# Geophysical and Geological Characteristics of the Crust of the Basin and Range Province

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GORDON P. EATON
*U.S. Geological Survey*

## INTRODUCTION

Selected geophysical and geological characteristics of the Basin and Range province of the western United States are examined here. They represent the effects of nearly 30 million years (m.y.) of crustal extension, preceded by subduction-related compressional deformation and magma genesis. The effects of this compressional regime appear to have exerted a significant influence on the mechanics and locus of late Cenozoic extension.

In its present state, the crust of the Basin and Range province is atypical of most continental crust, being appreciably thinner, warmer, and more highly fractured and permeable. Some of what is seen or measured geophysically at the surface in the province is influenced by processes or phenomena in the shallow crust. Faulting, for example, appears to be a condition of the upper 10 to 15 km, and some of the phenomena associated with it serve to screen or distort information from deeper levels. As an example, convection of groundwater in the fractured, shallow crust complicates the determination of heat flow from deeper crustal levels and is apparently also capable of creating regional geoelectric anomalies.

One of the remarkable features of the Basin and Range province is the broadly distributed nature of its extension. Most other regions of continental extension consist of singular or branching large rifts, like the contemporary, neighboring Rio Grande rift in New Mexico or the Rhine graben of Europe. Although marginal seas bordering continents, like the Seas of Okhotsk or Japan [both of which are at present inactive with regard to crustal spreading but nevertheless have high heat flow (Karig, 1974)], have a similarity to the Basin and Range province, they have significant contrasts as well. The same is true for actively spreading intraoceanic, back-arc basins, like the Bonin, Mariana, New Hebrides, and Lau-Havre troughs.

The Great Basin consists of continental crust, stands nearly 2 km above sea level, has a very broad, but well-developed, bilateral symmetry in its geophysical characteristics (Eaton *et al.*, 1978), and has a perimeter marked by active crustal seismicity and young volcanism. In contrast, marginal seas near the Asian continent and intraoceanic spreading basins consist of oceanic crust, the upper surface of which is submerged *below* sea level. They show little or no distributed or peripheral crustal seismicity (see Barazangi and Dorman, 1969). No well-developed or extensive pattern of symmetry nor any persistent, peripheral volcanism has been described for them. Although paired linear magnetic anomalies exist locally, they are relatively incoherent compared with those at spreading ocean ridges, and their amplitudes are much smaller.

Extension in the Great Basin *began* within and behind an andesitic volcanic arc, typical of where back-arc basins form and exist. Although much of that arc is now dead, extensional deformation continues. Finally, active continental volcanic arcs persist today in the Cordillera of Central and South America but have no Basin–Range structure behind them. According to Uyeda and Kanamori (1979), this is probably the result of a compressional state of stress in the overriding American plate; it follows from a relatively low angle of subduction and rapid rate of plate convergence. A fundamental distinction is made between what they term a "Chilean," or convergence-related compressional mode, and a "Marianas," or convergence-related extensional mode, of subduction.

Interpretations of crustal thickness of different parts of the western United States vary somewhat (compare Prodehl, 1970; Warren and Healey, 1973; and Smith, 1978), but it appears that the Great Basin crust is no more than 25–30 km thick. The crust of the rest of the Basin and Range province ranges from 20 to 30 km thick. By contrast, the crust of the unextended Colorado Plateaus and Great Plains provinces to the east is 40 and 50 km thick, respectively. Upper-mantle velocities beneath the Great Basin are anomalously low (Smith, 1978) and the lithosphere is anomalously thin (≤65 km). It is one of the few *continental* areas of the globe beneath which a low-velocity zone is known to be present in the mantle.

Not all of the Basin and Range province is tectonically active at present. The Great Basin section is extending, the western Mojave Desert is in horizontal dextral shear, and the rest of the province is tectonically inactive and has been for several millions of years.

## PRESENT AND PAST CRUSTAL EXTENSION

Normal faults, the surface evidence of crustal extension, are widespread in the western United States. Their distribution in the Basin and Range province is broad and pervasive, but in the Rio Grande rift [see Figures 9.5(a) and 9.6 for identification of the principal geographic features mentioned in the text] they define a narrow band typical of continental grabens. On the basis of the geomorphology of the faulted region, crustal spreading is a continuing process, still active in some areas, but inactive in others. The Sonoran Desert section of the Basin and Range province has a general elevation and topography characteristic of profound erosion and tectonic inactivity (Fenneman, 1931; Lobeck, 1939; Hunt, 1967). In sharp contrast, Quaternary faults of Nevada are amenable to detailed age classification based on the degree of erosion, the slope angle of the scarp, and the width of the crestal break in slope (Slemmons, 1967; Wallace, 1977).

Figure 9.1 is a map of the contemporary fault-displacement field of the United States. Active normal-fault displacements are restricted almost exclusively to the Great Basin and Rio Grande rift. Local, scattered thrusting events in the West are seen to the northwest, east, and southwest, but displacements of this type are found mostly in the middle and eastern United States.

The azimuths of the *P* (pressure) and *T* (tension) axes cannot be equated with those of the principal stress directions, $\sigma_1$ and $\sigma_3$, with certainty. McKenzie (1969) demonstrated that in the general case of triaxial stress the only restriction on the spatial relation between the two is that the stress directions must lie in quadrants containing the related displacement axes but may diverge from them by many tens of degrees. He argued, as did Brace (1972a), that because most crustal earthquakes occur on pre-existing fault planes, they are unrepresentative of the ideal situation in which initial failure takes place in virgin, homogeneous material. Only in this case will stress directions necessarily bear direct correspondence to displacement. Because of the long history of normal faulting in the Great Basin we may assume that those earthquakes whose fault-plane solutions are shown in Figure 9.1 took place on pre-existing faults and that the orientation of these faults with respect to the stress field determined both the direction and nature of displacement.

To investigate the potential discrepancy between displacement and stress direction, *in situ* measured directions of $\sigma_1$ for the eastern United States were plotted in Figure 9.1. Together, the two kinds of data suggest that on a regional basis *P* and $\sigma_1$ have much the same orientation—northeast to east.

In a related plot (Figure 9.2), a comparison is made between locally paired, measured directions of *T* and $\sigma_3$ in the West. Most of the directions of $\sigma_3$ determined from *in situ* stress measurements are not shown in Figure 9.1 because of the density of other data. They were taken from sources listed in the caption. The data selection process required only that the earthquake and *in situ* stress measurement locality in each pair be within 200 km of one another.

Figure 9.2 indicates moderately good agreement between extensional stresses and displacement directions, despite McKenzie's (1969) well-reasoned caution. On this limited basis we may assume that the displacement field of the normal faults is a measure of the direction of current extension. On average, it is east–west. The strike of these faults in relation to that of the displacements is, in much of the region, orthogonal or nearly so, hence the spreading is generally in consonance with that of continental spreading elsewhere in the world (Ranalli and Tanczyk, 1975) and with that of seafloor spreading (Moore, 1973).

Figure 9.1 shows two other significant characteristics of the region of active continental spreading: (1) it stands high, having been uplifted to an elevation of 1100 m and

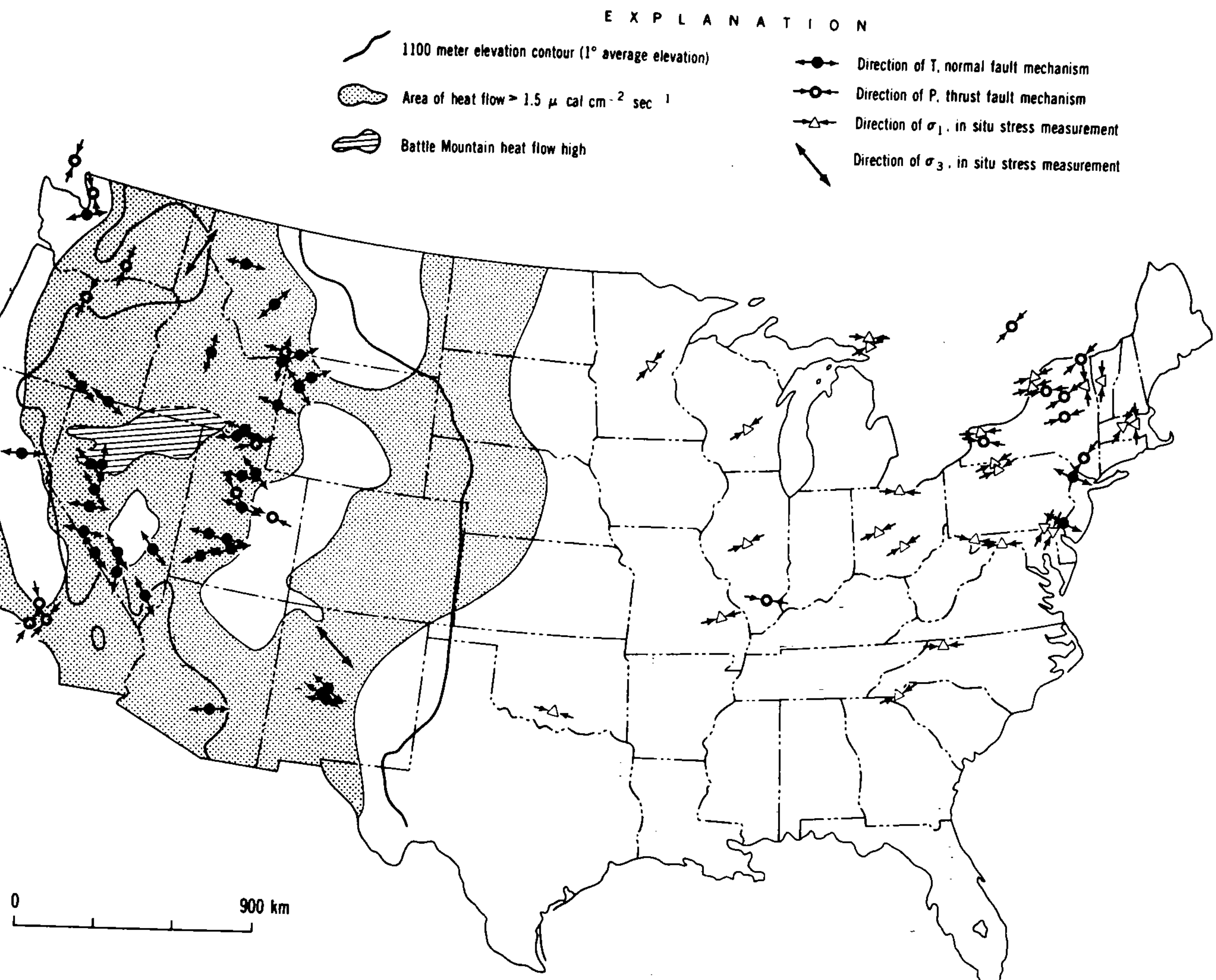


FIGURE 9.1 Horizontal components of fault displacements and maximum principal stress directions. Compiled from data in Couch *et al.* (1976), Haimson (1977), Jaksha *et al.* (1977), Lachenbruch and Sass (1977), Malone *et al.* (1975), Mott (1976), Rogers (1977), Sanford *et al.* (1977), Sbar and Sykes (1973, 1977), Smith (1978), Smith and Lindh (1978), and Strange and Woollard (1964). Thrust faulting and horizontal compression dominate in the eastern United States, normal faulting and extension, in the west.

more; and (2) it is hot; heat-flow values nearly everywhere are greater than 1.1 $\mu$cal cm$^{-2}$ sec$^{-1}$, the average value for stable continents. We return to these observations below.

Figure 9.3(a) is a map of most of the major faults in western North America that have steep dips, regardless of age. The longer ones are strike-slip faults, the shorter ones, normal faults. The latter are much more abundant and widespread. *Active* strike-slip faults that have major displacements of several tens of kilometers are today restricted largely to a 500-km-wide corridor paralleling the west coast of the United States. Those farther inland, in Canada, have not been active in late Cenozoic time—the period of crustal spreading with which we are concerned.

Because the azimuthal range of these faults is great, it is difficult to see well-defined patterns from which we might draw immediate conclusions about variations in spreading. To do this one must filter the data. Figure 9.3(b) is temporally filtered, showing only those faults active in the past 10–15 m.y. for which some Quaternary movement is suspected. It distinguishes the Great Basin section of the Basin and Range province from the Sonoran Desert section. It also clearly defines the Rio Grande rift. The fault data, like the geomorphic data, suggest that extensional spreading in the Sonoran Desert region ended some time ago. Eberly and Stanley (1978) alleged that block faulting began to wane about 10.5 m.y. ago in the Sonoran Desert and that throughgoing drainage of the Gila River was well established by 6.0 m.y. ago. The cessation of crustal spreading at these latitudes may be related to the opening of the Gulf of California on the west. If so, it represents the continental equivalent of a kind of ridge jump—from distributed extension on the east to narrowly focused extension on the west.

Figure 9.4 shows that the orientation of basin ranges in

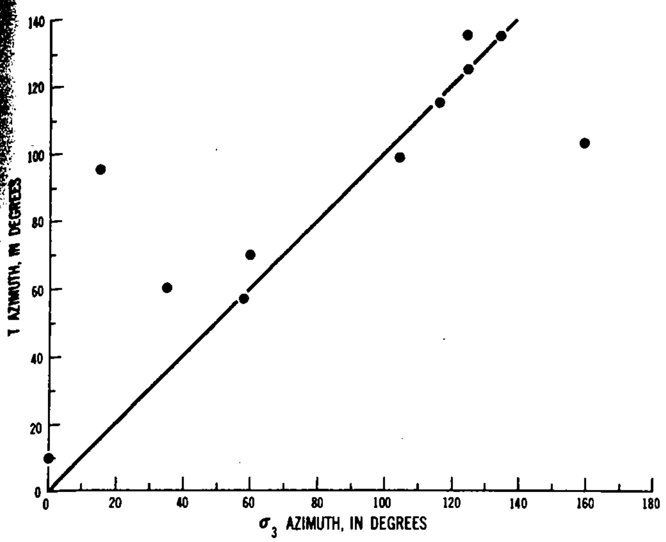


FIGURE 9.2 Azimuthal relations between horizontal components of minimum principal stress ($\sigma_3$) and *T* axes of normal-fault earthquake mechanisms. Localities in each set of paired measurements (stress measurement and fault displacement) were less than 200 km apart. Data from sources listed in caption of Figure 9.1. Solid line defines locus of stress directions and fault displacements having the same horizontal direction.

the Sonoran Desert is significantly different from that of the Great Basin and Rio Grande rift. If the topography is a reflection of block faulting, this difference, in conjunction with the data of Figures 9.1 and 9.2, suggests that extensional stresses had a markedly different orientation in the Sonoran Desert region than they do in the Great Basin today. They were generally southwest to west–southwest, in contrast to later east–west to west–northwest extension in the rest of the western United States.

## THE THERMAL REGIME

Heat lost at the surface of the earth may reflect any of several phenomena: (1) conduction of heat into and through the lithosphere, the base of which may approximate an isothermal surface marking the upper bound of a region of partial melt in the mantle; (2) upward mass transport of heat through the lithosphere by ascending magma, a penetrative form of convection; (3) production of heat in the crust itself by radioactive decay of U, Th, and $^{40}$K and/or by conversion from mechanical energy, as in faulting; (4) transient cooling of the lithosphere; and (5) convection of heat in the shallow crust by circulating groundwaters. The last phenomenon is especially troublesome in permeable rocks. It can greatly disturb the heat flow associated with deeper crustal regimes.

Many heat-flow measurements have been made in the extended regions of the western United States (Lachenbruch and Sass, 1977). The average value is roughly twice that of the tectonically stable interior and eastern part of the North American continent, after accounting for crustal radioactivity. Although heat-flow values in the Pacific Northwest are also higher than those in the continental interior, the reduced values (those for which the contribution of heat from heat production within the crust has been subtracted) appear to be somewhat lower than those of the Great Basin (compare Figures 1 and 14 of Lachenbruch and Sass, 1977). At present, there are an insufficient number of heat-flow measurements in the Sonoran Desert to be certain of what its thermal regime is, but thermal lag may keep crustal temperatures moderately high if extension there ceased as recently as 7 m.y. ago. The Rio Grande rift has heat-flow values as high as those of the Great Basin (Reiter *et al.*, 1975), and they decay systematically outward, much as they do at spreading ocean ridges.

Lachenbruch and Sass (1977; 1978) examined the thermal regime of the U.S. continental crust, evaluated various factors affecting crustal temperature, and proposed a model in which the high, but variable, heat flow in the Great Basin is accounted for by (1) regionally distributed basaltic intrusions of the lithosphere, accompanied by lithospheric thinning and possibly magmatic underplating and (2) spatial variations of the extensional strain rate that controls access of this basalt to the lithosphere. The lithosphere is pulled apart and basalt wells up into it from the asthenosphere.

Basaltic intrusions in this model are viewed as vertical, dike-like or bleb-like bodies that accommodate extension while preserving the continuity of the lithosphere. Instead of pulling open at a single place, as it does at most spreading ridges, the lithosphere pulls open in a broadly distributed fashion at a great many places. Spatial and temporal variations in rates of extension control the intensity of the upward flux of basalt and, hence, that of surface heat flow. The model is mechanically somewhat like one proposed by Thompson (1959; 1966), who postulated a Great Basin lithosphere dilated by intrusive dikes. It provides a rationale for the local occurrence of shallow magmatic systems that convey heat from the asthenosphere to the surface by mass transport.

## CRUSTAL MAGMAS

The Mesozoic and Cenozoic history of the western United States is one of repeated intrusion of the shallow crust by magmas. Many broke through to the surface, creating large volcanic fields. Figure 9.5 shows the spatial distribution of igneous rocks of post-Paleozoic age. From earliest Mesozoic through early Miocene time, rocks of intermediate, calc-alkaline composition were formed, probably as a result of subduction of the Farallon plate beneath the region (Lipman *et al.*, 1972; see Chapter 14). From Miocene time on, however, they were dominantly bimodal (basaltic and rhyolitic) in composition, reflecting extension of the lithosphere that followed cessation of subduction (Christiansen and Lipman, 1972; Christiansen and McKee, 1978). Figure 9.5(a) shows the regional limits of Mesozoic igneous rocks and individual areas of outcrop

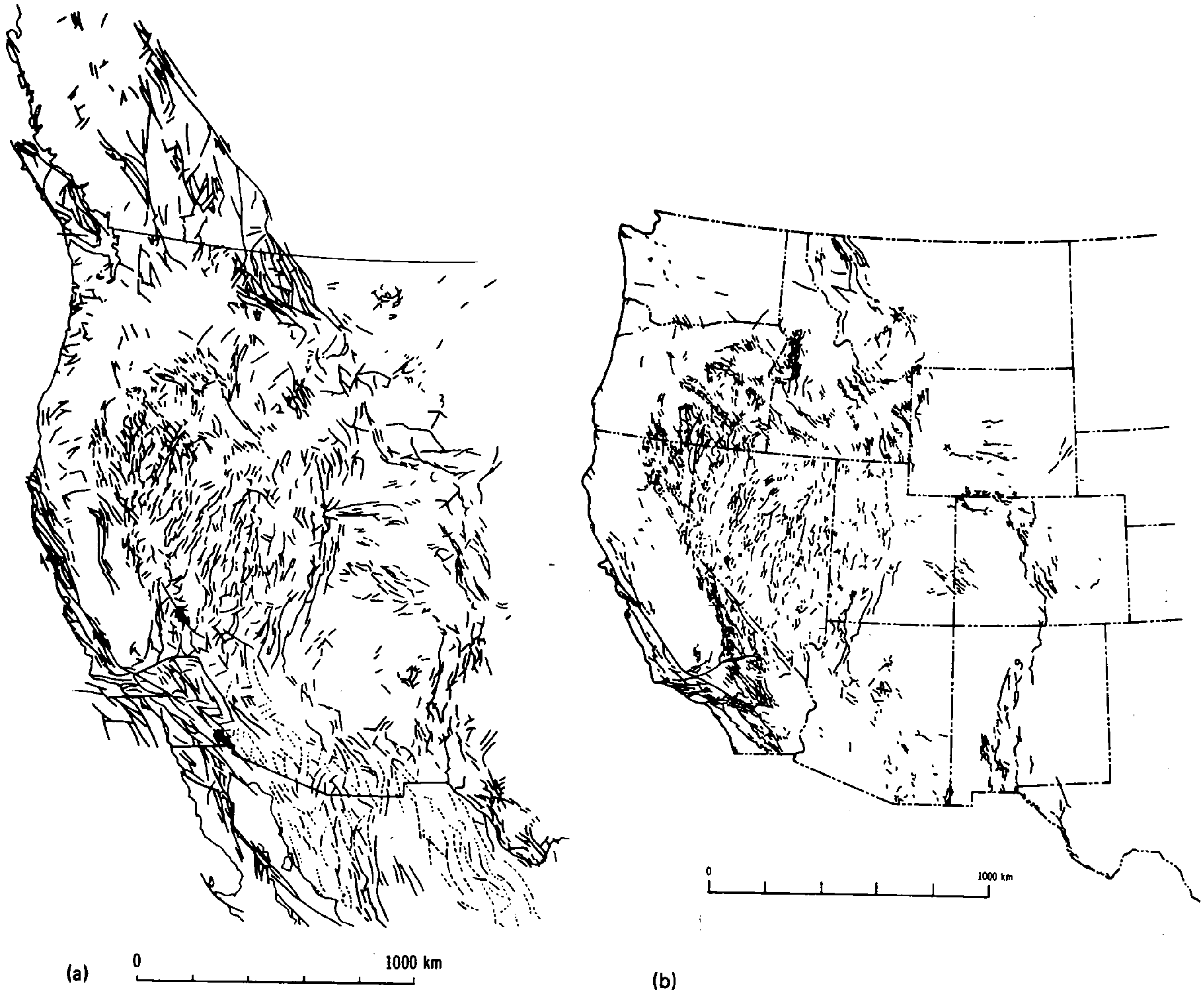


FIGURE 9.3 (a) Steeply dipping faults in western North America (from King and Beikman, 1974; King, 1969; and Cohee, 1961). Faults are shown without regard to age of initiation or last movement. Longer ones are generally strike-slip faults; shorter ones, normal faults. Dashed lines in southern part of map mark long axes of block ranges whose boundary faults are obscured by upper Cenozoic alluvium. (b) Faults active in the past 10–15 m.y. for which some Quaternary movement is suspected (after Howard *et al.*, 1978).

of Paleocene through Oligocene igneous rocks. Clearly, the Great Basin had abundant igneous activity in Mesozoic and early Tertiary time.

Igneous rocks of Miocene, Pliocene, and Quaternary age are shown in Figures 9.5(b), 9.5(c), and 9.5(d). An outward restriction of magmatic activity with time toward the margins of the Great Basin is illustrated by these three diagrams. Areas immediately adjacent to the Great Basin had a generally similar history of magmatism, as the four illustrations show. Magmatism alone does not set the Great Basin or Rio Grande rift apart from the rest of the region.

Figure 9.6 shows the spatial distribution of known Cenozoic plutons and ash flow-related calderas. The calderas represent large magma chambers at crustal levels shallow enough to permit catastrophic eruptions of great volumes of pyroclastic material. Such magma chambers are in the upper 5 to 10 km of the crust and give up large amounts of heat there. More than 50 of these features have been identified thus far in the Great Basin and the region immediately north of it. Another 22 are seen in a broad north–south corridor that includes the Rio Grande rift and its western environs.

The distribution of these shallow crustal heat sources coincides generally with that of normal faults active during the past 10 m.y. to 15 m.y. [Figure 9.3(b)]. Although the former span the entire Cenozoic Era, they are spatially distributed more or less like the regions of young extensional faulting, suggesting some sort of genetic relationship. Mackin (1960) proposed direct cause and effect:

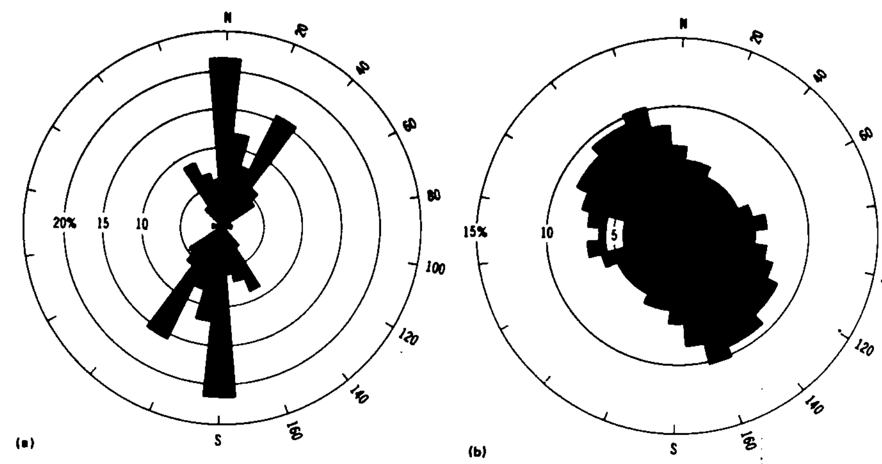


FIGURE 9.4 Rose diagrams of the orientations of range fronts. (a) 545 range-front segments, 50 km long, in the Great Basin section of the Basin and Range province and the Rio Grande rift (Nevada, Utah, and northern and central New Mexico); (b) 548 range-front segments, 50 km long, in the Mojave Desert, Sonoran Desert, and Mexican Highlands section of the Basin and Range province (southeastern California, southern Arizona, and southern New Mexico). Note difference in maximum frequency values of the outermost rings of the two diagrams and the greater scatter of range-front directions in (b).

withdrawal of large volumes of magma to the surface allowed collapse and spreading of the overlying slab. I do not subscribe to this view.

## HYDROTHERMAL CONVECTION

Heated groundwaters appear at the surface as hot springs and warm springs, the surface manifestations of convecting hydrothermal systems. In some places they are related directly to shallow bodies of magma, as at Yellowstone National Park (Eaton *et al.*, 1975; Smith and Shaw, 1975). They mark sites where the regional conductive heat flow is perturbed. The areal distribution of thermal springs is shown in Figure 9.7(a), where boundaries have been drawn about regional clusters of springs. The guideline followed in drawing the boundaries required that individual members of a cluster should be no more than 100 km from one of its neighbors in the same cluster.

The hydrothermal systems in Figure 9.7(a) are active today, yet, with two exceptions, their distribution seems generally to mimic that of extensional faults of the last 15 m.y. [Figure 9.3(b)]. Areas of prominent exception are (1) a broad corridor along the San Andreas Fault, in coastal California [see Lachenbruch and Sass (1973) for an explanation of the high-heat flow there]; and (2) a corridor along the Cascade Range of Washington, Oregon, and northern California, a belt of active, subduction-related andesitic volcanoes. These hot springs are thus associated with regions of (1) crustal extension and basaltic volcanism, (2) active calc-alkaline volcanism, and (3) transform plate motion. Hydrothermal systems in the extensional regime doubtless use the high regional permeability provided by faults and related fractures.

Fossil equivalents of these hydrothermal systems are shown in Figure 9.7(b). They are sites of Tertiary and Mesozoic (and some still older) epigenetic ore deposits, many of hydrothermal origin. To a limited extent their distribution may reflect accidents of erosion. In some regions, however, such as the Colorado Plateaus, subsurface data are abundant enough to indicate a general scarcity, if not absence, of such deposits on a regional scale, both now and in the past. The boundaries drawn in Figure 9.7(a) are repeated in Figure 9.7(b) to illustrate the fact that although some of the older hydrothermal systems are in locations quite different from those active today, a great many are in the same areas. The Great Basin hot springs, those north of the Snake River Plain (in central Idaho and southwestern Montana), and those in and near the Rio Grande rift (in Colorado and New Mexico) have ancestors in very much the same places, dating back tens of millions of years. Clearly, hydrothermal convection has been a feature of the shallow crust in these areas during times of both crustal compression and extension, whether of subduction-related or postsubduction origin.

What can be said of the exceptions, those clusters of springs or areas of spring-absence identified in Figure 9.7(b) by letters? Significantly, the most numerous exceptions are older deposits and hydrothermal systems that have not persisted to the present, rather than younger ones that sprang up anew, as they have in virgin areas B, G, and K. Deposits in areas C and D occur largely in oceanic crust accreted to the continent.

Areas E and F are of special interest because they tell us something about significant contrasts within the Basin and Range province, such as the strong difference between the Great Basin and Sonoran Desert sections. Hydrothermal convection is dying out in the Sonoran Desert region, just as crustal extension has.

Figures 9.5–9.7 show that the Great Basin and Rio Grande rift have been perturbed thermally for a very long time and that crustal temperatures were high in these areas long before extension began. High temperature probably influenced extensional deformation from its outset; at the very least it thermally weakened the crust and thereby helped to determine where extension and thinning would ultimately take place.

## THE DISTRIBUTION OF EARTHQUAKES

The spatial distribution of earthquakes in the western United States is shown in Figure 9.8. Many are concentrated in a broad corridor parallel to the San Andreas Fault

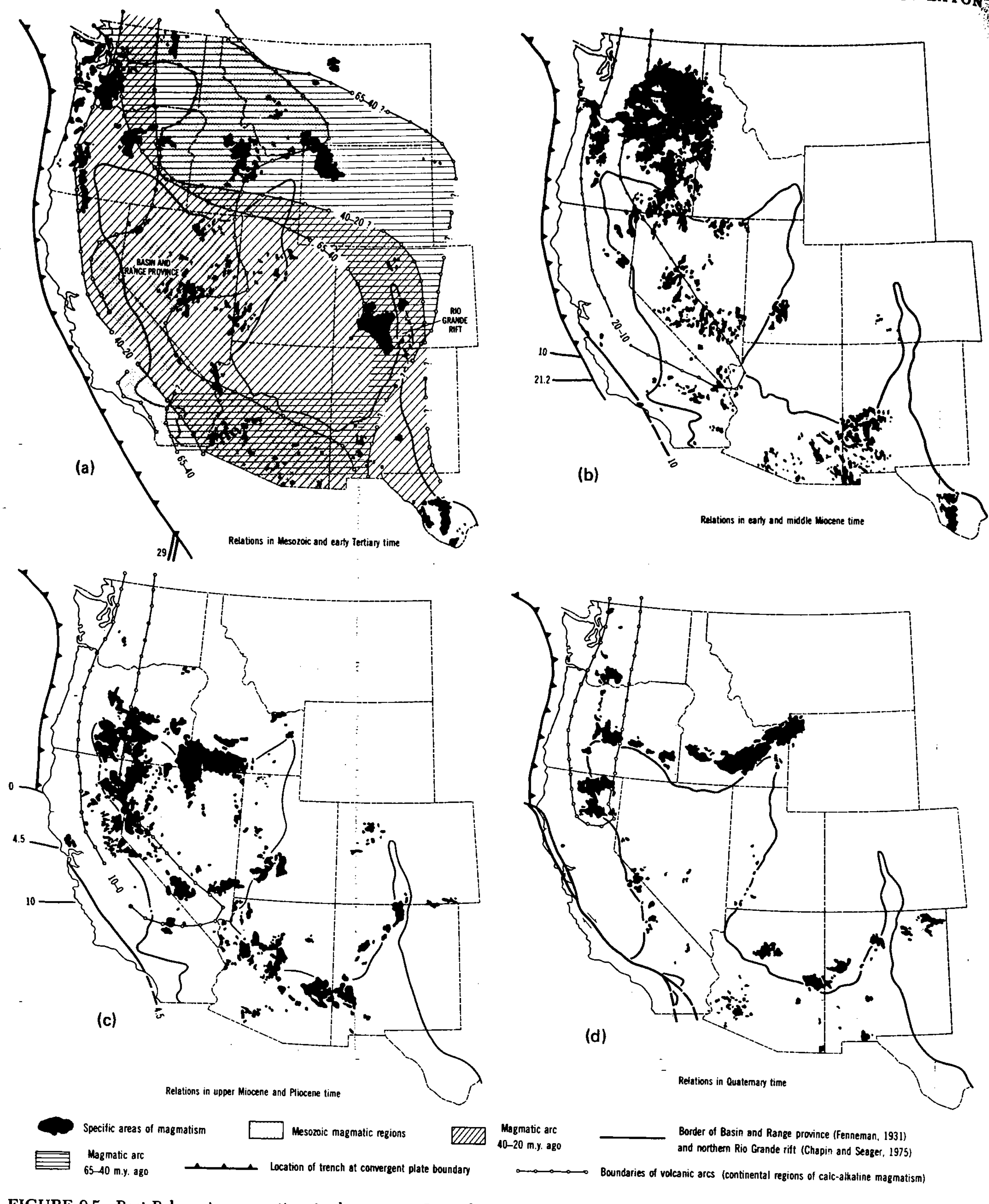


FIGURE 9.5 Post-Paleozoic magmatism in the western United States (data from King and Beikman, 1974; Snyder *et al.*, 1976). The heavy line is the present-day border of the Basin and Range province, after Fenneman (1931), with addition of the northern Rio Grande rift from Chapin and Seager (1975).

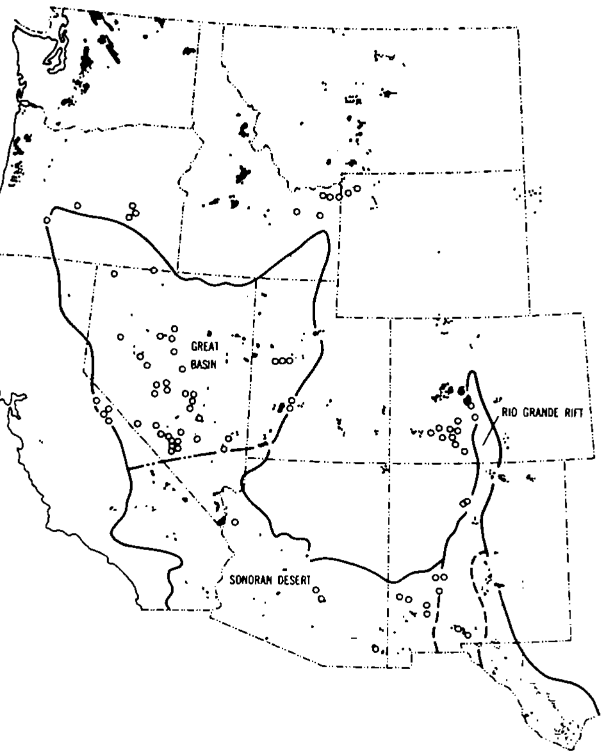


FIGURE 9.6 Shallow Cenozoic intrusive masses and ash-flow related calderas in the western United States (data from King and Beikman, 1974; E. H. McKee, USGS, unpublished data; T. A. Steven, USGS, unpublished data; and Eaton *et al.*, 1978). Small irregular black areas and tiny dots are exposed intrusive rocks; open circles are sites of ash-flow related calderas. The calderas represent large bodies of magma at crustal levels shallow enough for their roofs to founder. They mark sites where shallow local crustal temperatures were, or still are, anomalously very high. The heavy line is Fenneman's (1931) boundary of the Basin and Range province, plus Chapin and Seager's (1975) boundary for the Rio Grande rift, the southern part of which is dashed. The dash–dot–dot line marks the midpoint of a steep, regional gravity gradient between the Great Basin and Sonoran Desert sections of the Basin and Range province (after Eaton *et al.*, 1978).

system in coastal California, but many are well inland, some as far as 1700 km from the coast. Atwater (1970) considered the San Andreas to be the key element in the transform boundary between the Pacific and North American plates. She viewed tectonic activity inland, at least to the eastern edge of the Great Basin, as reflecting deformation in a broad, soft zone functioning as part of that boundary.

The smoothly curving line in Figure 9.8 marks the inland edge of the most abundant earthquakes. Although Smith (1978) used a lower threshold magnitude for plotting California earthquakes, the line I have drawn is, for the most part, within California, hence represents a real boundary. This line approximates the *present* inboard limit of pronounced right-lateral shearing motion between the Pacific and North American plates. East of this line, strike-slip first motions in the southern Great Basin are left-lateral in displacement sense, with the active nodal plane striking west to southwest (see the compilation of Smith and Lindh, 1978).

Strike-slip motion has not been recorded or observed in the eastern Great Basin, but in west-central Nevada, in a zone 100 to 150 km wide east of the boundary shown in Figure 9.8, oblique-slip faults with dextral components of shear are observed. The nature of displacements has been confirmed by fault-plane solution, by geodetic measurement, and by direct observation in the field (Thompson and Burke, 1974). Slemmons (1967) observed strike-slip faults across the entire Nevada section of the Great Basin, but those east of the zone in question are of much smaller displacement, suggesting that the direct effects of dextral shear related to plate interaction die out inland in the western Great Basin. The total seismic-strain energy released in the eastern two thirds of the Great Basin is less, in the aggregate, than that related to strike-slip faulting in the curve-bounded region to the west (Ryall *et al.*, 1966; Crampin *et al.*, 1976).

Figure 9.9 shows statistical and geographical variations in the focal depths of earthquakes. Hypocenters are distributed from the near surface to a depth of 20 km but are seldom found at depths greater than 15 km. In the eastern part of the Great Basin (histograms A, D, G, and H) earthquakes are limited to the upper 15 km of the crust and concentrated in the upper 10 km, with some of the modal values (e.g., sites D and G) in the upper 5 km. Sites D and G also show a systematic decrease in the numbers of earthquakes downward. Western Great Basin earthquakes (histograms I through N) show a tendency toward somewhat greater depth, and at sites K and L there is a systematic *increase* downward, but, as just noted, this is an area of profound strike-slip faulting, hence the deformational regime is different.

These depth characteristics provide clues to the nature of crustal extension. Because sudden instabilities that create earthquakes in the shallow crust are probably related to stick-slip displacements on pre-existing faults (Brace, 1972a), one may interpret their depths as depths of instantaneous faulting. Extensional faulting accompanied by abrupt stress drop appears to continue in places scarcely deeper than 10 km. If the faults themselves continue to deeper levels their displacements are characterized by stable sliding or fault creep rather than stick-slip (Byerlee, 1968; Brace and Byerlee, 1968; Brace, 1972b).

Almost certainly the extensional faults do *not* continue to deeper levels. Thompson (1966) and Hamilton and Myers (1966) pointed out that if the normal faults in the Great Basin continued as planes having dips like those observed at the surface to depths where facing pairs intersected, the fault blocks could be no thicker than 10–15 km. Many students of the region, beginning with Longwell (1933; 1945), have described features of these faults that suggest they flatten with depth (see Moore, 1960; Mackin, 1960; Hamblin, 1965; Anderson, 1971; Wright and Troxel, 1973; and Proffett, 1977), leading to fault blocks even thinner than 10–15 km. In addition, regional-gravity data (Eaton *et al.*, 1978) indicate that the local mountain ranges

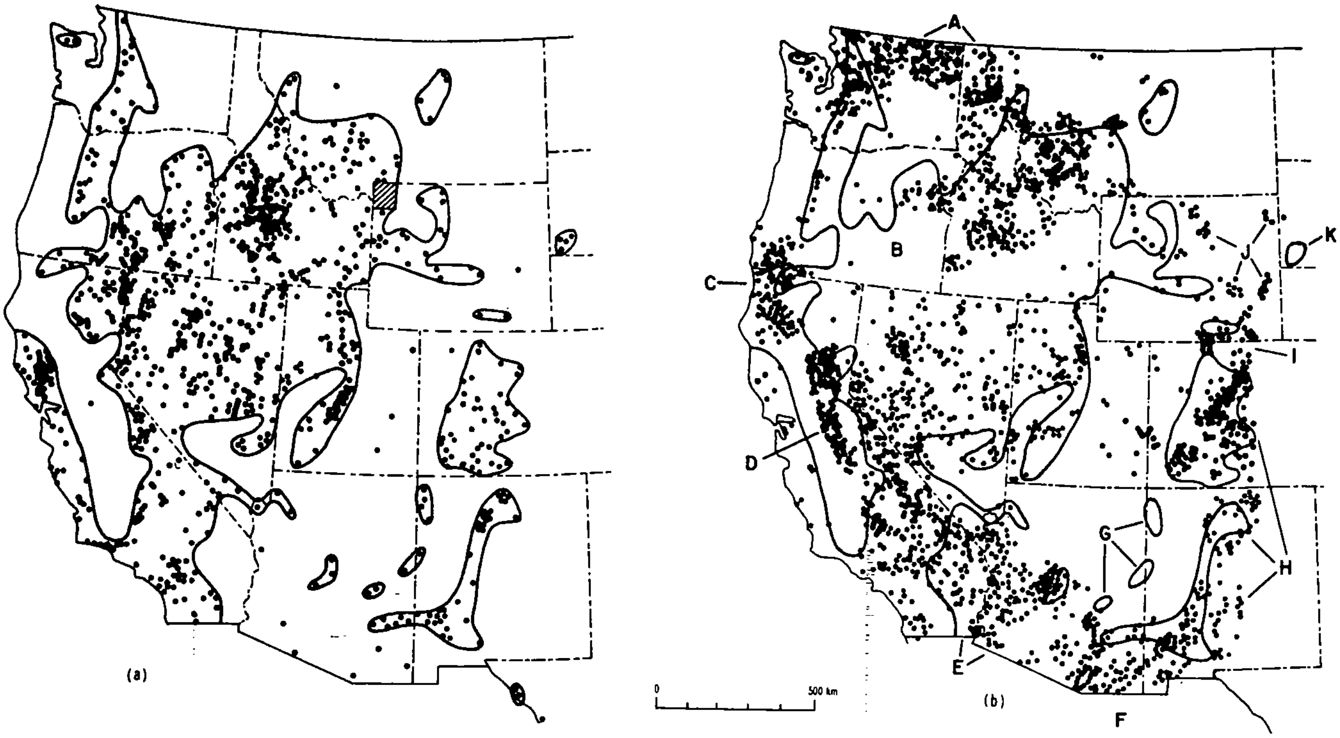


FIGURE 9.7 Active and fossil hydrothermal systems in the western United States. Data from Waring (1965) and Jerome and Cook (1967). (a) Active warm springs and hot springs the surface temperatures of which exceed the local mean annual air temperature by more than 8°C. Yellowstone National Park, the site of numerous springs, is cross hatched. Boundaries enclose local and regional clusters of these springs. (b) Sites of ore deposits, largely of epigenetic origin, many of which mark the location of older hydrothermal circulation systems. Boundaries from diagram on left are repeated on right to illustrate similarities and differences in areas of occurrence. Letters identify areas of principal difference between the two maps, some of which are discussed in text.

generally are not compensated isostatically, suggesting, in turn, that faults serving as surfaces of adjustment do not pass through the lithosphere. All of these data tend to suggest that the depth limit of earthquakes may be the depth limit of faulting. Stewart (1978) illustrated alternative interpretations of how such faults may terminate downward.

Figure 9.10 compares the aggregate earthquake-depth distribution for the entire region with Basin and Range widths. The two histograms are generally similar in form, both having intermediate values between 0 and 5 km, peaking between 5 and 10 km, and having relative low levels of occurrence for values greater than 20 km. Of the earthquakes, 97 percent occur in the upper 15 km of the crust, which in the Great Basin is the upper *half* of the crust. Of Basin and Range block widths, 88 percent are no wider than 20 km. Theoretical and experimental analyses of the depth and spacing of fractures formed in extension (both tensile fractures and extensional *shear* fractures) suggest that the two values should be similar, at least within an order of magnitude (Lachenbruch, 1961; Sowers, 1972).

In a mechanical analysis of block gliding in which horsts and grabens formed, Voight (1973) derived an approximate equation for the width of a block formed by extensional faulting of an initially continuous slab. It is $W = 2T(45° - \phi/2)$, where $W$ is the width of the block; $T$, the thickness of the faulted slab; and $\phi$, the coefficient of internal friction for effective stresses. Application of this equation to the Basin and Range province requires the assumption of a surface or zone of translatory sliding at the base of the fragmenting upper crust. It will be shown below that the Great Basin crust may have such a zone.

In order to solve Voight's equation we must have a value for $\phi$. Byerlee's (1968) experimental study of the brittle–ductile transition in rocks indicated that friction is independent of composition. Rocks under confining pressures of from 0 to 5.2 kilobars (the depth equivalent of 0 to 16.5 km) revealed a remarkably systematic variation between increasing normal stress ($\sigma$) and shear stress ($\tau$) for friction. The limiting slopes, $\tau/\sigma = \tan \phi$, of Byerlee's friction data curve are 36° and 46°. Substitution of these values in Voight's equation yields the following widths for fault blocks formed in this manner: for shallow crustal slabs initially 10 km thick, widths of 8.1–10.2 km; for slabs 15 km thick, 12.1–15.3 km; and for slabs 20 km thick, 16.2–20.4 km. These results are in good agreement with the observed Basin and Range block width–earthquake depth relation (Figure 9.10), hence Voight's model of extensional sliding may be judged to have potential relevance to an understanding of the mechanics of Basin–Range faulting.

Thompson (1959; 1966) was the first to suggest that extension in the deeper basin–range crust takes place via plastic stretching or injection of dikes. Hamilton and Myers (1966), Stewart (1971), and Proffett (1977) all accepted the first of these concepts, viewing Basin and Range structure as fragmentation of a shallow crustal slab riding on a plastically extending substratum. Lateral dilation of the lithosphere by magma from below is implicit in the thermomechanical model of Lachenbruch and Sass (1978). Wright and Troxel (1973) called upon both mechanisms (plastic stretching *and* intrusion) to extend the deeper crust beneath the fault-fragmenting surface slab of the western Great Basin.

## NORMAL FAULTING AND BASAL SLIDING

Laboratory-scale models of extensional faulting that use sand or dry mortar as the deforming media (Hubbert, 1951; Stewart, 1971) yield structures similar to single, simple grabens and distributed arrays of alternating grabens and horsts. Stewart's model, which was designed specifically to resemble Great Basin structure, had a significant feature—a constructed surface of translatory sliding at its base. In order to predestine the spacing and plan of individual horsts and grabens, Stewart placed segmented sheets of paper beneath dry mortar. These sheets constituted a basal dislocation between the locally fragmenting mortar above and a sheet of uniformly extending rubber (the model's analog of a plastic substrate) beneath. Translatory sliding took place between the paper and the rubber sheet. In Hubbert's (1951) model, horizontal sliding took place at the base of the sand section, and it is not difficult to imagine striations developing on the floor of the deformation box parallel to the direction of extension after repeated runs.

In a real earth, such a dislocation could take one of two forms: (1) a simple surface of sliding or (2) a thin zone of

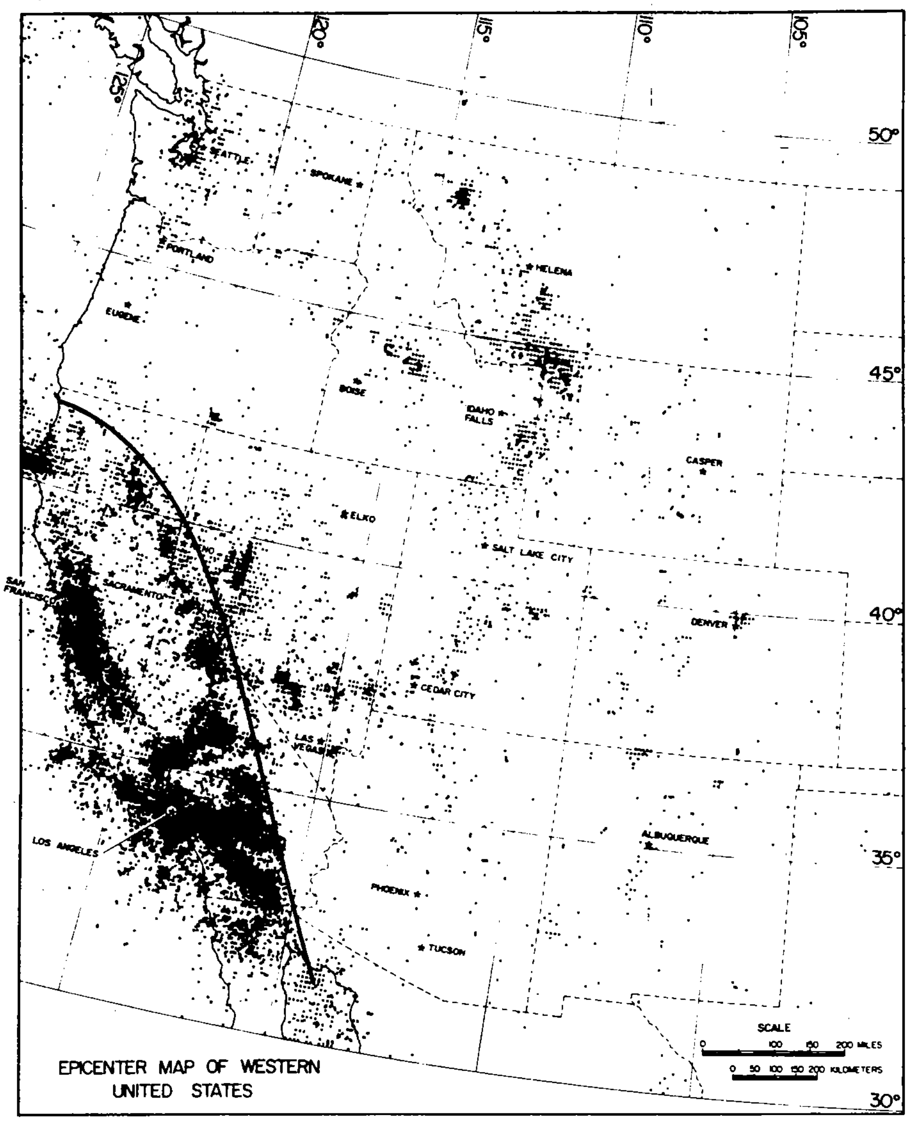


FIGURE 9.8 Seismicity of the western United States (from Smith, 1978). Lower threshold magnitudes were used in plotting California data. Heavy line, based on seismic data, marks inboard limit of highest earthquake event frequency and areal density in California, high cumulative seismic-strain energy release, and major strike-slip faulting related to dextral shear of the western plate boundary in Holocene time. Major faults within 150 km east of this line show oblique slip, with active strike-slip components, but farther east, the dominant mechanism is simple extension. (Reprinted from *Geological Society of America Memoir 152*, with permission.)

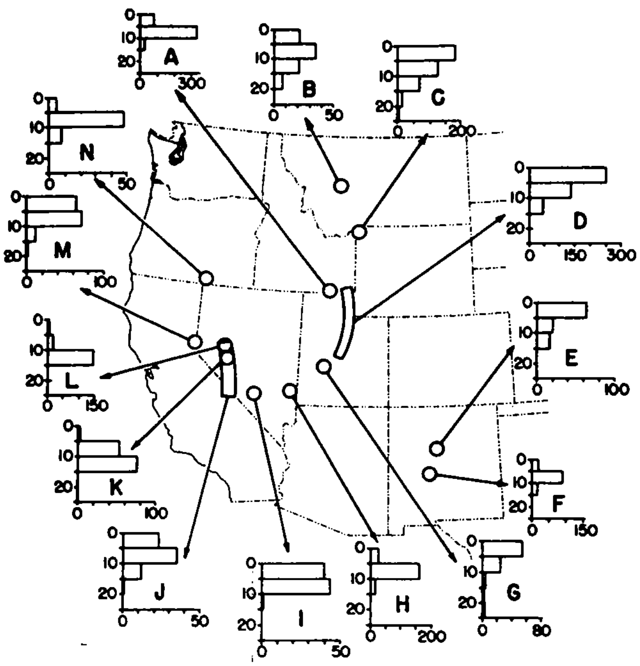


FIGURE 9.9 Histograms of earthquake focal depths (in kilometers) for 14 areas within the region of active extension in the western United States. Data from Gumper and Scholz (1971); Jaksha *et al.* (1977); Ryall and Savage (1969); Shuleski *et al.* (1977); and Smith (1978).

distributed shear or décollement. A subhorizontal striated surface, a thin zone of mylonite, or a thick layer of dynamothermally metamorphosed rock might thus be anticipated, depending on depth, effective pressure, temperature, composition, and shear stress.

The Turnagain Heights translatory slide in Alaska may be cited as an example of such deformation. It was described and illustrated by Hansen (1965) and mechanically analyzed by Voight (1973). It mimics the Basin and Range province in structure. The original thickness of the fragmenting slab was only 20 m, hence there was little tendency for normal faults to flatten appreciably with depth. Extension and sliding of the originally intact surface slab was possible because of an absence of lateral support on one side and a perceptible (2.2°) slope of the basal surface. Genetically, the structure was a gravity slide. It resulted from a loss in strength of material in the vicinity of the sliding surface as a result of excitation by a major earthquake. Owing to the backward retreat of its headwall, as more and more of the fragmenting surface slab slid away laterally, Voight termed the feature a "retrogressive block-glide." I do not suggest that the driving force of Basin–Range faulting is the same, only that kinematics and resultant structures are grossly similar and, for this reason, instructive.

Evidence suggests that the Great Basin may be growing laterally, i.e., that it may be consuming neighboring regions on both west and east (Smith *et al.*, 1976; Eaton *et al.*, 1978). Both margins show transitional regions that have geophysical anomalies characteristic of the Great Basin extending tens of kilometers into the neighboring provinces (see Ryall and Stuart, 1963; Shuey *et al.*, 1973; Smith and Sbar, 1974; Keller *et al.*, 1975; and Eaton *et al.*, 1978). The eastern margin of the Basin and Range province also has geological characteristics suggestive of transition (Best and Hamblin, 1978; Howard *et al.*, 1978; Luedke and Smith, 1978). This state may relate, at least superficially, to the retrogressive aspect of the Turnagain Heights slide. If so, the Sierran and Wasatch fronts (opposed headwalls) may be retreating from each other as a result of plastic stretching at depth, thus effecting a growth of the province at the expense of adjoining regions. This could account for the observed outward restriction of magmatism with time. Uplift in the regions of these headwalls is probably a thermal phenomenon that is essentially contemporaneous with the extensional faulting itself, just as it is in the oceans. Although heat flow in the Sierra Nevada is anomalously low, it must, in part, reflect the appreciable thermal time constant of the crust, for the mass deficiency characteristic of the Great Basin continues beneath the Sierra Nevada (Eaton *et al.*, 1978).

The Great Basin has a well-developed bilateral symmetry in certain aspects of its geology, but far more obviously in its geophysical fields (Proffett, 1977; Eaton *et al.*, 1978). In Proffett's model the western half of the shallow, fragmenting, crustal slab translates eastward relative to the extending substrate beneath (the middle and lower crust), and that of the eastern half, westward.

According to Voight (1973) retrogression cannot take place if the glide blocks are rigid (as opposed to internally deformable) unless fluid pressures within fractures are sufficiently high to perform the function of plastic wedges. Basin and Range blocks are sufficiently fractured at the surface to suggest internal deformation. The significant deformational model thus appears to be lateral spreading with deformable block gliding.

## GEOPHYSICALLY ANOMALOUS LAYERS IN THE SHALLOW CRUST

The possibility of a surface of sliding or a zone of ductile flow (mylonite or other metamorphic rocks) beneath the fault-fragmented surface slab of the Basin and Range province raises the question of their detectability by geophysical means. We examine this issue only briefly, but in the last section of the chapter offer a tentative crustal model, based primarily on geology, heat flow, and earthquake data, to which the other geophysical observations are fit by hypothesis. The hypothesis needs specific testing.

In the past decade and a half an increasing number of reports of a seismic low-velocity layer in the shallow crust have been published. One example is in the eastern Great Basin, near its boundary with the Colorado Plateaus (Mueller and Landisman, 1971; Landisman *et al.*, 1971; Braile *et al.*, 1974; Keller *et al.*, 1975; Smith *et al.*, 1975;

and Braile, 1977). Because such a feature is also associated with the Rhine graben (Landisman *et al.*, 1971) it is tempting to conclude that it is a feature characteristic of extensional regimes.

Shurbet and Cebull (1971) suggested that the crustal low-velocity layer in the Great Basin is a zone of decreased rigidity that provides a means of absorbing the displacements of Basin–Range faults. According to them, the top of this zone is at levels of 5 km or so, and the base, at 8–9 km. Braile *et al.* (1974) suggested that surface extension by normal faulting at the surface is absorbed in a soft, plastically extending region of lowered seismic velocity.

Braile (1977) later published the following additional information: (1) the layer has a compressional wave velocity perhaps as low as 5.5 km sec$^{-1}$ (compared with velocities above and below of 6.0 and 6.5 km sec$^{-1}$, respectively); and (2) it has a heightened Poisson's ratio; (3) an anomalously low $Q$ (quality factor) for the transmission of compressional seismic waves; and (4) a top at 9.5 km and base at 15 km. Although the depth and thickness of this layer are different in the Shurbet and Cebull model, the values were derived in very different ways. The figures of Braile (1977) are preferred. Both sets of authors agree on the existence and gross mechanical properties of the layer, and on its role in accommodating extension at the surface. Neat as this picture is, it is marred by the fact that crustal low-velocity layers are not peculiar to extensional regimes.

Two collections of papers on the physical properties and conditions of the continental crust (Heacock, 1971, 1977) reveal that the phenomenon is widespread, found in areas of young extension as well as in stable Precambrian shield areas (Berry and Mair, 1977). Such features have been observed in the crust of all the continents except Antarctica. Mueller (1977) has incorporated it as a key element in a generalized model of the continental crust. According to him, it is found fairly consistently at depths of 5–15 km. Some investigators, however, place it as deep as 20 km, e.g., Landisman and Chaipayungpun (1977). Its origin has been ascribed to high temperatures (Smith *et al.*, 1975), to the presence of a zone of granitic intrusions (Mueller, 1977), to high pore-fluid pressures (Berry and Mair, 1977), or to some combination of any or all of these factors (Mueller, 1977).

Laboratory experiments (Nur and Simmons, 1969; Todd and Simmons, 1972; and Brace, 1972b) demonstrate that as pore pressures rise toward lithostatic values (thereby reducing effective pressure) seismic velocities fall toward those observed at exceedingly shallow levels in the crust. If a consensus as to the origin of the crustal seismic low-velocity layer is emerging, it is high pore pressure. In the Great Basin, high regional heat flow, which implies high crustal temperatures, probably plays an important supportive role.

Pore water at high pressures in a closed system is capable of lowering seismic velocity, $Q$ values, and rock strength, conditions that appear to occur in the Great Basin crust. As Berry and Mair (1977) point out, however,

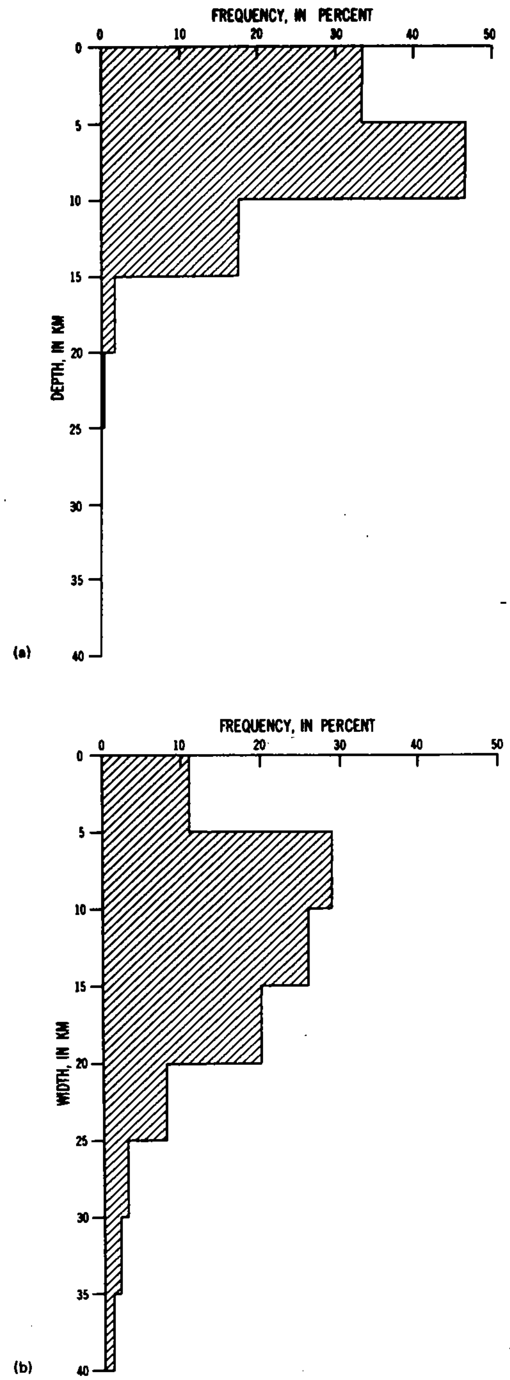


FIGURE 9.10 Earthquake focal depths and widths of Basin and Range blocks. (a) Histogram of focal depths of 2,475 earthquakes in the region of extension; (b) widths of individual basins and ranges in the Great Basin scaled from the map of King and Beikman (1974). Both characteristic dimensions (depth and width) show modal values in the range 5–10 km, with most values (>85 percent) in the range 0–20 km.

this explanation requires that rocks in the layer in question have finite porosities at depths of 5–15 km, while those in the zone immediately above it must be free of permeability because the hydraulically pressured layer must have some sort of impermeable cap. This cap in the Great Basin may be a layer of rock extended by ductile flow, the one whose upper surface may mark the base of the region of brittle faulting. If so, the seismic low-velocity zone may not coincide with this layer but is perhaps *beneath* it. Depths to surfaces or zones of Tertiary sliding in the Great Basin have been estimated by Armstrong (1972) to be at least 8 km on geological grounds, but the thick, ductile zones must be deeper still, because in many places the glide faults do not rest directly on ductile rocks. Because depths of somewhat more than 8 km are in reasonable agreement with Braile's (1977) seismic estimate of 9.5 km to the *top* of the seismic low-velocity layer, I tentatively regard the ductile layer (if it truly exists) as a possible cap. Pore fluids could be trapped at high pressure in fractures in rock immediately beneath such a ductile layer. Internal displacements or structural adjustments within rocks at this level would take place by stable sliding (Brace, 1972a; 1972b), and earthquakes would not be common at such depths. As we have already seen, they are not.

Pore water in closed-rock systems will also lower electrical resistivity, as will increasing temperature, which elevates the ionic mobility of such fluids. At very high temperatures the onset of partial melting could do much the same thing; the melting temperature of the rocks is lowered by the presence of water. For these reasons one might anticipate the presence of electrical conductors in the Great Basin crust, and, in fact, they are observed.

Some investigators (e.g., Landisman and Chaipayungpun, 1977; Lienert and Bennett, 1977) have equated the crustal low-velocity layer with a low resistivity layer in tectonically active or high heat-flow areas. Much remains to be done to substantiate this equivalence, and also to establish equivalence between a subhorizontal low-velocity layer in the crust and a porous zone capped by a stratum of ductile impermeable rock. The data in hand are permissive, but thus far hardly conclusive. The electrical data are reviewed briefly to provide an idea of what is currently known. I am indebted to my friend and colleague, J. N. Towle, for providing the information summary that follows.

Schmucker's (1970) geomagnetic variometer investigations in California indicated the presence of an electrical conductor in the western Cordillera at the eastern base of the Sierra Nevada, at and near its boundary with the Great Basin. Stanley *et al.* (1976b) identified a shallow (2–7 km), highly conductive layer in the crust beneath the Carson sink in western Nevada by means of magnetotelluric soundings. Stanley *et al.* (1976a) also studied the electrical structure of the Long Valley geothermal system in the western Great Basin by means of direct current and electromagnetic techniques, concluding that hydrothermal activity is reflected in discrete conductive zones in the crust, which are controlled, in turn, by regional faulting. Lienert and Bennett (1977) have identified a crustal conductor in the western Great Basin at a depth of 20 km using controlled-source geomagnetic variometry.

Reitzel *et al.* (1970) and Porath and Gough (1971) observed generally reduced vertical geomagnetic field variations in the eastern Great Basin that they interpreted as reflecting a shoaling of the mantle. Ambiguities in their interpretation of crustal thickness will doubtless be resolved as the evidence mounts both for a shallow, strongly conducting, crustal layer in the Great Basin as a whole and for local shoaling of the asthenosphere.

Studies by W. D. Stanley and colleagues at the U.S. Geological Survey (personal communication, 1978) have revealed the presence of a conductive crustal layer near the boundary between the Great Basin and Snake River Plain on the north. Depths to its top range from 2 to 10 km; its thickness may be as great as 10 km. Stanley *et al.* (1977) have also identified a conductor beneath the Snake River Plain region at depths of only 5 km in the Yellowstone caldera, but deepening to 20 km on the southwest. In the vicinity of the Raft River geothermal area, in the northeastern Great Basin, it is 7 km deep. This conductor may be related directly to the presence of magma, at least in the Yellowstone area, and, therefore, may or may not be directly related to the crustal low-velocity layer under discussion.

On the basis of these limited data it appears that an electrically conductive layer is a common feature of the shallow Great Basin crust. Depth estimates place its top between 2 and 20 km, and in several areas, at less than 10 km. Possibly this conductor coincides with the crustal low-velocity layer, but too little is known about it to be certain. Coincidence might be anticipated simply because some of the factors that lower seismic velocity also raise electrical conductivity (high temperature, high porosity, the presence of a pore fluid, or the presence of a silicate melt). An electrically conductive layer by itself does not require abnormally high pore pressures and, hence, does not require the presence of an impermeable cap to keep the system closed. The presence of conductive minerals such as metallic sulfides or those having high ion-exchange capacity, like clays or zeolites, can also lower rock resistivities without affecting seismic velocity. The low-resistivity layer in the crust could just as well be *above* the low-velocity layer, reflecting some combination of high porosity, temperature, pore-fluid salinity, or hydrothermal alteration in the lower part of the shallow crust.

## CRUSTAL MODEL FOR THE BASIN AND RANGE PROVINCE: A SUMMARY AND INTERPRETATION

The crust of the Great Basin section of the Basin and Range province (and its immediate environs) is higher in elevation, thinner, warmer, more highly fractured, and more well endowed with hot springs than that of surrounding regions, excluding the area immediately north of the Snake River Plain. The fractures (mostly faults) extend a third of the way to halfway through the crust. They are loci of abundant shallow earthquakes and vigorous hydrothermal circulation. Crustal extension takes place by faulting near the surface, but probably takes place by other modes at depth, most likely by dike intrusion and stretching of the lower crust and lithospheric mantle, and by ductile shear flow (distributed décollement) in a relatively thin layer at some intermediate level.

Repeated magmatic invasions of the crust have taken place during the past 100 million years. Some of these magmas broke the surface, but some have come to rest within the crust, giving up their heat there. Shallow magmatic systems serve to drive hydrothermal convection in the shallow crust, as does high regional heat flow from the deeper crust. The phenomena have been long lived.

At present, extension is taking place in an east–west or west–northwest direction; earlier, it was directed southwest or west–southwest (Eaton *et al.*, 1978; Zoback and Thompson, 1978). The Sonoran Desert was included in the initial episode of extension but is not included in the present one. Because the Sonoran Desert is deeply eroded, it exposes the effects of crustal extension at deeper levels. A large part of the Sonoran Desert in southwestern Arizona reveals evidence of subhorizontal, unidirectional plastic strain of middle Tertiary age (Davis, 1977; Davis *et al.*, 1977; Davis, see Chapter 8; Rehrig and Reynolds, 1977) that initially developed before block faulting began but that may have served as the base of the faulted, shallow slab. The mylonitization and metamorphism probably took place at lithostatic pressures of several kilobars.

Armstrong (1972) reviewed evidence of translatory displacements in the Basin and Range province and argued that some of the subhorizontal surfaces of sliding in the Great Basin are certainly Tertiary in age, that many of them *may* be Tertiary, and that they are more likely related to Basin and Range faulting than to the Sevier orogenic event, the youngest episode of pre-extension thrusting. Most of these dislocations place younger strata over older. He noted that some of the structures are of relatively deep-seated origin (at least 8 km). It is possible that these subhorizontal zones of sliding first developed as thrust soles during crustal compression, later to evolve into extensional décollements. These observations and speculations lend themselves to the interpretation that prolonged thermal conditioning of the crust plus horizontal shearing simply may have continued earlier initiated dynamothermal metamorphism of rocks that now serve as a boundary layer between parts of the lithosphere extending by fundamentally different mechanisms. As young normal faults developed near the surface in the regime of crustal extension, they became listric to (they came to sole on) older thrust zones.

The mechanical model of Kehle (1970), in which a décollement is distributed through the middle (relatively more ductile) layer of a crustal or lithospheric triad, may be applicable. Shearing in such a layer would lead to the mechanical generation of heat. Its magnitude would be controlled by the rate of shearing, which, judging from rates of extension measured at the surface, should be lower than that generated along the San Andreas Fault (see Lachenbruch and Sass, 1973). Some part of the high heat loss in the Great Basin could be due to shearing, however. The greater part has been ascribed to penetrative convection of the lithosphere by basaltic magma (Lachenbruch and Sass, 1978). Part is ascribable to convective groundwater circulation in the shallow crust and locally, to young, hot, volcanic systems residing in the upper crust. If this model, based largely on geological, heat flow, and earthquake data, is generally correct, it has implications for the surface patterns of deformation, the regional distribution of geological resources, and some of the effects of earthquakes. The model is shown in schematic form in Figure 9.11. To summarize its implications:

1. The location and extent of the Basin and Range province may have been largely predetermined by the location and extent of early Tertiary and Mesozoic magmatism that preheated (thermally weakened) the crust and augmented a regime of compressional thrusting in which subhorizontal dislocation surfaces or ductile zones (distributed décollements) first developed at middle to shallow crustal levels. Such zones could be used later as basal dislocations for normal faulting and might also serve (at depth) as crustal membranes impermeable to the deeper circulation of groundwaters but allowing the upward passage of magma by intrusion. Normal faults at the surface probably are listric to the deepest of these zones, inasmuch as the shallowest ones are exposed in the uplifted fault blocks themselves.
2. Zones of translatory, ductile shear would constitute near-horizontal surfaces of mechanical decoupling or inefficient coupling within the crust. As a result, deformations and kinematic motions in the lower crust would not always be clearly or faithfully reproduced at the surface.
3. The regional maintenance of long-continued high temperatures and high permeability assures the continuation of vigorous hydrothermal circulation and attendant epigenetic deposition of minerals through both compressional and extensional regimes. They may, on the other hand, be responsible for what appears to be a regional scarcity of oil and gas in the Basin and Range province. Where unfavorably situated, such fluids could be driven to the surface, where they could escape, except for local conditions of entrapment. For the same reason, consideration of Great Basin sites for the isolation and storage of radioactive wastes carries with it the requirement of a critical evaluation of the local hydrologic and seismotectonic regime.
4. Seismic energy traversing the crust of the extended region is probably absorbed to a somewhat greater degree than it is in the relatively less intensely fractured, shallow crust of the central and eastern United States, hence the

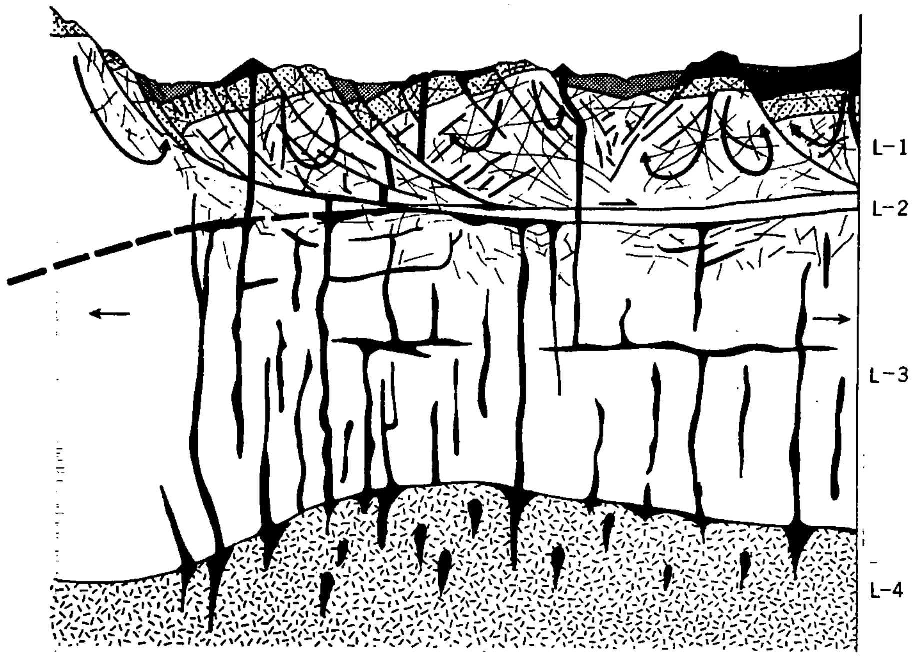


FIGURE 9.11 Interpretive model of possible crustal structure of the Great Basin (simplified, schematic, and not to scale); based on surface geology, heat flow, and earthquake distribution. (See Stewart, 1978, for alternative interpretations of the near-surface structure.) The crust is composed of three layers (L-1, L-2, L-3) having different lithologies and physical properties. Each fails or yields in extension by a different physical mode. Erosion in the Sonoran Desert region generally has cut down to the level of L-2. L-4 is lithospheric mantle. Characteristics of these layers are as follows: L-1, Fault-fragmented, surface layer, 8–15 km thick, composed of rocks of a great variety or origins, compositions, and ages, all exposed at the surface somewhere in the region; diagram shows Cenozoic continental sedimentary and volcanic rocks at the surface (patterns of dense stippling and solid black, respectively) overlying older sedimentary rocks (open stippling) and granitic and metamorphic basement rocks (plain white). Although the diagram does not show it, stratified rocks extend well into L-2 and probably into L-3 (as in Arizona). All of these rocks are highly fractured, as indicated by the plexus of fine, irregular lines. The layer fails in semibrittle fashion by normal faulting, fault-block rotation, pervasive fracturing, and slumping. The deformation creates high fracture porosity and permeability, allowing convective circulation of groundwater (curved arrows) driven both by the high heat flow from the deeper crust and by local, young intrusions (black, dike-like bodies). The upper part of the crustal low-resistivity layer may coincide with the lower part of this layer. The base of the layer generally marks the maximum depth of earthquakes. L-2 ductile intermediate layer, 0–3 km thick, composed of pervasively sheared, mylonitized, and/or dynamothermally metamorphosed Miocene and older rocks of a wide variety of original compositions (medium stippling). At one extreme, the layer is a vanishingly thin stratum of mylonite, 1–10 mm thick; at the other, a layer of granitic augen gneiss, schist, or amphibolite, 1–3 km thick. This layer is locally or regionally lineated and extends by laminar plastic flow. It is generally impermeable to groundwater circulation except where later uplifted and fractured in the brittle regime, but at depth it may be cut by dikes of Tertiary igneous rocks (solid block). It developed first as a regional thrust sole (heavy dashed line) during earlier crustal compression. L-3, lower crustal layer, 10–20 km thick, composed near its top of igneous and metamorphic basement rocks like those of layer L-1 but grading downward into increasing proportions of old granites, migmatites, gneisses, amphibolites, and felsic to mafic granulites, in approximately that order. This layer extends by a combination of diking (by basalts from the asthenosphere, solid black) and solid-state convection (stretching and underplating). It is rigid at relatively high and intermediate strain rates. These modes of penetrative convection are responsible for the mass transport of heat from the deeper mantle, causing the anomalously high heat flow observed in the province. The seismic low-velocity zone may coincide with the uppermost part of this layer as a result of anomalously high pore pressure in a system capped by the impermeable layer, L-2, and high temperature. L-4, lithospheric mantle, 25–35 km thick, composed of ultramafic rock devoid of finely disseminated melt. This layer, like the lower crustal layer above it, extends by diking (perhaps via rising, bleb-like bodies) and solid-state convection. It is immediately underlain by asthenospheric mantle.

geographical extent of isoseismal boundaries for an earthquake of given magnitude is generally less in the West than it is in the Midwest and East.

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