithosphere are thin, heat flow is anomalously high, and upper mantle and crustal seismic wave velocities are relatively low. Evidence of Oligocene and Miocene normal faulting, the resultant structures trending northwest and preserved in the interior of today's uplifted, north-northeast-trending ranges, reveals the direction of an associated minimum principal stress that was orthogonal to the arc and trench of that time.

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