13

Late Cenozoic volcanic and tectonic evolution
of the Great Basin and
Columbia Intermontane regions

ROBERT L. CHRISTIANSEN
EDWIN H. MCKEE
U.S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

ABSTRACT

The Great Basin is a tectonically youthful region that shares some features with the Columbia Intermontane region but is separated from the more mature southern part of the Basin and Range province by a zone of active seismicity and geophysical contrasts. Sedimentary, physiographic, and structural features show that during the past 17 m.y., extensional linear normal faulting has been active in the Great Basin region, and extension also is indicated by numerous dikes in the High Lava Plains and the Columbia Plateau. Cumulative tectonic extension in the Great Basin is more than 100 km. Since about 14 m.y. ago, tectonic activity in the Great Basin region has tended to become progressively more concentrated toward the margins, and extension has been taken up by a wide transform zone along the High Lava Plains. Within several tens of kilometres north of the High Lava Plains of Oregon and Idaho, cumulative extension is generally less than a few kilometres and has been nearly inactive since about 14 m.y. ago.

Volcanism in the past 17 m.y. has been characterized by basaltic and bimodal rhyolite-basalt suites. Between 17 and 14 m.y. ago, the predominant volcanism was basaltic, being somewhat alkalic and of relatively small volume in the central Great Basin, more voluminous and less alkalic northward into the plateaus of southern Oregon and the High Lava Plains, and extremely voluminous and tholeiitic in the Columbia Plateau. Since about 14 m.y. ago, basaltic and bimodal volcanism has occurred throughout the Great Basin region but generally has tended to erupt in successively narrower zones near its margins, probably in direct correspondence to the increasing concentration of normal faulting toward these margins. The High Lava Plains have been characterized during this same time by two linearly propagating volcanic systems, in which major cycles of rhyolitic volcanism have been initiated successively farther northwest and northeast. These two volcanic systems have propagated away from a region in the center

of the High Lava Plains at about the same rate that faulting and volcanism in the Great Basin have been concentrated toward its margins.

A model that accounts for this evolution relates tectonic extension to the regional stress fields that result from the motions and changes in the interactions of the North American, Pacific, and Farallon lithospheric plates. In this model, geophysical and volcanic features of the region are interpreted to be due to a chain of heating events caused by this extension but conditioned by the stress and thermal history of the continental plate. Stress relief at the base of the lithosphere causes basaltic magma generation of varying amounts and at varying depths in the upper mantle, depending on the thickness and history of the overlying crust. The generation of basaltic magmas and their intrusion into and through the crust during continued extension have increased regional heat flow, lowered the rigidity of the lithosphere, caused crustal thinning, produced flowage and decreased seismic velocities in the upper mantle, caused regional uplift by thermal expansion, and produced rhyolitic magmas by localized partial melting of the lower crust.

According to the model, initial rifting occurred between 17 and 14 m.y. ago when northward migration of the Mendocino triple junction caused the continental-margin subduction zone to become short enough to allow partial coupling between two zones of transform displacement of the Pacific and North American plates. The increased coupling between these two zones caused extension in the North American plate perpendicular to the continental margin. Since about 14 m.y. ago, continued tectonic extension and basaltic magma generation have (1) caused a wide zone of oblique extension to become successively hotter and less rigid near the zone's central axis, (2) increasingly concentrated brittle deformation and high-level magmatism outward toward the margins of the Great Basin region, and (3) produced concentrated zones of extension and crustal melting at the intersections of the resulting marginal zones with the transitional northern transform boundary of the extending region. This accounts for the symmetrically propagating volcanic systems of the High Lava Plains. The Yellowstone melting anomaly, whose locus was controlled initially by an old structural boundary, was favorably oriented to be augmented by shear melting at the base of the lithosphere; it has become self-sustaining because of the initiation of a thermal feedback cycle and the development of a root in the mantle by inward flow around a dense, sinking, unmelted residuum.

INTRODUCTION AND REGIONAL SETTING

The Cenozoic basin-range region of the Western United States is a region of extensional block faulting that includes the Basin and Range physiographic province and some other bordering areas. A number of models have been proposed recently for the origin of the basin-range region (for example, Menard, 1964; Hamilton and Myers, 1966; Cook, 1969; Atwater, 1970; Scholtz and others, 1971; Christiansen and Lipman, 1972; Noble, 1972; Thompson and Burke, 1974; Suppe and others, 1975). Many of these models, based in part upon time-space patterns of Cenozoic volcanism and regional structures or upon global tectonic models, have focused on the Great Basin of Nevada, western Utah, and adjacent areas. To the extent that the view has widened, it commonly has been cast southward toward the rest of the Basin and Range province from southern California to New Mexico.

In this paper we attempt to build a conceptual model of late Cenozoic volcanic and tectonic evolution by focusing on relationships between the northern part of the basin-range region, here referred to as the Great Basin region, and the largely volcanic plains and plateaus that lie farther north (Fig. 13-1), sometimes referred to as the Columbia Intermontane province

(for example, see Thornbury, 1965). Volcanic and tectonic features that are common to these two regions or that mark a unique transition between them may (as suggested by Hamilton and Myers, 1966) denote a genetic linkage that is as fundamental as the relationship between the Great Basin region and the southern part of the Basin and Range province.

The most important pattern common to the Great Basin and Columbia Intermontane regions is a two-stage Cenozoic volcanic and tectonic development that seems to have been nearly synchronous. From about 50 to 17 m.y. ago this entire area was a postorogenic terrane characterized in places by predominantly calc-alkalic volcanism of intermediate to silicic composition (McKee and others, 1970; Lipman and others, 1972; Stewart and Carlson, 1976 and this volume). During that time, subduction was active along the Pacific margin west of the area although after about 29 m.y. ago, subduction ceased to the south (Atwater, 1970; Atwater and Molnar, 1973). Volcanism in the area before 17 m.y. ago probably was related to that subduction (Lipman and others, 1972). A brief lull in volcanism about 17 m.y. ago (McKee and others, 1970) was followed by progressive decline of the subduction zone to the west and by regional tectonic extension and predominantly basaltic or bimodal rhyolite-basalt volcanism (Christiansen and Lipman, 1972; Stewart and Carlson, 1976 and this volume).

During the time since about 14 to 17 m.y. ago, the typical basin-range pattern of subparallel fault-bounded linear ranges separated by alluviated basins has typified most of the Great Basin region, while the Columbia Intermontane region has evolved mainly as a series of volcanic
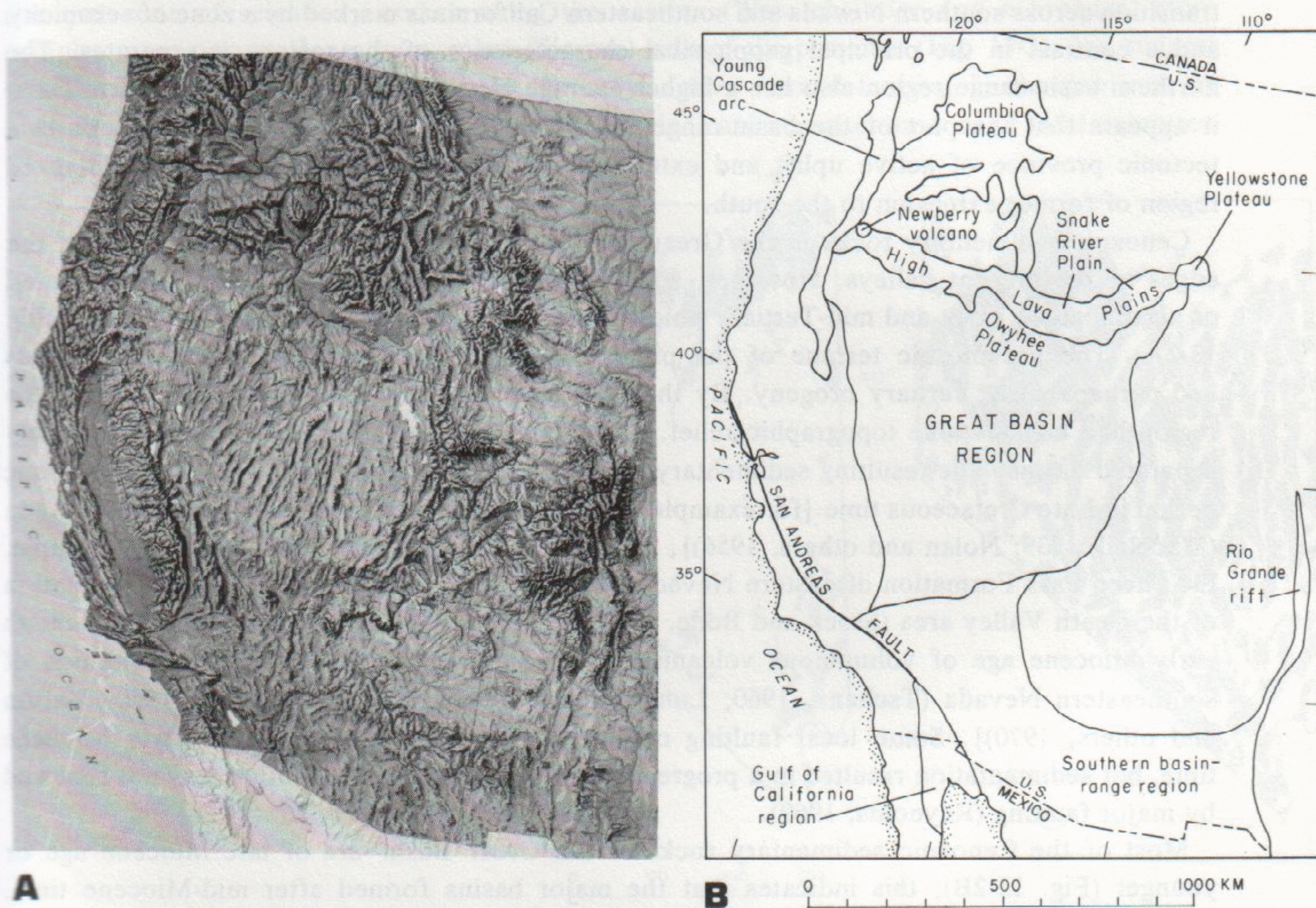

Figure 13-1. Physiography of the Western United States. (A) Shaded relief map; note well-defined physiographic break between the northern and southern parts of the Basin and Range province (after U.S. Geol. Survey, *National Atlas of the United States of America*, p. 56). (B) Index to some late Cenozoic tectonic and volcanic features shown on Figure 13-1A.

plains. Basin-range topography is well developed south of the High Lava Plains of central and eastern Oregon and southern Idaho (Fig. 13-1), but within several tens of kilometres north of those plains, this topographic pattern generally declines and is no longer predominant. Conversely, the volume of predominantly basaltic lavas of late Cenozoic age is greatest in the Columbia Plateau, declines southward across Oregon and Idaho into the northern Great Basin, and is relatively small in the central and southern Great Basin.

PHYSIOGRAPHY, BASIN DEPOSITS, AND STRUCTURE

Great Basin Region

The parallel linear basins and ranges of the Great Basin region are youthful fault-bounded features. The range fronts are generally steep, and fault scarps break many of the bordering alluvial fans (Slemmons, 1967). On a regional scale these characteristic features of linearity, parallelism, and conspicuous youthfulness contrast with the Basin and Range province south of a sharply defined east-trending zone through southern Nevada and California (Fig. 13-1). In the southern basin-range region the mountains commonly are lower, have less well defined fronts, and have wide surrounding complexes of pediments and alluvial fans; fresh fault scarps and active seismicity are uncommon. Eaton and others (this volume) note that this abrupt transition across southern Nevada and southeastern California is marked by a zone of seismicity and a contrast in the principal geophysical characteristics of the regions it separates. The northern basin-range region also has a higher average elevation than the southern part. Thus, it appears that the part of the basin-range region north of the southern Nevada zone is a tectonic province of active uplift and extension that is separated from a nearly stabilized region of former extension to the south.

Cenozoic sedimentary rocks in the Great Basin region most commonly occur in or on the edges of the present valleys. However, some of the older sedimentary rocks that predated or accompanied early and mid-Tertiary volcanism occur high in the block-faulted ranges (Fig. 13-2A). The prevolcanic terrane of the present Great Basin succeeded a Late Cretaceous and perhaps early Tertiary orogeny. By the time most orogenic movements had ceased, the region had considerable topographic relief, and fluvial and lacustrine sediments filled widely separated basins. The resulting sedimentary units range in age from that of the major orogenic period in Late Cretaceous time [for example, the Newark Canyon Formation of central Nevada (MacNeil, 1939; Nolan and others, 1956)], through Eocene and early Oligocene [for example, the Sheep Pass Formation of eastern Nevada (Winfrey, 1960) and the Titus Canyon Formation of the Death Valley area (Stock and Bode, 1935; Reynolds, 1969)], into the Oligocene through early Miocene age of voluminous volcanism [for example, the Horse Spring Formation of southeastern Nevada (Tschanz, 1960; Longwell and others, 1965; Armstrong, 1970; Marvin and others, 1970)]. Some local faulting continued between Cretaceous and early Miocene time, but sedimentation resulted in a progressive reduction of local relief that was not renewed by major faulting (Reynolds, 1969).

Most of the Cenozoic sedimentary rocks of the Great Basin are of late Miocene age or younger (Fig. 13-2B); this indicates that the major basins formed after mid-Miocene time. Progressive development of basin-range topography and concurrent basin filling are recorded by thick alluvial deposits (Fig. 13-2C) and numerous fault or fault-line scarps. Facies gradations are abrupt from range-front conglomerates to tuffaceous fine-grained strata; this indicates that they accumulated in basins much like those of today. The physiography and sedimentary

rocks, therefore, represent a distinct geologic regime that started abruptly in mid-Miocene time and has since remained active. Where the basin-filling sedimentary rocks have been dated radiometrically or where their fossils can be calibrated to a chronology, it is clear that the large separate basins were well defined by at least about 13 m.y. ago (Robinson and others, 1968; Stewart, this volume).

It is difficult to document the age of initial basin-range faulting in detail, especially since faulting has continued virtually to the present throughout much of the region. There is, however, widespread structural evidence that basin-range topography did not exist as a regional landform prior to about 17 m.y. ago (McKee and others, 1970; McKee, 1971; Stewart, 1971; Christiansen and Lipman, 1972; Noble, 1972; McKee and Noble, 1974). Numerous sheetlike welded ash-flow tuffs of early Miocene age (25 to 20 m.y. old) in central Nevada indicate that there was little topographic relief comparable to present-day basin-range topography in that part of the province, and the relative scarcity of sedimentary rocks older than about 17 to 18 m.y. is consistent with that evidence.

Estimates of the cumulative tectonic extension across the Great Basin during the past 17 m.y. vary widely. Hamilton and Myers (1966) estimated extension of 100 to 300 km. Thompson and Burke (1974) reviewed convincing evidence for a minimum of 50 to 100 km, roughly a 10% increase in width. Stewart (this volume) reviews published evidence for cumulative extension and shows that most estimates range from less than 10% to about 35%. Although there is abundant evidence for very young basin-range faulting throughout the entire province, a major zone of seismicity occurs along the eastern (Wasatch) edge and a broader, more complex zone occurs along the western (Sierran) side of the province (Fig. 13-3). We suggest that these zones represent the loci of the most active normal faulting in the province as a whole. Stewart (this volume) and Proffett (1977) indicate a greater amount of total extension
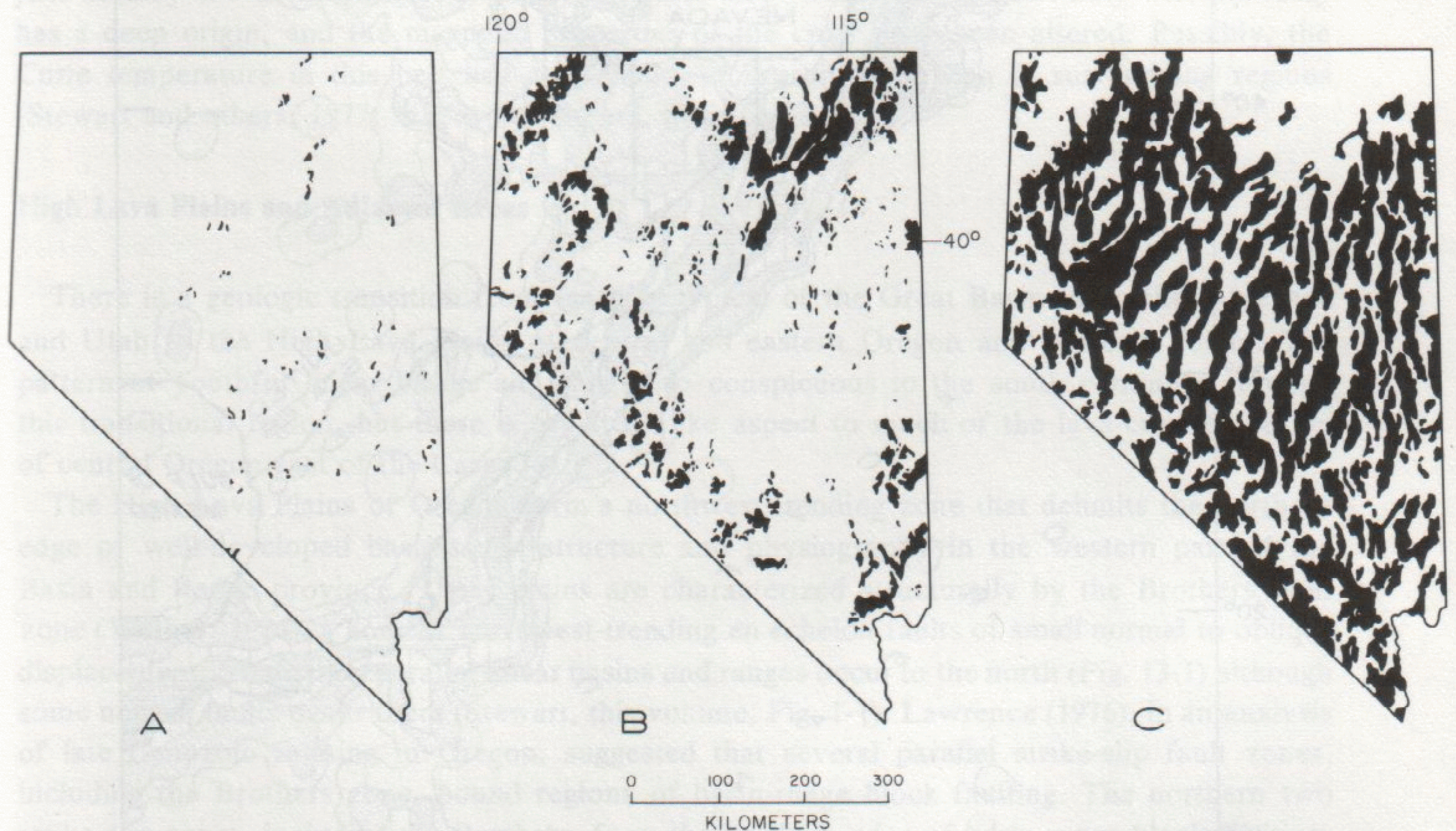

Figure 13-2. Maps showing surface distribution of Tertiary sedimentary rocks and Quaternary sediments in Nevada, taken as an example of the Great Basin region. (A) Sedimentary rocks 40 to 17 m.y. old (before basin-range faulting). (B) Tertiary sedimentary rocks less than 17 m.y. old. (C) Quaternary sediments. (In part after Stewart and Carlson, 1976.)

in the Lahontan and Bonneville basins (near the margins of the Great Basin) than in the center. Similarly, Wright (1976) demonstrated greater cumulative displacement on the western side of the Great Basin than across its central region.

Scott and others (1971) and Scholz and others (1971) suggested that basin-range faulting and extension started first near the central part of the Great Basin and spread progressively toward the present margins throughout later Cenozoic time. However, Noble (1972) gave evidence that extensional faulting began simultaneously over a wide front across the Great Basin. Noble and Slemmons (1975) noted a 12.5-m.y.-old dike emplaced into an older normal fault near the Sierran front and gave evidence that a third of the tilting of the Sierra Nevada occurred before 9.5 m.y. ago. We note further that at least some of the present basins and ranges near the edges of the province were well defined by at least 10 to 14 m.y. ago. For example, the Esmeralda Formation and related units began to accumulate before about 13 m.y. ago in basins near the western edge of the Great Basin (Robinson and others, 1968). Similarly,
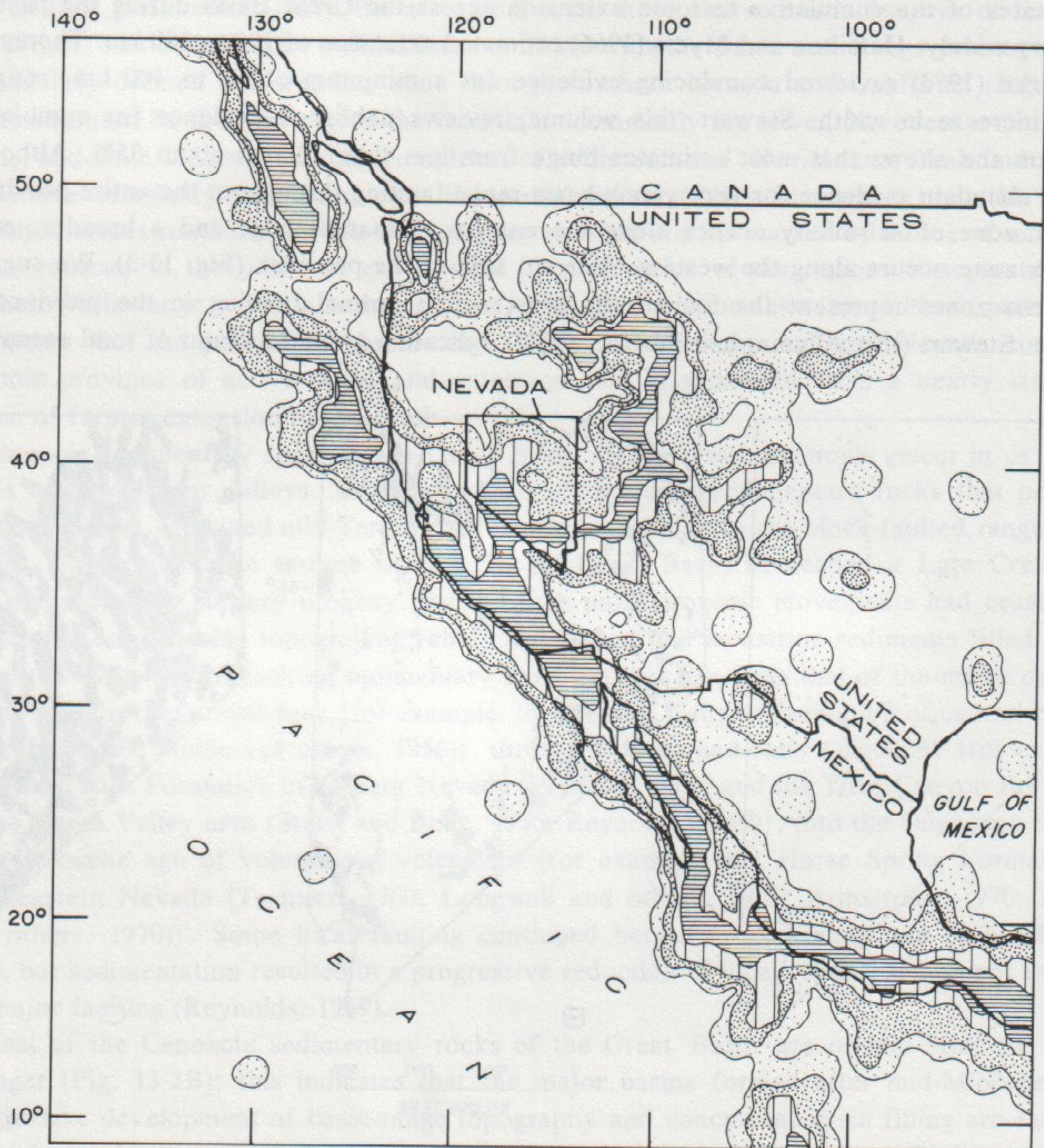

Figure 13-3. Map of seismicity of the Western United States. Based on epicenters for 1961 through 1967 (after Barazangi and Dorman, 1969); contours are 1, 2, 3, and 4 or more epicenters per circular area with radius 1° of latitude.

the Teton Range and the adjacent block-faulted basin of Jackson Hole at the northeasternmost edge of the tectonic province had formed by normal faulting oblique to Laramide structures before initial deposition of the Teewinot Formation about 10 m.y. ago (Love, 1956; Love and others, 1972).

Thus, several lines of evidence point to development of the entire province within a few million years of initial rifting. Although the present low seismicity cannot be regarded as evidence of complete tectonic inactivity in the central Great Basin (Slemmons, 1967), the present zones of greatest seismicity and cumulative extension are concentrated near the margins of the Great Basin. The pattern of tectonic extension appears to have evolved from an early stage of uniform distribution through stages of concentration into successively narrower zones toward the margins.

By analysis of first arrivals from earthquakes in the Great Basin region, Smith and Sbar (1974) and Smith (this volume) show that normal-fault mechanisms predominate and that the directions of minimum principal stress generally lie between northwest-southeast and east-west. Numerous regional geophysical studies have shown that the Basin and Range province is characterized by high heat flow (Sass and others, 1971; Lachenbruch and Sass, 1977), a thin crust and low upper-mantle seismic velocity (Hill, this volume; Smith, this volume), and a low regional gravity field (Eaton and others, this volume).

Near the symmetry axis of the Great Basin is a north-trending zone (the "quiet zone" of Stewart and others, 1977) defined by relatively low aeromagnetic intensities and a distinct lack of the high-frequency anomalies common elsewhere in the region (Fig. 13-5). This zone is especially well defined between lat 38°N and 39°N, where it appears as an aeromagnetic low between regions with a conspicuous high-frequency magnetic signature. Rocks at the surface in the low-anomaly belt are mainly middle and late Cenozoic rhyolitic ash-flow tuffs just as they are in aeromagnetic highs on either side. Thus, the low-anomaly belt probably has a deep origin, and the magnetic properties of the crust have been altered. Possibly, the Curie temperature in this belt lies at a shallower crustal level than in surrounding regions (Stewart and others, 1977; Mabey and others, this volume).

High Lava Plains and Adjacent Areas

There is a geologic transition from features typical of the Great Basin in northern Nevada and Utah to the High Lava Plains of central and eastern Oregon and southern Idaho. The pattern of youthful linear basins and ranges so conspicuous to the south continues through this transitional region, but there is a plateau-like aspect to much of the lava-covered region of central Oregon east of the Cascades.

The High Lava Plains of Oregon form a northwest-trending zone that delimits the northern edge of well-developed basin-range structure and physiography in the western part of the Basin and Range province. These plains are characterized structurally by the Brothers fault zone (Walker, 1974), a zone of northwest-trending en echelon faults of small normal to oblique displacement. No distinct parallel linear basins and ranges occur to the north (Fig. 13-1) although some normal faults occur there (Stewart, this volume, Fig. 1-1). Lawrence (1976), in an analysis of late Cenozoic faulting in Oregon, suggested that several parallel strike-slip fault zones, including the Brothers zone, bound regions of basin-range block faulting. The northern two strike-slip zones, including the Brothers, form the northern edge of basin-range block faulting.

The eastern Snake River Plain trends northeast across southwestern to east-central Idaho (Fig. 13-1). Along the southern margin of the Snake River Plain, prominent basin-range structure and topography are characteristic of the Great Basin region. The plain itself is a lava-covered

region of low relief that marks a partial northern boundary to the basin-range tectonic province. One group of parallel block-faulted northwest-trending ranges (the Lemhi, Lost River, and Beaverhead Ranges) lies north of the eastern Snake River Plain, but farther northward these ranges abut the Idaho batholith terrane (Fig. 13-1), and even farther north and northeast the predominant physiographic elements are rejuvenated Mesozoic structural features of the Northern Rocky Mountains (see Pardee, 1950; Robinson, 1960, 1961). Eaton and others (1975) and Mabey and others (this volume) interpret the eastern Snake River Plain to lie along a regional structural feature of the Precambrian basement, marked by a pattern of aeromagnetic anomalies that extends beyond the plain southwestward into northern Nevada and northeastward across the Rocky Mountains and Great Plains to the Canadian border.

The western Snake River Plain is a northwest-trending basalt-filled rift that dates from mid-Miocene time (Mabey, 1976).

Seismicity along the margins of the Great Basin region continues northward nearly to the High Lava Plains; in the east, seismicity continues to the Snake River Plain (Fig. 13-3). A conspicuous east-trending belt of seismicity (the Idaho seismic belt of Smith and Sbar, 1974) marks this northward transitional boundary of the active region north of the Snake River Plain, although relatively minor seismicity continues northward to about the Canadian border. The relatively high heat flow typical of the Great Basin also characterizes the Northern Rocky Mountains (Blackwell, 1969), but the area of highest regional heat flow, like the seismically most active region, extends only a short distance north of the Snake River Plain (Lachenbruch and Sass, 1977).

Columbia Plateau

The northern part of the Columbia Intermontane region is the Columbia Plateau, a vast basaltic province that formed during middle and late Miocene time but is now quiescent. The generally flat topography of the Columbia Plateau, incised by major throughgoing drainages, contrasts markedly with the structure and topography of the Great Basin region. Large young folds characterize parts of the Columbia Plateau, but its gross structure is dominated by the large Pasco basin at its center, formed by regional subsidence during eruption of the Columbia River Group, mainly about 17 to 14 m.y. ago (Swanson and others, 1975; Watkins and Baksi, 1974). The other principal structural element of the Columbia Plateau comprises numerous more-or-less parallel north-northwest–trending basaltic dikes, the feeders for the Columbia River basaltic flows (Taubeneck, 1970; Swanson and others, 1975). These dikes represent aggregate extension of more than 1 km (Taubeneck, 1970), most of it during a short span of the mid-Miocene at about the same time as initial extension and basaltic volcanism in the Great Basin.

Thus, despite the younger folding evident in the Columbia Plateau, an initial tectonic regime of roughly east-west minimum stress was comparable to the generally east-west to northwest-southeast extension of the Great Basin region. Present seismicity of the Columbia Plateau is weak.

Regional Tectonic Pattern

In summary, the regional structure, topography, drainage, and basin deposits of the Great Basin and Columbia Intermontane regions show the following situations: (1) The late Cenozoic structures form a pattern of uplift and roughly east-west to northwest-southeast extension that is superimposed across a heterogeneous early and middle Tertiary postorogenic volcanic

terrane. (2) The present pattern of structure and topography began to form about 17 m.y. ago and was essentially fully developed at least by 12 to 14 m.y. ago. (3) After an initial unified stage 17 to 14 m.y. ago, the region became divided into a zone of continuing extensional tectonism in the Great Basin region to the south and a zone of declining activity in the Columbia Plateau to the north; the High Lava Plains of Oregon and Idaho mark a transitional boundary between these two zones. (4) In contrast to the more mature-looking southern basin-range region, the entire Great Basin shows evidence of youthful tectonism. However, the most active seismicity occurs in the southern, western, and eastern parts of the Great Basin region, and the zones of greatest cumulative extension tend to be near its western and eastern margins. Thus, since 12 to 14 m.y. ago, the most active faulting appears to have been increasingly concentrated toward the margins of the Great Basin. (5) The transitional northern boundary zone along the High Lava Plains is generally aseismic (Fig. 13-3); the western branch roughly parallels the Brothers fault zone of probable strike-slip displacement, whereas the eastern branch is the subsiding eastern Snake River Plain.

VOLCANISM

Great Basin Region

Upper Cenozoic volcanic rocks, predominantly basaltic and subordinately rhyolitic, are common in the Great Basin, especially around its margins. The basalts and associated silicic rocks are most voluminous in the northern part of the province (Fig. 13-4; also see Stewart and Carlson, 1976 and this volume). Almost all the Cenozoic basalts of the Great Basin are less than about 17 m.y. old, and most are less than about 12 m.y. (McKee and others, 1970; McKee and Noble, 1974; Snyder and others, 1976).

The oldest of the upper Cenozoic basalts of Nevada are dikes and flows in north-central Nevada that have been dated by K-Ar at various places as about 17 to 14 m.y. old. These basalts form a north-northwest-trending outcrop belt that is broken by younger north-trending tectonic basins. The belt, however, is continuous beneath the basins, as shown by a prominent linear aeromagnetic high extending from central Nevada to the Idaho border (Fig. 13-5). This anomaly has been described in detail by Mabey (1966) and Robinson (1970) and is the southern part of the Oregon-Nevada lineament of Stewart and others (1975), who emphasized a direct relationship between the aeromagnetic anomaly and the mafic dikes. Although outcrops of basalt closely follow the aeromagnetic high (compare Figs. 13-4 and 13-5), analysis of the anomaly indicates that it is not caused by the surface basalts alone. The amplitude of the anomaly and its narrow linear form must also reflect a deeper zone of basaltic rocks (Mabey, 1966). This sharply defined, long (200 km), narrow (~10 km) anomaly of deep origin (10 to 15 km; Robinson, 1970) near the center of the Great Basin suggests a basaltic dike swarm close to the line of symmetry of the province although somewhat to the west. Exposed dikes in the swarm have the same trend as the belt as a whole. In the Roberts Mountains, near the south end of the belt, all the exposed basalts are dikes in a 7-km-wide swarm (Fig. 13-6). Farther northwest, most of the dikes are buried by more widespread basaltic flows, although continued presence of the dikes at depth can be inferred from the continuity of the aeromagnetic high (Fig. 13-5).

The region containing voluminous basalts and associated silicic volcanic rocks 17 to 14 m.y. old widens northward and becomes diffuse as it approaches the Nevada-Idaho-Oregon border area. The axial belt crosses the Owyhee Plateau region (Figs. 13-1, 13-4) as a zone
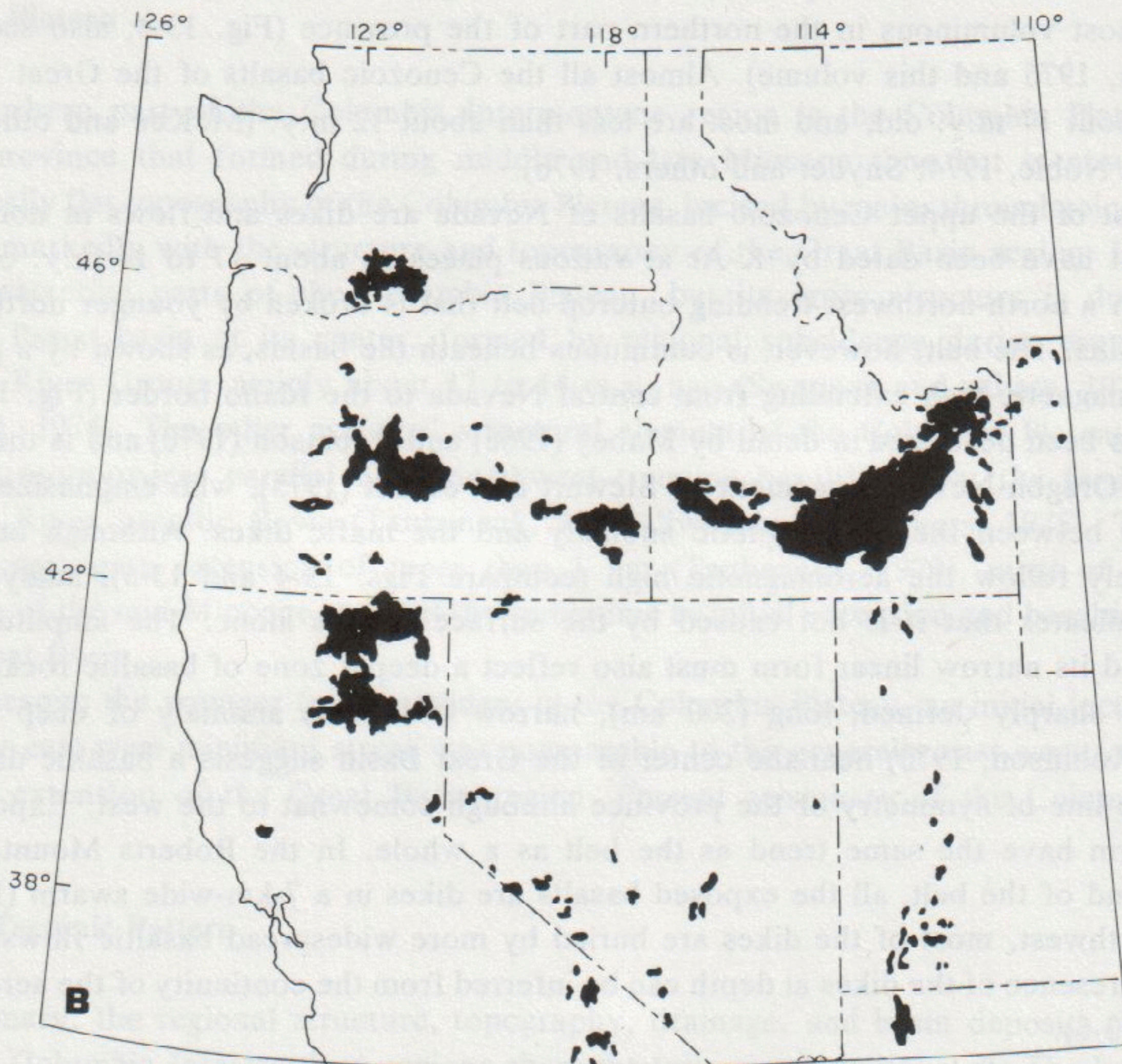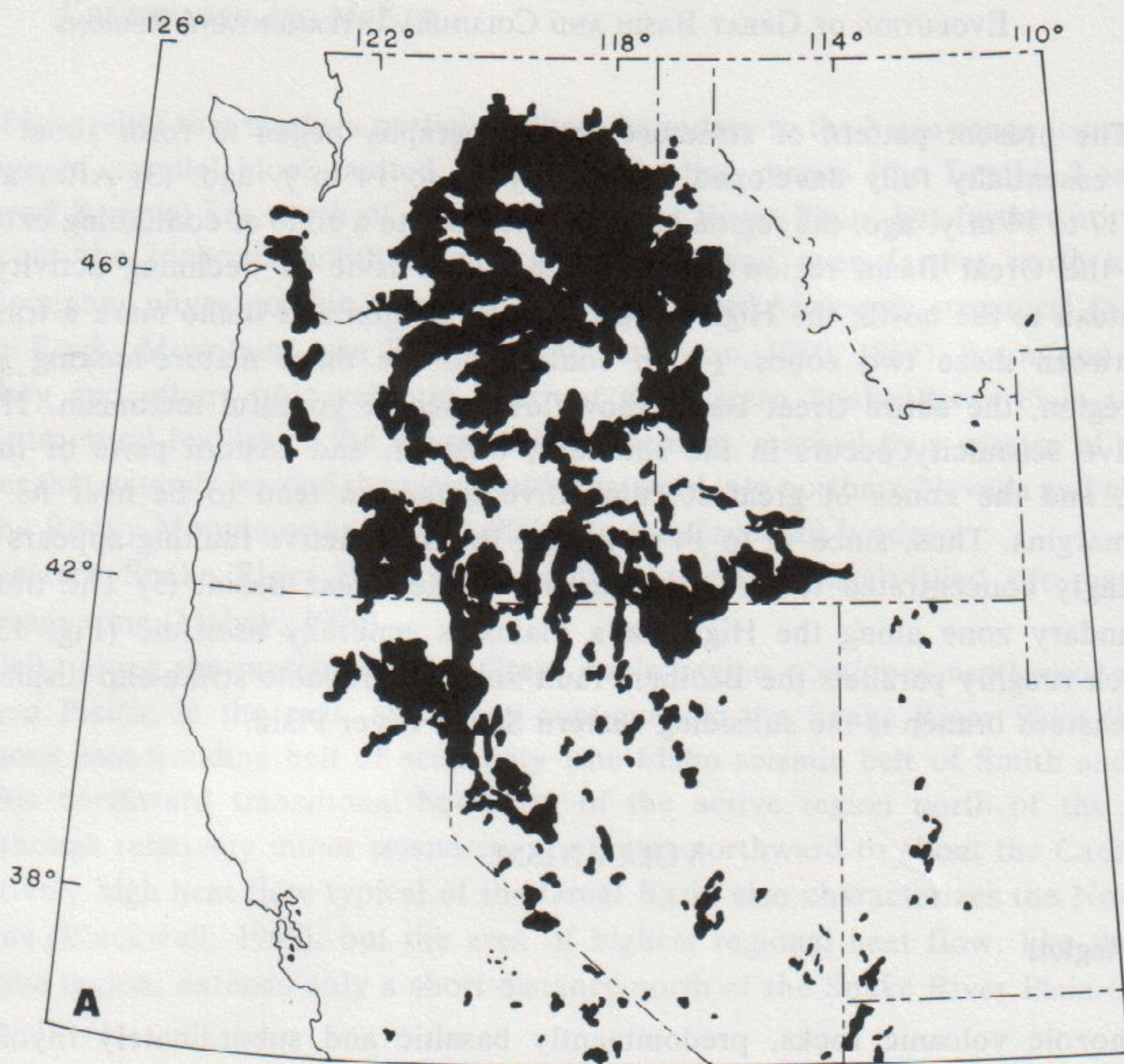
Figure 13-4. Maps showing distribution of Tertiary and Quaternary basalt in the Great Basin and Columbia Intermontane regions. (A) Tertiary basalts less than 17 m.y. old. (B) Quaternary basalts. (In part after Stewart and Carlson, 1976.)
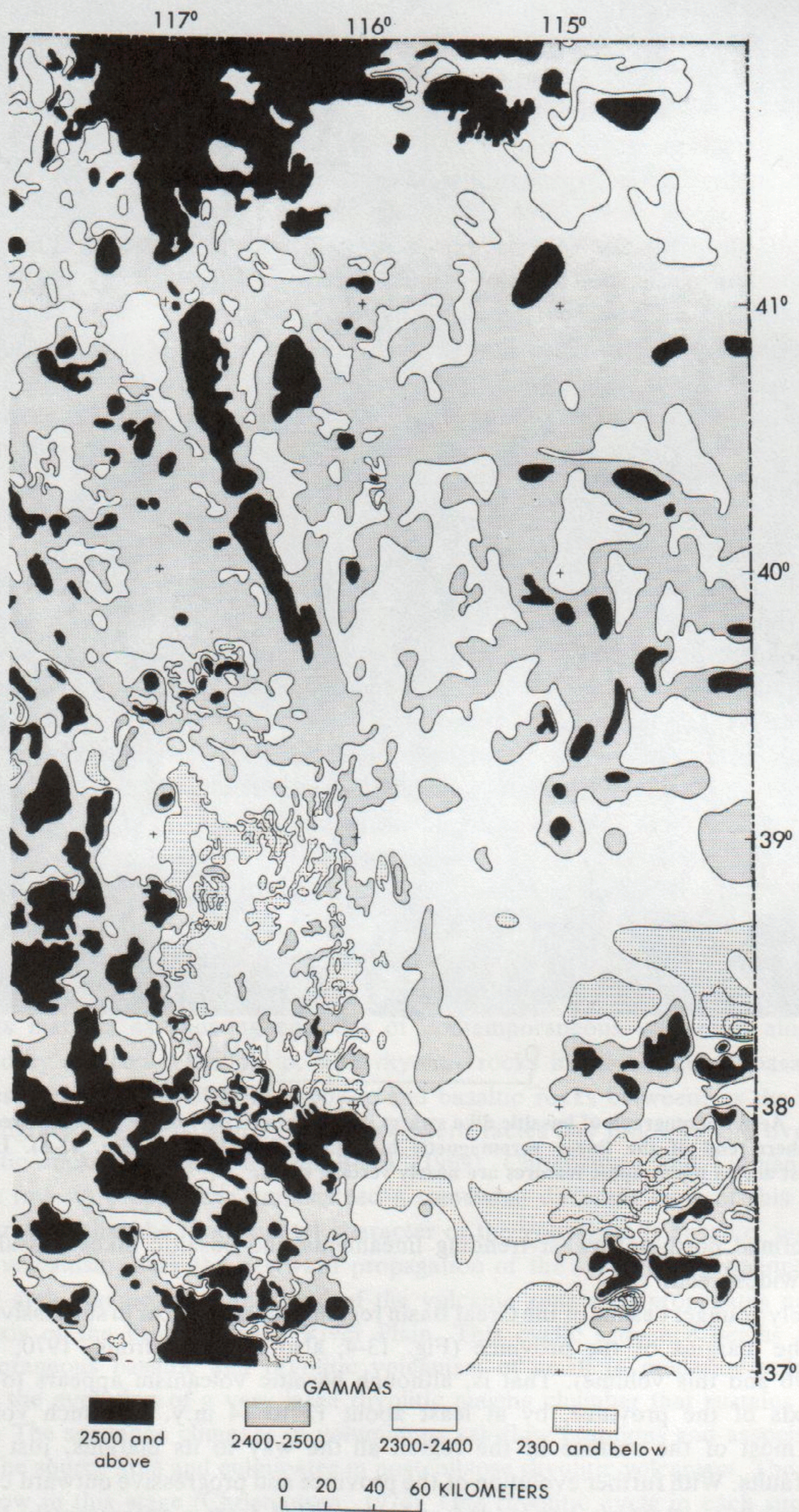
Figure 13-5. Aeromagnetic map of central and northeastern Nevada. Note the similarity between the outcrop pattern of basalts (Fig. 13-4) and many of the aeromagnetic highs, especially the north-trending belt in the central to northern part of the state. (After Stewart and others, 1977.)

Figure 13-6. Aerial photograph of basaltic dike swarm in the western part of the Roberts Creek Mountains, near the southern end of the linear aeromagnetic high of central Nevada (Fig. 13-5). Lines trending north-northwest across topographic features are nearly vertical dikes.

of poorly defined north-northwest–trending lineaments and basaltic dikes. Basaltic flows of this age are widespread.

Successively younger basalts of the Great Basin region generally occur in successively narrower belts near the margins of the province (Fig. 13-4; also see Armstrong, 1970; Stewart and Carlson, 1976 and this volume). That is, although basaltic volcanism appears to have begun along the axis of the province, by at least about 12 to 14 m.y. ago such volcanism was active over most of the surface of the region all the way to its margins, just as were the basin-range faults. With further evolution of the province and progressive outward concentration of faulting, basaltic volcanism has tended to be excluded from a progressively larger central region and to be concentrated in successively narrower zones around the margins of the Great Basin.

Most basalts of the Great Basin region are alkali-rich (generally K-rich) types (Leeman and Rogers, 1970). Some tholeiitic basalts occur near the province boundaries.

Upper Miocene and younger silicic rocks are less abundant than basaltic rocks in most of the Great Basin, but the volcanic rocks as a whole form a bimodal rhyolite-basalt suite. Rhyolitic volcanic rocks younger than 17 m.y. are particularly abundant in a belt of caldera-related source areas along the southern margin of the region (Noble, 1972), including the Kane Wash caldera (Noble, 1968), the Timber Mountain caldera complex (Christiansen and others, 1977), the Black Mountain caldera (Christiansen and Noble, 1965, 1968; Noble and Christiansen, 1968, 1974), the Long Valley caldera (Bailey and others, 1976), and the Little Walker volcanic center (Noble and others, 1974). Abundant silicic volcanic rocks also are associated with the basaltic rocks of the northern Great Basin (Willden, 1964; McKee and Silberman, 1970; Noble and Parker, 1975; McKee and others, 1976; Stewart and Carlson, 1976 and this volume). Minor rhyolites are associated with basalts in other localities such as the Mineral Range and Cove Fort areas of Utah (Lipman and others, 1975). The ages of these silicic rocks span the same ranges as the basalts with which they are areally associated.

Snake River Plain–Yellowstone Plateau

The Snake River Plain is a predominantly basalt-covered belt across Idaho, but along the margins of the plain are both basaltic and rhyolitic volcanic rocks that mainly predate the rocks of the plain itself. The plain comprises two segments. The northwest-trending western segment consists of younger basalts superimposed across a zone of middle and upper Miocene rhyolitic and basaltic volcanic rocks (Malde and Powers, 1962; McIntyre, 1972; Mabey, 1976). By contrast, stratigraphic and geochronologic relations among volcanic rocks marginal to the eastern segment show that major volcanism along it has propagated from southwest to northeast (Armstrong and others, 1975). This volcanism began at least 14 m.y. ago in the region of southwestern Idaho and has migrated northeastward at a rate of several centimetres per year to form a linear zone (Christiansen and Blank, 1969; Armstrong and others, 1975). The youngest part of this propagating zone is the Yellowstone Plateau, which overlies a very large active rhyolitic magma chamber (Christiansen, 1974; Eaton and others, 1975).

Armstrong and others (1975) interpreted geochronologic data for the eastern Snake River Plain and its margins as showing episodes of contemporaneous volcanism along the plain characterized by a facies relationship, with rhyolitic rocks in the northeast, basaltic rocks to the southwest, and an area of both rhyolitic and basaltic rocks between; as the northeastern facies has migrated northeastward, the southwestern facies has followed and overprinted the older rhyolitic rocks. Christiansen and Blank (1969), Christiansen and Lipman (1972), and Christiansen (ms. in prep.) have emphasized a somewhat different view of this relationship; they have stressed that the fundamental character of the volcanism is basaltic, but that cycles of rhyolitic volcanism mark the eastward propagation of the system. A sequence is repeated in each new area that becomes the head of the volcanic zone as it propagates northeastward along the axis of the eastern Snake River Plain. This cyclic sequence begins with basaltic or contemporaneous basaltic and rhyolitic volcanism of small to moderate volume. This is followed by the evolution of a very large rhyolitic magma chamber that sustains the eruption of rhyolites. The sequence climaxes in voluminous ash-flow eruptions and associated caldera-collapse of the source area and culminates in postcollapse rhyolitic volcanism. The Yellowstone Plateau is now in this stage (Christiansen, 1974). One rhyolitic cycle in each successive area has a duration on the order of 1 to 2 m.y., during which time basaltic volcanism continues around the margins but not within the principal rhyolitic source area. A rhyolitic cycle ends

with solidification and fracturing of the large rhyolitic magma chamber; this allows basalts to erupt through the former rhyolitic source area, accompanied only occasionally by small volumes of rhyolite or mixed basalt-rhyolite complexes. Island Park, just west of Yellowstone, is now in this stage (Hamilton, 1965; Christiansen, 1975). Ultimately, as the head of the volcanic zone propagates eastward past an area, the activity in that area reverts to continued basaltic volcanism while the axis of the volcanic zone subsides, as on most of the Snake River Plain.

The basalts of the Snake River Plain–Yellowstone region are predominantly olivine tholeiites (Stone, 1967; Hamilton, 1963, 1965; Tilley and Thompson, 1970; Christiansen, ms in prep.). The younger tholeiites that flood the subsided axis of the Snake River Plain are generally somewhat more potassic and iron-rich than the earlier tholeiites associated more closely with rhyolitic volcanism, now exposed on the plains margins and around Yellowstone.

Oregon High Lava Plains

Basaltic flows are widespread in southeastern Oregon, blanketing large parts of the region and locally, as at Steens Mountain, forming sequences 1,000 m or more thick. More typically, one or a few flows form mesas of 10 km^2 or less. The basaltic vents are not commonly seen, but where observed, they generally form north- or northwest-trending dike swarms. In most of the region, basalts are the youngest volcanic rocks, but they occur throughout the upper Cenozoic sequence as well. The oldest basalts are about 17 to 16 m.y. old and are known from the Steens Mountain area eastward and southeastward toward Idaho and northern Nevada. The Quaternary basalts of central and eastern Oregon (Fig. 13-4B) are restricted to a zone slightly oblique to the trend of the Brothers fault zone, extending from the Newberry volcano to near the Oregon-Idaho border.

Rhyolitic domes, lava flows, and ash flows crop out widely in southeastern Oregon (Fig. 13-7). More than 100 dome complexes have been recognized between Newberry volcano and the Owyhee Plateau in western Idaho (MacLeod and others, 1976). These domes are distributed more or less equally across the region with local concentrations such as near Newberry. Rhyolitic ash-flow tuffs are associated with some of the dome complexes, especially near the Harney basin, about midway between Newberry and the Owyhee Plateau (Greene and others, 1972; Walker, 1973). The combined volume of rhyolite in these ash-flow sheets probably is greater than 300 km^3.

The oldest rhyolitic rocks lie mainly in eastern Oregon, western Idaho, and north-central Nevada and have ages in the 17- to 14-m.y. range. A few occur in central Oregon as well. The younger rhyolitic domes, flows, and tuffs record progressively younger periods of volcanism from southeast to northwest across Oregon, ranging from about 11 m.y. east of the Harney basin to less than 1 m.y. at Newberry volcano (Fig. 13-7; also see MacLeod and others, 1976). The succession of rhyolitic domes defines a broad belt more than 250 km long trending N75°W across southeastern Oregon in which the age progression is well defined and uniform. The rate of propagation calculated by MacLeod and others (1976) is about 1 cm/yr in the western part of the belt, where the domes are <1 to 5 m.y. old, and about 3 cm/yr along the central and eastern parts of the belt, where the domes are 5 to more than 11 m.y. old.

This west-northwest–trending belt of progressively younger rhyolitic rocks virtually mirrors the northeastward rhyolitic progression of the eastern Snake River Plain and Yellowstone, although the eruptive volumes are less in Oregon. The lengths of the belts, rates of propagation of rhyolitic volcanism, and general age spans are similar in the two belts, and the starting area for both is in the Owyhee region near the Idaho-Oregon-Nevada boundary. Geochronologic

evidence indicates similar successions of events in the two propagating volcanic systems. In Oregon as in the eastern Snake River Plain and Yellowstone, both basaltic and rhyolitic volcanism occurred at an early stage. They are closely associated now in the very young Newberry volcano. Rhyolitic volcanism occurs in a given area for only about 1 m.y. or less after the initial volcanism, but basaltic activity has continued all along the principal volcanic axes of both systems through Quaternary time.

Columbia Plateau

The Columbia River Group forms by far the largest Cenozoic basalt field in North America. It represents the northern part of the fundamentally basaltic volcanic association of the Great Basin and Columbia Intermontane region. The basaltic flows commonly are relatively thick (20 m or more) and very widespread, some occurring tens or even hundreds of kilometres from their linear fissure vents (Wright and others, 1973; Swanson and others, 1975). That individual flows can be traced so far attests to the subdued topography of the region at the time of eruption. Many flows ponded in the Pasco basin, in the center of the Columbia Plateau; this indicates continued subsidence of the basin during the volcanism. Within a region about 300 km across there is an estimated 200,000 km³ of erupted basalts. A variety of compositional types is present, but all are tholeiitic (Wright and others, 1973; McDougall, 1976).
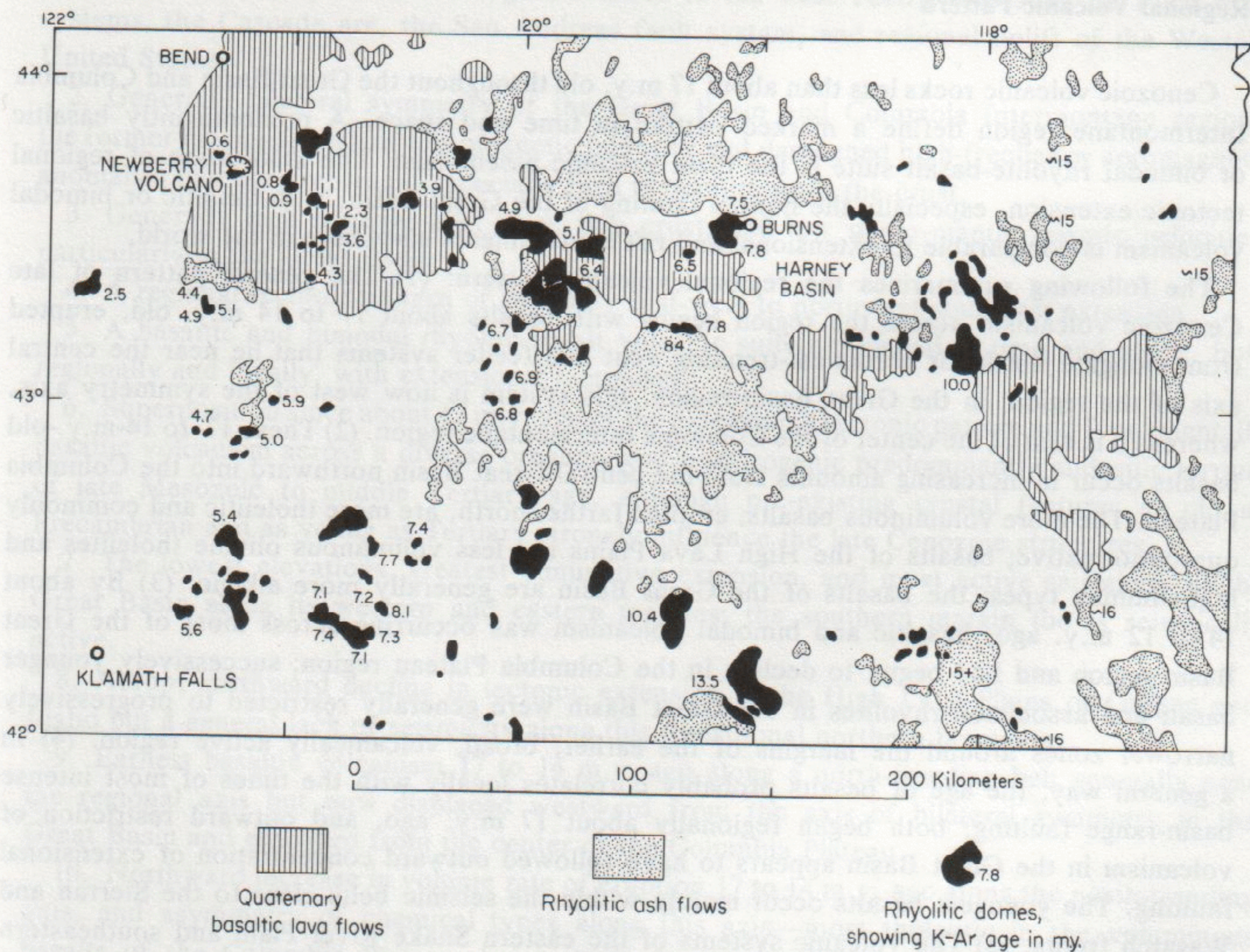

Figure 13-7. Map showing distribution of upper Cenozoic volcanic rocks in south-central Oregon. Rhyolite domes and ash-flow tuffs decrease in age from about 17 m.y. in the southeast to less than 1 m.y. in the northwest. (After MacLeod and others, 1976.)

Known feeder dikes are found only in the central and eastern part of the Columbia Plateau. Most of these dikes occur in the Monument dike swarm in north-central Oregon and the Chief Joseph dike swarm of northeastern Oregon and adjacent parts of Washington and Idaho. Within these large swarms, vents for specific flows are not easily recognized because of subsequent burial by younger flows. However, several vent systems for regionally extensive flows or groups of flows have been recognized; they form linear systems that are tens of kilometres long and a few kilometres wide (for example, see Swanson and others, 1975). These include the Roza vent system near the east edge of the Columbia Plateau and the Ice Harbor system near the center of the plateau. Other linear vents are inferred from flow-outcrop patterns, alignments of cinder and spatter cones, and local accumulations of thin pahoehoe flows. All the known vents and vent systems are parallel, trend roughly north to northwest, and are located east of the center of the Columbia Plateau.

Most of the Columbia River Group has been determined by K-Ar dating to be between about 16 and 14 m.y. old (Watkins and Baksi, 1974). Successively erupted volumes generally became successively smaller although with some exceptions. A relatively small amount of basalt that forms intracanyon flows along the Snake River floods the center of the Pasco basin (Wright and others, 1973; Swanson and others, 1975) and is as young as about 6 m.y. (McKee and others, 1977), but these final basaltic eruptions marked a protracted waning stage of volcanism.

Regional Volcanic Pattern

Cenozoic volcanic rocks less than about 17 m.y. old throughout the Great Basin and Columbia Intermontane region define a marked pattern in time and space. A predominantly basaltic or bimodal rhyolite-basalt suite is the main volcanic assemblage. The association of regional tectonic extension, especially the normal faulting of the Great Basin, with basaltic or bimodal volcanism is comparable to extensional and rift environments elsewhere in the world.

The following summarizes the regional volcanic pattern: (1) The general pattern of late Cenozoic volcanism across the region begins with basalts about 17 to 14 m.y. old, erupted from elongate north- or northwest-trending vent and feeder systems that lie near the central axis of the region; in the Great Basin region, this system is now west of the symmetry axis, whereas it is east of the center of the Columbia Intermontane region. (2) These 17- to 14-m.y.-old basalts occur in increasing amounts from the central Great Basin northward into the Columbia Plateau. The more voluminous basalts, erupted farther north, are more tholeiitic and commonly quartz-normative; basalts of the High Lava Plains are less voluminous olivine tholeiites and high-alumina types; the basalts of the Great Basin are generally more alkalic. (3) By about 14 to 12 m.y. ago, basaltic and bimodal volcanism was occurring across most of the Great Basin region and had begun to decline in the Columbia Plateau region; successively younger basalt and associated rhyolites in the Great Basin were generally restricted to progressively narrower zones around the margins of the earlier, broad, volcanically active region. (4) In a general way, the age of basalts probably correlates locally with the times of most intense basin-range faulting; both began regionally about 17 m.y. ago, and outward restriction of volcanism in the Great Basin appears to have followed outward concentration of extensional faulting. The youngest basalts occur mainly within the seismic belts close to the Sierran and Wasatch fronts. (5) The volcanic systems of the eastern Snake River Plain and southeastern to west-central Oregon define symmetrical but oppositely directed propagating systems; at an early stage as each of these systems propagated into a new area, the system evolved into a major rhyolitic center, represented now by the active Yellowstone Plateau volcanic

field at the eastern end and the basaltic and rhyolitic Newberry volcano at the western end. (6) The northeastward-propagating system has produced greater volumes of eruptive rocks than the northwestward-propagating system; the major rhyolitic magmatic systems along the Snake River Plain–Yellowstone axis have been of batholithic size. (7) Since about 14 m.y. ago, the two systems have propagated from their common area of origin at about the same rate that the inner margins of the principal zones of Great Basin basaltic volcanism have retreated outward from the center. Basalts, however, continue to erupt along the axes of both propagating systems.

A MODEL FOR VOLCANO–TECTONIC EVOLUTION

Background

We propose here a qualitative conceptual model for late Cenozoic volcanic and tectonic evolution of the Great Basin region, the Snake River Plain–Yellowstone system, the plateaus of southern Oregon, and the Columbia Plateau. Such an ambitious objective can only be approached in the most tentative manner, yet such models can be worthwhile in helping to focus studies toward an understanding and integration of tectonic and magmatic histories of the region. The principal relations that our model seeks to represent are as follows:

1. The position of the whole region relative to the East Pacific and Juan de Fuca ridge systems, the Cascade arc, the San Andreas fault system, and regional uplift of the Western United States.

2. Generally bilateral symmetry of the Great Basin and Columbia Intermontane regions, the former having an axis of low magnetic intensity and dampened high-frequency aeromagnetic anomalies that may represent especially high temperatures in the crust.

3. Generally high regional heat flow, thin crust, and low upper-mantle seismic velocities, particularly in the Great Basin.

4. A regional tectonic pattern of generally east-west to northwest-southeast extension.

5. A basaltic and bimodal rhyolite-basalt volcanic suite associated in time and space, both regionally and locally, with extensional tectonism.

6. Superposition since about 17 m.y. ago of the extensional tectonic pattern and fundamentally basaltic volcanism across a diverse orogenic and postorogenic predominantly andesitic terrane of late Mesozoic to middle Tertiary age, although pre-existing crustal features as old as Precambrian and as young as Tertiary strongly influence the late Cenozoic structures.

7. The lowest elevations, greatest cumulative extension, and most active seismicity of the Great Basin along its western and eastern margins; the southern margin too is seismically active.

8. Abrupt northward decline in tectonic extension at the High Lava Plains of Oregon and Idaho but a general lack of seismicity along this transitional northern boundary.

9. Earliest basaltic volcanism 17 to 14 m.y. ago along a north-trending belt generally near the regional axis but now displaced westward from the axis of bilateral symmetry in the Great Basin and eastward from the center of the Columbia Plateau.

10. Northward increase in volume rate of eruption 17 to 14 m.y. ago along the north-trending axis, and asymmetry of chemical types along this axis—most tholeiitic in the voluminous basalts of the Columbia Plateau, most alkalic in the less voluminous basalts of the Great Basin.

11. General decline in basaltic volcanism in the Columbia Plateau after about 14 m.y. ago and nearly total lack of such volcanism after about 8 m.y. ago.

12. Since about 14 m.y. ago, progressive restriction of Great Basin basaltic volcanism to successively narrower marginal zones with time apparently coincided with a outward concentration of major extensional faulting.

13. Propagation of rhyolitic volcanism both northwestward and northeastward after about 14 m.y. ago along the High Lava Plains from a starting area across the north-trending axis of earlier volcanism, apparently at about the same rate as the outward concentration of faulting and volcanism toward the margins of the Great Basin.

14. Occurrence of major rhyolitic volcanism in the propagating systems of the High Lava Plains within only about the first 2 m.y. of the initial volcanic pulse in each new area but continuation of basaltic volcanism along both belts of the High Lava Plains in the wake of volcanic propagation.

15. Much greater volume productivity of the Snake River Plain–Yellowstone propagating system than of the Oregon system.

16. Parallelism of the Snake River Plain–Yellowstone axis with two directions of possible tectonic significance—(a) a regional structural feature of the Precambrian basement along this axis that continues beyond it in both directions (Eaton and others, 1975) and (b) the probable direction of motion of the North American plate relative to the asthenosphere if the Hawaiian melting anomaly is assumed to mark the motion of the Pacific plate relative to the asthenosphere (compare with Smith and Sbar, 1974; Shaw and Jackson, 1973; Suppe and others, 1975).

Previous Concepts

Models applied recently to all or parts of the late Cenozoic volcanic and tectonic evolution of the region are considered in some detail in this section. They can be considered in two general categories:

1. Models relating this evolution to contemporary interactions of the North American, Pacific, and Farallon (or its remnants) plates; these have sometimes been called "passive" models.

2. Models based upon an active upwelling beneath the North American lithospheric plate; these models can be further subdivided by the primary mechanism envisioned for upwelling—(a) overriding by the North American plate over the East Pacific Rise, which is pictured as a convective source of energy and spreading motion; (b) buoyant upward displacement of previously subducted lithosphere of the Farallon and Pacific plates, perhaps after cessation of subduction (active "back-arc spreading" or "ensialic marginal basin"); and (c) uplift and rifting over a rising central thermal anomaly ("hot spot," "convection plume," or "chemical plume") or in the wake of one or more "hot spots" that are laterally stable relative to the lower mantle but in motion relative to the North American plate.

In this paper we adopt a plate-interaction model, but first we consider some arguments concerning models that feature primary upwelling beneath the lithosphere. The concept of an overridden East Pacific Rise, once widely popular (Menard, 1964; Cook, 1969; McKee, 1971), is no longer tenable from a plate-tectonics view of the role of spreading mid-oceanic ridges. The ridges are now recognized as results rather than causes of plate divergence; they represent merely thinning of the lithospheric plate and accretion of new lithospheric material at the divergent plate boundary (for example, Le Pichon and others, 1973). The East Pacific Rise has not been subducted and overridden by the North American plate. Rather, the geometrical and physical constraints of relative motion when the Pacific and American plates came in direct contact caused cessation of the processes that were active at both the rise and the subduction zone. Both ceased to exist as functional entities and were replaced by two triple

junctions and a transform fault system (McKenzie and Morgan, 1969; Atwater, 1970).

The concept of the Great Basin as an ensialic marginal basin is currently popular, but several arguments persuade us that it is less attractive than a plate-interaction model. One of the original suggestions for such an active "back-arc spreading" model came from Scholtz and others (1971) on the basis of a concept advanced by Armstrong and others (1969) that Cenozoic volcanism in the Great Basin began in a core area about 40 m.y. ago and has since migrated outward toward the margins. That history, however, especially before 17 m.y. ago, has been shown to be incorrect (see McKee and others, 1976; Stewart, this volume; Stewart and others, 1977). Scholtz and others (1971) further suggested that block faulting moved outward in a similar manner, but we have cited evidence earlier in this paper that major faulting near the present margins of the Great Basin is at least 10 to 14 m.y. old. Also, the region of the proposed ensialic marginal basin, as several papers in this symposium volume emphasize, is a region mainly of uplift, not of subsidence like the supposedly analogous features of the western Pacific (Karig, 1971). By comparison, subduction along the margin of South America has continued beneath a continental plate longer than it did beneath North America but has not resulted in a major episode of back-arc spreading; thus, such a process is not, as sometimes proposed, an inevitable consequence of longevity in continental-margin subduction systems. Triggering of back-arc spreading by the cessation of compressive subduction (as suggested by Scholtz and others, 1971) seems problematical because of the relative timing of these events. Regional extension and basaltic and bimodal volcanism did begin in the southern part of the Basin and Range province and Southern Rocky Mountains at the time of cessation of subduction, but this tectonic and volcanic activity migrated with time as subduction ceased over a widening region (Christiansen and Lipman, 1972). However, similar processes affected the entire Great Basin and Columbia Intermontane regions simultaneously while subduction continued in a shrinking zone to the west (Christiansen and Lipman, 1972; Snyder and others, 1976).

If there were one part of this entire volcano-tectonic system that was in some ways analogous to an oceanic marginal basin, it would be the Columbia Plateau, which was a region of extension, voluminous tholeiitic volcanism, and subsidence for a period of several million years. Elsewhere the concept does not fit well.

"Hot spot" models are too diverse to discuss succinctly in general. However, the most commonly held forms of those models argue that a melting anomaly at Yellowstone (or at Yellowstone and at Raton in northeastern New Mexico) represents a convection plume from the deep mantle (Morgan, 1972; Matthews and Anderson, 1973; Smith and Sbar, 1974; Suppe and others, 1975). One of us (Christiansen, 1973, and ms. in prep.) has shown that such a model does not explain well the geologic, petrologic, or regional tectonic data for the Yellowstone–Snake River Plain region. In particular, (1) the mantle-plume model does not account for spontaneous origin of the melting anomaly between 17 and 14 m.y. ago at the same time as the inception of regional extension; (2) it does not explain the occurrence of volcanism not merely as a propagating "hot spot" but also as continuing basaltic activity for at least 14 m.y. along the trace; (3) it does not account for the simultaneous and symmetrical northeastward and northwestward propagation of rhyolitic volcanism along the eastern Snake River Plain and across central Oregon; and (4) although the eastern Snake River Plain correlates in orientation with the kinematic prediction of the mantle-plume model, the orientation also correlates with the prediction of the shear-melting concept and with an ancient structural feature of the crust.

Another variant model places a broader "hot spot" at the junction between the two propagating volcanic axes of the High Lava Plains and the central axis of the Great Basin; this model

derives the two axes of volcanic propagation, the Great Basin, and the Columbia Plateau as tensional rifts across a resulting uplift (Prostka and others, 1976). In our view, such a model fails to relate closely either to (1) the Great Basin as a wide region of extension rather than as a linear rift or (2) the decline of the voluminous early basaltic volcanism in the Columbia Plateau just when the rhyolitic systems of the High Lava Plains began to propagate. Furthermore, since such a system could not reflect a deep-mantle convection plume unless that plume were migrating synchronously with the North American plate, Prostka and others (1976) proposed that an anomalous concentration of radiogenic heat-producing elements in the lithosphere drives the upwelling. That concentration presumably was the result of a "chemical plume" from the deep mantle before or during formation of the lithospheric plate. Two points argue against this concept, in our minds. First, there is no record of this chemical anomaly before 17 m.y. ago in any pattern that is coherent with the effects ascribed to it since. Second, this feature, presumably derived from some ancient deep upwelling, is centered in an area where regional geology and geophysics show that the lithosphere has grown in several accretionary events. East of the central zone of the proposed chemical plume—in the Yellowstone region and its surroundings—is an area of very old Precambrian crust and upper mantle; younger Precambrian basement underlies areas farther northwest in Montana and eastern Washington; farther west and southwest the lithosphere formed during Mesozoic arc-type magmatism and tectonism; the Columbia Plateau on the north side of the proposed chemical anomaly probably was oceanic lithosphere in Mesozoic time.

Basis of a Plate-Interaction Model

The foregoing arguments help to persuade us against any of the specific models proposed for dynamic upwelling from the asthenosphere or the deep mantle; our underlying reason for preferring a plate-interaction model, however, is that it can be viewed as a relatively simple, even inevitable consequence of the tectonic interactions between moving lithospheric plates. In this vein, we accept in their most general form the concepts of Atwater (1970) that the region of late Cenozoic tectonic extension is a diffuse zone of transform displacement between the Pacific plate and the tectonically stable part of the North American plate. Our model builds on Atwater's view that a region of oblique extension, having components both of rifting and of lateral shear, has resulted as a migrating triple junction caused the previously intervening Farallon plate to cease interacting with the Pacific and North American plates in a progressively widening zone. We also follow the general concepts of Christiansen and Lipman (1972) that "fundamentally basaltic" magma generation in this region of tectonic extension is a consequence of that extension and that the Mesozoic and Cenozoic thermal and stress histories of each part of the region determine just how that part responds to superposition of the extensional stress system.

We wish to emphasize that regional extension in such a model does not require that shear stresses generated at the San Andreas fault be transmitted across the entire region, as some critics of plate-interaction models have supposed. In most plate-tectonics models, regardless of the motive forces called upon to drive the plates, it is presumed that each plate is in some state of stress that characterizes it throughout. Although the stress field within a plate may vary in space and time and although there are pre-existing anisotropies in the continental lithosphere, each part of each plate reflects in some way the stresses that are imposed upon it by the motions of the plates and by their mutual interactions. The stresses are resolved at the plate boundaries in ways that are consistent with geometric and physical constraints of the relative plate motions. When the Pacific and North American plates began to interact,

probably about 29 m.y. ago (Atwater and Molnar, 1973), the stresses along their mutual boundary began to reflect both the narrow zone of contact and the continuing interactions with the two large remnants of the Farallon plate, which was being subducted. At present, there is a wide zone of Pacific–North American plate interaction and a very narrow zone of weak interaction between the North American plate and the Juan de Fuca plate (the northern remnant of the Farallon plate). The temporal change between these conditions resulted in an altered stress field within the North American plate. The new stress field causes the regional extension shown by the course of late Cenozoic tectonic and magmatic evolution of the Cordilleran region.

The direct result of the extensional tectonic regime in the past 25 m.y. or so has been a chain of heating events. A few million years probably were required for thermal equilibration of the plates after initial contact of the Pacific and North American plates and local cessation of plate-margin spreading and subduction (Atwater, 1970). With the change from subduction to tectonic extension, stress at the base of the North American lithosphere is reduced; this facilitates partial melting of upper-mantle peridotites that were initially part of the lithosphere, and basaltic magmas are produced. In this model, a relatively small percentage of partial melting at considerable depths produces the alkali-rich basalts that characterize much of the region. Generally only at the boundaries between tectonic provinces has there been greater extension, locally enough to produce higher percentages of partial melting, perhaps at shallower depths in the upper mantle, to generate tholeiitic magmas. The basaltic magmas tend to rise buoyantly into and through the crust where favorable structures or particularly large deviatoric stresses accommodate this rise. Thus, the magmas erupted mainly in areas where extensional faulting was active simultaneously. Elsewhere these magmas would have tended to intrude the crust and to reside there for long periods. The presence of basaltic melts in the upper mantle and crust—both passing locally upward through the crust and residing in the crust over much of the region—has increased the regional heat flow, lowered the rigidity of the lithosphere, caused further extension, thinned the crust, produced thermal expansion and regional uplift, caused isostatically compensating mantle flowage, and lowered seismic velocities in the upper mantle.

The actual course of volcanic and tectonic evolution of each part of the Cordilleran region has varied, depending partly on the tectonic response of that region as conditioned by its preceding stress and thermal history. The zone of extension in the North American plate started at the continental margin of Mexico and southern California about 29 m.y. ago (Atwater and Molnar, 1973) and produced its first notable effects within the basin-range region (Christiansen and Lipman, 1972; Armstrong and Higgins, 1973; Snyder and others, 1976). During the period of 25 to 17 m.y. ago, only those areas that were directly inland from the coastal transform zone and that had undergone previous Laramide compressive deformation and Laramide or early to mid-Tertiary intermediate to silicic volcanism subsequently underwent extensional normal faulting and basaltic volcanism. These were the Southern Rocky Mountains and the southern part of the present Basin and Range province in New Mexico, Arizona, Sonora, and southeastern California; the Colorado Plateau remained unaffected. A pre-existing structural grain from Paleozoic and Mesozoic deformations has forced a pattern of oblique extension in the Great Basin during the past 17 m.y.

The approximate northern boundary of the region of maximum cumulative extension coincides in its eastern portion with an ancient structural alignment. With increasing extension in the Great Basin after about 14 m.y. ago, this alignment and a symmetrical tear to the west appear to have acted as a boundary for that extension. Whereas cumulative extension across the central and northern Great Basin probably is more than 100 km, comparable extension in

areas north of the High Lava Plains is generally a few kilometres and, at most, less than a few tens of kilometres. Thus, the High Lava Plains are a transitional transform boundary zone of the Great Basin. In the same way that Atwater's basin-range model represents a "soft" plate margin, our model emphasizes the axis of the High Lava Plains as a "soft" edge to the zone of oblique extension that lies mainly to the south (Fig. 13-8). This transitional boundary is essentially aseismic and is characterized along its entire length, especially in the Snake River Plain, by Quaternary volcanic fields, presumably indicative of high crustal temperatures. The transitional zone continues northward from the eastern Snake River Plain to the Idaho seismic belt.

Interpretation of Late Cenozoic Volcano-Tectonic Evolution

By 17 m.y. ago the remnant volcanic arc along the continental margin was restricted to a narrow belt north of the latitude of southern or south-central Nevada (Fig. 13-8). For a long time before 17 m.y. ago, the eastern Pacific spreading ridge probably had intersected the North American continental margin near Vancouver Island (Atwater, 1970). Therefore, by about 17 m.y. ago, the remnant arc and subduction system was no more than about 2,200 km long and was bracketed at both ends by transform boundaries between the Pacific and North American plates. The stresses that resulted within the North American plate from this changing configuration of relative plate motions and interactions caused the western part of the North American plate to be dominated by the stress field resulting from the growing
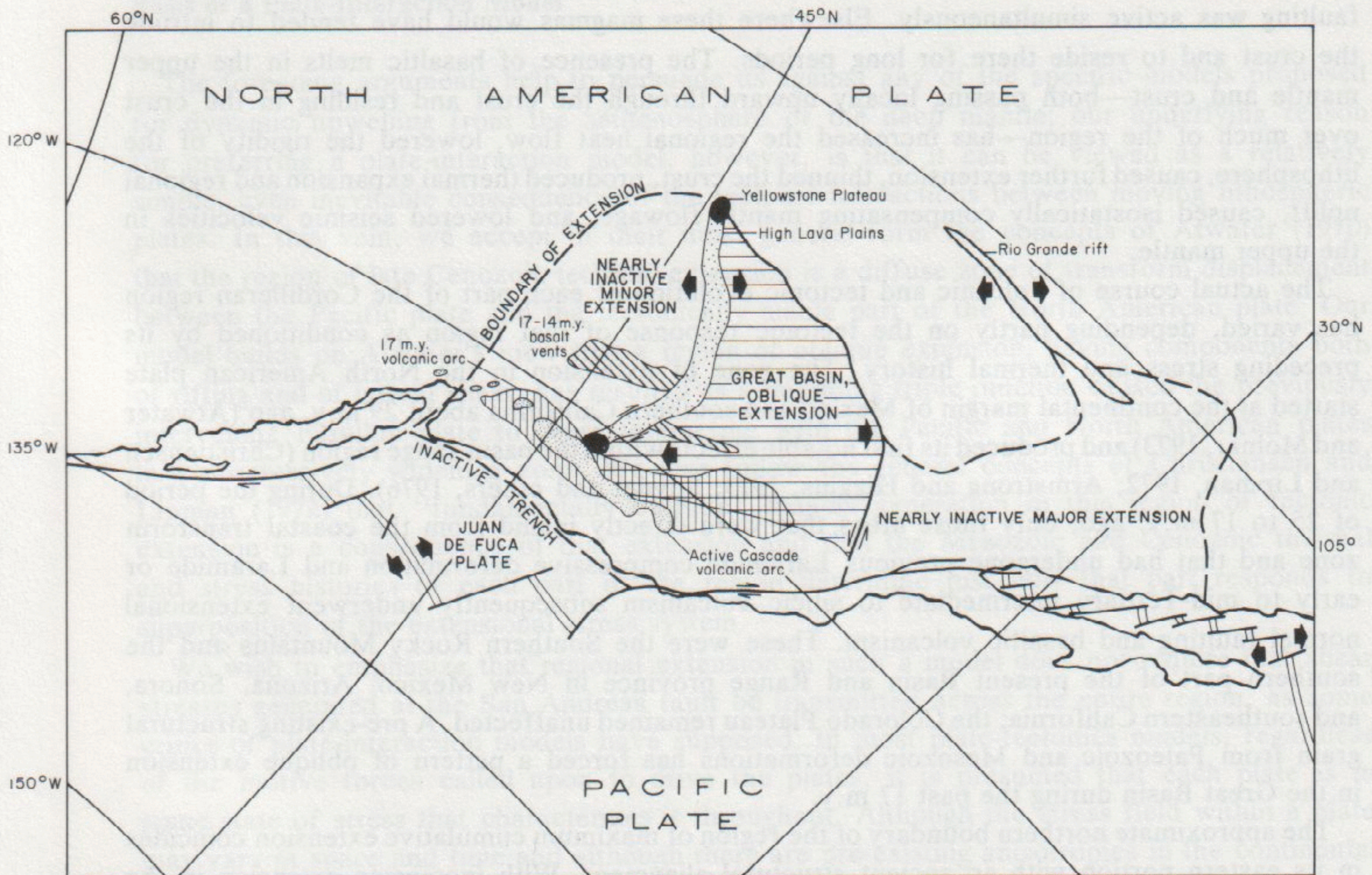

Figure 13-8. Map of the Western United States outlining the late Cenozoic tectonic features and upper Cenozoic volcanic rocks in relation to Pacific–North American plate interaction. Shown on the oblique Mercator projection of Atwater (1970); horizontal lines of the projection are directions of pure transform displacement; vertical lines are directions of pure dilation or compression.

Pacific–North American interaction instead of the declining stress field resulting from the Farallon–North American interaction. Under these conditions, in our model, the previously heated and deformed part of the North American plate did not behave in the ideally rigid manner of plate tectonics but allowed partial coupling between the two separated transform zones. This forced extension across an axis roughly parallel to the continental-margin arc and about 300 to 400 km inland from it (Fig. 13-8). Extension roughly perpendicular to this axis opened a linear rift that was continuous from central Nevada, through the western Snake River Plain, to the eastern Columbia Plateau. Subsequently, this rift axis has been offset by transform displacement at the northern boundary of the Great Basin. Stress relief at the base of the lithosphere along this rift produced the initial basaltic magma generation in the upper mantle. Under that part of the lithosphere having a Precambrian crystalline basement or a tectonically thickened Paleozoic geosynclinal accumulation, slight partial melting at considerable depth produced alkali-rich basalts. Farther north, where the 17- to 14-m.y.-old volcanic axis crossed the Mesozoic plutonic-orogenic belt, larger volumes of melt were produced at shallower levels and less-alkalic basalts, commonly of high-alumina character, were erupted. In the Columbia basin, according to the model, basalts which formed voluminous tholeiitic floods were generated by high percentages of partial melting under a thin crust that had been oceanic in Mesozoic time.

Within a few million years, continued tectonic extension was expressed across the entire region previously heated by early and mid-Tertiary volcanism; regional melting at the base of the lithosphere decreased lithospheric rigidity so that much subsequent deformation took place at considerable depth by thinning of the crust and flowage of the mantle. Under these conditions, much of the basaltic magma accumulated in the crust; this further increased regional heat flow, caused regional uplift, and reduced lithospheric rigidity. Continuously higher temperatures and decreasing rigidity of the lithosphere along the axis of the region resulted in progressive restriction of brittle deformation at upper-crustal levels away from the axial zone. Basaltic eruption occurred mainly in areas that were broken at shallow levels by normal faults. This accounts for the progressive concentration of both faulting (seismicity) and basaltic volcanism toward the Great Basin margins, even though the entire region is under extension and the axial zone, with its low seismicity and aeromagnetic "quiet zone," may even be hotter at depth than the margins.

The approximate northern boundary of the region of maximum cumulative extension (an ancient structural boundary in the east and a symmetrical tear in the west) is a diffuse transform boundary to the region of oblique extension (Fig. 13-8) and localizes the zone of stress relief. The eastern part (the eastern Snake River Plain) is a linear zone of subsidence flanked by uplifts. The western part (the Brothers fault zone and parallel zones) is essentially a right-lateral en echelon system (Lawrence, 1976). Basaltic volcanism along the full boundary zone has tended to narrow the transition with time so that tectonic extension has become nearly inactive in most of the region farther north. Progressive transform displacement along the transitional boundary has tended to displace the 17- to 14-m.y.-old rift axis westward in the Great Basin relative to the vent systems of the Columbia Plateau.

The most voluminous and diverse volcanism has tended to concentrate where the two zones of successively more localized lithospheric extension and stress relief marginal to the Great Basin intersect its northern transform boundary. As the inner edges of these zones have retreated toward the margins, their intersections with the boundary have propagated away from the initial axis of regional extension. At these intersections there has been intensely localized lithospheric extension and a relatively high percentage of melting in the upper mantle to produce significant volumes of tholeiitic magma. Intrusion of this magma into and through

the lower crust has caused localized melting of the lower crust. The rhyolitic magmas thus produced in the lower crust are emplaced to shallow levels and sustain cycles of rhyolitic volcanism. Continued tectonic extension to the south and transform displacement along the High Lava Plains result in continued basaltic magmatism, heating of the crust, and an aseismic shadow within the northern part of the border zone of earthquakes around the Great Basin.

By far the most productive system of magma generation marginal to the Great Basin has been the eastern Snake River Plain–Yellowstone system, along which very large volumes of basaltic and rhyolitic magma have erupted. Christiansen (ms. in prep.) suggests that the generation of large volumes of rhyolitic and basaltic magma by the Yellowstone melting anomaly reflects a system of magma generation that has become self-sustaining in much the same way as the Hawaiian melting anomaly. The Yellowstone melting anomaly is not a "hot spot" in the sense of a fundamental thermal anomaly of the mantle; the geologic history of the Yellowstone melting anomaly cannot readily be reconciled with a deep-mantle convection plume. We thus favor a concept like that proposed by Shaw and Jackson (1973) for Hawaii, plus the additional effect of lower-crustal melting in a continental plate. In our view, the ancestor of the Yellowstone and Newberry systems was initiated where the original axis of extension and basaltic magma generation from the Great Basin through the Columbia Plateau intersected the structurally controlled transform northern boundary zone of the Basin and Range province. The melting anomalies have been augmented by concentration of extension and stress relief at this intersection as the central axis was replaced by two zones of progressively more concentrated faulting toward the margins of the Great Basin. The Yellowstone melting anomaly probably has become a self-sustaining system by the accumulated effects, guided by the fortuitously oriented old structural boundary, of (1) shear melting at the base of the lithosphere where relative motion parallels this structure, (2) the onset of a thermal feedback cycle, (3) partial melting in the lower crust as well as the mantle, and (4) development of a deep root, in which a zone of inward flow compensates for downward displacement of a dense unmelted mantle residuum (the "gravitational anchor" of Shaw and Jackson, 1973). This flow replenishes the supply of undepleted mantle for continued basaltic magma generation. The concept of such a root accounts for the seismic observations of Iyer (1975; also 1977, written commun.) that indicate a velocity anomaly to a depth of about 300 km in the mantle beneath Yellowstone. The high-velocity core of the mantle structure that Hadley and others (1976) interpreted beneath Yellowstone would be at least as consistent with the concept of a dense refractory residuum as it is with their hypothesized "chemical plume" for Yellowstone.

A model for the late Cenozoic volcano-tectonic evolution of the Great Basin and Columbia Intermontane regions may be summarized as follows: (1) The late Cenozoic relative motions of three mutually interacting lithospheric plates formed two separate zones of transform displacement between the Pacific and North American plates. (2) By 17 m.y. ago, the remnant subduction zone and continental-margin arc became so short between these transform zones that the two became partially coupled through a previously stressed and heated region behind the arc. This coupling produced an east-west to northwest-southeast minimum compressive stress within part of the North American plate, and the resulting extension caused a rift to open along an axis 300 to 400 km behind the arc. (3) Because of its complex recent tectonic and magmatic history, the response of the continental plate to the lithospheric stress field evolved by 14 m.y. ago to a pattern of oblique extension; an old structural boundary guided development of the transform northern boundary of this oblique extension. (4) Stress relief at the base of the lithosphere in the extensional region causes partial melting in the upper mantle and regional basaltic volcanism. (5) Continued basaltic magma generation and intrusion of the crust by the basaltic magma, in addition to continued extension, cause increased regional

heat flow, progressive thinning of the crust, flowage in the upper mantle, decreasing rigidity of the crust, and thermal expansion with regional uplift; brittle deformation tends to be progressively excluded from the central, hottest part of the region. (6) The consequent progressive concentration of normal faulting toward the Great Basin margins has localized maximum cumulative extension toward the margins and has guided a similar outward narrowing of the zones of most active basaltic and related volcanism. (7) Intersection of these marginal zones with the transform northern boundary of the Great Basin has produced localized zones of intense lithospheric extension, basaltic magmatism, and lower-crustal partial melting to sustain brief cycles of rhyolitic volcanism; the parallelism of part of this transform boundary with the direction of plate motion relative to the asthenosphere has caused the eastern Snake River Plain–Yellowstone system to become a self-sustaining melting anomaly.

ACKNOWLEDGMENTS

REFERENCES CITED

Armstrong, R. L., 1970, Geochronology of Tertiary igneous rocks, eastern Basin and Range province, western Utah, eastern Nevada, and vicinity, U.S.A.: Geochim. et Cosmochim. Acta, v. 34, p. 203–232.

Armstrong, R. L., and Higgins, R. E., 1973, K-Ar dating of the beginning of Tertiary volcanism in the Mojave Desert, California: Geol. Soc. America Bull., v. 84, p. 1095–1100.

Armstrong, R. L., Ekren, E. B., McKee, E. H., and Noble, D. C., 1969, Space-time relations of Cenozoic silicic volcanism in the Great Basin of the Western United States: Am. Jour. Sci., v. 267, p. 478–490.

Armstrong, R. L., Leeman, W. P., and Malde, H. E., 1975, K-Ar dating, Quaternary and Neogene volcanic rocks of the Snake River Plain, Idaho: Am. Jour. Sci., v. 275, p. 225–251.

Atwater, Tanya, 1970, Implications of plate tectonics for the Cenozoic tectonic evolution of western North America: Geol. Soc. America Bull., v. 81, p. 3513–3536.

Atwater, Tanya, and Molnar, P. H., 1973, Relative motion of the Pacific and North American plates deduced from sea-floor spreading in the Atlantic, Indian and South Pacific Oceans, in Kovach, R. L., and Nur, Amos, eds., Tectonic prob-lems of the San Andreas fault system, Conf., Proc.: Stanford Univ. Pubs. Geol. Sci., v. 13, p. 136–149.

Bailey, R. A., Lanphere, M. A., and Dalrymple, G. B., 1976, Volcanism, structure, and geochronology of Long Valley caldera, Mono County, California: Jour. Geophys. Research, v. 81, p. 725–744.

Barazangi, M., and Dorman, J., 1969, World seismicity map of ESSA Coast and Geodetic epicenter data for 1961–1967: Seismol. Soc. America Bull., v. 59, p. 369–380.

Blackwell, D. D., 1969, Heat-flow determinations in the northwestern United States: Jour. Geophys. Research, v. 74, p. 992–1007.

Christiansen, R. L., 1973, Rhyolitic ash flows of the Yellowstone Plateau volcanic field—Wyoming, Idaho, Montana—and generation of the rhyolite-basalt association: Geol. Soc. America Abs. with Programs, v. 5, p. 21–22.

——1974, Quaternary volcanism of the Yellowstone rhyolite plateau region, Wyoming-Idaho-Montana [abs]: EOS (Am. Geophys. Union Trans.), v. 56, p. 1189.

——1975, Origin and geothermal potential of Island Park, eastern Idaho: Geol. Soc. America Abs. with Programs, v. 7, p. 595–596.

Christiansen, R. L., and Blank, H. R., Jr., 1969, Volcanic evolution of the Yellowstone rhyolite plateau and eastern Snake River Plain, U.S.A. [abs.], in Symposium on volcanoes and their roots, volume of abstracts: Oxford, England, Internat. Assoc. Volcanology and Chemistry Earth's Interior, p. 220–221.

Christiansen, R. L., and Lipman, P. W., 1972, Cenozoic volcanism and plate tectonic evolution of the western United States—Pt. II, late Cenozoic: Royal Soc. London Philos. Trans., ser. A, v. 271, p. 249–284.

Christiansen, R. L., and Noble, D. C., 1965, Black Mountain volcanism of southern Nevada [abs.]: Geol. Soc. America Spec. Paper 82, p. 246.

——1968, Geologic map of the Trail Ridge quadrangle, Nye County, Nevada: U.S. Geol. Survey Geol. Quad. Map GQ-774.

Christiansen, R. L., Lipman, P. W., Carr, W. J., Byers, F. M., Jr., Orkild, P. P., and Sargent, K. A., 1977, The Timber Mountain–Oasis Valley caldera complex of southern Nevada: Geol. Soc. America Bull., v. 88, p. 943–959.

Cook, K. L., 1969, Active rift system in the Basin and Range province: Tectonophysics, v. 8, p. 469–511.

Eaton, G. P., Christiansen, R. L., Iyer, H. M., Pitt, A. M., Mabey, D. R., Blank, H. R., Jr., and Gettings, M. E., 1975, Magma beneath Yellowstone National Park: Science, v. 188, p. 787–796.

Eaton, G. P., Wahl, R. R., Prostka, H. J., Mabey, D. R., and Kleinkopf, M. D., 1978, Regional gravity and tectonic patterns: Their relation to late Cenozoic epeirogeny and lateral spreading in the western Cordillera, in Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

Greene, R. C., Walker, G. W., and Corcoran, R. E., 1972, Geologic map of the Burns quadrangle, Oregon: U.S. Geol. Survey Misc. Geol. Inv. Map I-680, scale 1:250,000.

Hadley, D. M., Stewart, G. S., and Ebel, J. E., 1976, Yellowstone: Seismic evidence for a chemical mantle plume: Science, v. 193, p. 1237–1239.

Hamilton, Warren, 1963, Petrology of rhyolite and basalt, northwestern Yellowstone Plateau: U.S. Geol. Survey Prof. Paper 475-C, p. C78–C81.

——1965, Geology and petrogenesis of the Island Park caldera of rhyolite and basalt, eastern Idaho: U.S. Geol. Survey Prof. Paper 504-C, 37 p.

Hamilton, Warren, and Myers, W. B., 1966, Cenozoic tectonics of the Western United States: Rev. Geophysics, v. 4, p. 509–550.

Hill, D. P., 1978, Seismic evidence for the structure and Cenozoic tectonics of the Pacific Coast States, in Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

Iyer, H. M., 1975, Anomalous delays of teleseismic P-waves in Yellowstone National Park: Nature, v. 253, p. 425–427.

Karig, D. E., 1971, Origin and development of marginal basins in the western Pacific: Jour. Geophys. Research, v. 76, p. 2542–2561.

Lachenbruch, A. H., and Sass, J. H., 1977, Heat flow in the United States and the thermal regime of the crust: Am. Geophys. Union Geophys. Mon. 20, p. 626–675.

Lawrence, R. D., 1976, Strike-slip faulting terminates the Basin and Range province in Oregon: Geol. Soc. America Bull., v. 87, p. 846–850.

Leeman, W. P., and Rogers, J.J.W., 1970, Late Cenozoic alkali-olivine basalts of the Basin-Range province, U.S.A.: Contr. Mineralogy and Petrology, v. 25, p. 1–24.

Le Pichon, X., Francheteau, J., and Bonin, J., 1973, Plate tectonics: Amsterdam, Elsevier, 300 p.

Lipman, P. W., Prostka, H. J., and Christiansen, R. L., 1972, Cenozoic volcanism and plate-tectonic evolution of the Western United States—Pt. I, early and middle Cenozoic: Royal Soc. London Philos. Trans., ser. A., v. 271, p. 217–248.

Lipman, P. W., Rowley, P. D., and Pallister, J. S., 1975, Pleistocene rhyolite of the Mineral Range, Utah—Geothermal and archeological significance: Geol. Soc. America Abs. with Programs, v. 7, p. 1173.

Longwell, C. R., Pampayen, E. H., Bowyer, B., and Roberts, R. J., 1965, Geology and mineral deposits of Clark County, Nevada: Nevada Bur. Mines Bull. 62, 218 p.

Love, J. D., 1956, Summary of geologic history of Teton County, Wyoming, during late Cretaceous, Tertiary and Quaternary times: Wyoming Geol. Assoc., Guidebook 11, p. 140–150.

Love, J. D., Reed, J. C., Jr., Christiansen, R. L., 1972, Geologic block diagram and tectonic history of the Teton region, Wyoming-Idaho: U.S. Geol. Survey Misc. Geol. Inv. Map I-730.

Mabey, D. R., 1966, Regional gravity and magnetic anomalies in part of Eureka County, Nevada: Soc. Exploration Geophysicists Mining Geophysics Case Histories, v. 1, p. 77–83.

——1976, Interpretation of a gravity profile across the western Snake River Plain, Idaho: Geology, v. 4, p. 53–55.

Mabey, D. R., Zeitz, I., Eaton, G. P., and Kleinkopf, M. D., 1978, Regional magnetic patterns

in part of the Cordillera in the Western United States, *in* Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

MacLeod, N. S., Walker, G. W., and McKee, E. H., 1976, Geothermal significance of eastward increase in age of upper Cenozoic rhyolitic domes in southeastern Oregon, *in* Proc, Second United Nations symposium on the development and use of geothermal resources, Vol. 1; Washington, D.C., U.S. Govt. Printing Office, p. 465–474.

MacNeil, F. S., 1939, Fresh-water invertebrates and land plants of Cretaceous age from Eureka, Nevada: Jour. Paleontology, v. 13, p. 355–360.

Malde, H. E., and Powers, H. A., 1962, Upper Cenozoic stratigraphy of western Snake River Plain, Idaho: Geol. Soc. America Bull., v. 73, p. 1197–1220.

Marvin, R. F., Byers, F. M., Jr., Mehnert, H. H., and Orkild, P. P., 1970, Radiometric ages and stratigraphic sequence of volcanic and plutonic rocks, southern Nye and western Lincoln Counties, Nevada: Geol. Soc. America Bull., v. 81, p. 2657–2676.

Matthews, V., III, and Anderson, C. E., 1973, Yellowstone convection plume and break-up of the Western United States: Nature, v. 243, p. 158–159.

McDougall, I., 1976, Geochemistry and origin of basalt of the Columbia River Group, Oregon and Washington: Geol. Soc. America Bull., v. 87, p. 777–792.

McIntyre, D. H., 1972, Cenozoic geology of the Reynolds Creek experimental watershed, Owyhee County, Idaho: Idaho Bur. Mines and Geology Pamph. 151, 115 p.

McKee, E. H., 1971, Tertiary igneous chronology of the Great Basin of the Western United States—Implications for tectonic models: Geol. Soc. America Bull., v. 82, p. 3497–3502.

McKee, E. H., and Noble, D. C., 1974, Timing of late Cenozoic crustal extension in the Western United States: Geol. Soc. America Abs. with Programs, v. 6, p. 218.

McKee, E. H., and Silberman, M. L., 1970, Geochronology of Tertiary igneous rocks in central Nevada: Geol. Soc. America Bull., v. 81, p. 2317–2328.

McKee, E. H., Noble, D. C., and Silberman, M. L., 1970, Middle Miocene hiatus in volcanic activity in the Great Basin area of the western United States: Earth and Planetary Sci. Letters, v. 8, p. 93–96.

McKee, E. H., Tarshis, A. L., and Marvin, R. K., 1976, Summary of radiometric ages of Tertiary volcanic and selected plutonic rocks in Nevada—Pt. V: Northeastern Nevada: Isochron/West, no. 16, p. 15–27.

McKee, E. H., Swanson, D. A., and Wright, T. L., 1977, Duration and volume of Columbia River Basalt volcanism, Washington, Oregon, and Idaho: Geol. Soc. America Abs. with Programs, v. 9, p. 463–464.

McKenzie, D. P., and Morgan, W. J., 1969, Evolution of triple junctions: Nature, v. 244, p. 125–133.

Menard, H. W., 1964, Marine geology of the Pacific: New York, McGraw-Hill, 271 p.

Morgan, W. J., 1972, Plate motions and deep mantle convection: Geol. Soc. America Mem. 132, p. 7–22.

Noble, D. C., 1968, Kane Springs Wash volcanic center, Lincoln County, Nevada, *in* Eckel, E. D., ed., Nevada Test Site: Geol. Soc. America Mem. 110, p. 109–116.

——1972, Some observations on the Cenozoic volcano-tectonic evolution of the Great Basin, western United States: Earth and Planetary Sci. Letters, v. 17, p. 142–150.

Noble, D. C., and Christiansen, R. L., 1968, Geologic map of the southwest quarter of the Black Mountain quadrangle, Nye County, Nevada: U.S. Geol. Survey Misc. Geol. Inv. Map I-562.

——1974, Black Mountain volcanic center, *in* Guidebook to the geology of four Tertiary volcanic centers in central Nevada: Nevada Bur. Mines and Geology Rept. 19, p. 27–54.

Noble, D. C., and Parker, D. F., 1975, Peralkaline silicic volcanic rocks of the Western United States: Bull. Volcanol., v. 38, p. 803–827.

Noble, D. C., and Slemmons, D. B., 1975, Timing of Miocene faulting and intermediate volcanism in the central Sierra Nevada and adjacent Great Basin: California Geology, v. 28, p. 105.

Noble, D. C., Slemmons, D. B., Korringa, M. K., Dickinson, W. R., Al-Rawi, Y., and McKee, E. H., 1974, Eureka Valley Tuff, east-central California and adjacent Nevada: Geology, v. 2, p. 139–142.

Nolan, T. B., Merriam, C. W., and Williams, J. S., 1956, The stratigraphic section in the vicinity of Eureka, Nevada: U.S. Geol. Survey Prof. Paper 276, 77 p.

Pardee, J. T., 1950, Late Cenozoic block faulting in western Montana: Geol. Soc. America Bull., v. 61, p. 359–406.

Proffett, J. M., Jr., 1977, Cenozoic geology of the Yerington District, Nevada, and implications for the nature and origin of Basin and Range faulting: Geol. Soc. America Bull., v. 88, p. 247–266.

Prostka, H. J., Eaton, G. P., and Oriel, S. S., 1976, Cordilleran thermotectonic anomaly: II. In-

teraction of an intraplate chemical plume mass and related mantle diapir: Geol. Soc. America Abs. with Programs, v. 8, p. 1054–1055.

Reynolds, M. W., 1969, Stratigraphy and structural geology of the Titus and Titanothere Canyons area, Death Valley, California [Ph.D. dissert.]: Berkeley, Univ. California, 310 p.

Robinson, E. S., 1970, Relations between geological structure and aeromagnetic anomalies in central Nevada: Geol. Soc. America Bull., v. 81, p. 2045–2060.

Robinson, G. D., 1960, Middle Tertiary unconformity in southwestern Montana: U.S. Geol. Survey Prof. Paper 400-B, p. B227–B228.

——1961, Origin and development of the Three Forks basin, Montana: Geol. Soc. America Bull., v. 72, p. 1003–1014.

Robinson, P. T., McKee, E. H., and Moiola, R. J., 1968, Cenozoic volcanism and sedimentation, Silver Peak region, western Nevada and adjacent California, in Coats, R. R., and others, eds., Studies in Volcanology (Williams volume): Geol. Soc. America Mem. 116, p. 577–612.

Sass, J. H., Lachenbruch, A. H., Munroe, R. J., Greene, G. W., and Moses, T. H., Jr., 1971, Heat flow in the Western United States: Jour. Geophys. Research, v. 76, p. 6376–6413.

Scholz, C. H., Barazangi, M., and Sbar, M. L., 1971, Late Cenozoic evolution of the Great Basin, western United States, as an ensialic interarc basin: Geol. Soc. America Bull., v. 82, p. 2979–2990.

Scott, R. B., Nesbitt, R. W., Dasch, E. J., and Armstrong, R. L., 1971, A strontium isotope evolution model for Cenozoic magma genesis, eastern Great Basin, U.S.A.: Bull. Volcanol., v. 35, p. 1–26.

Shaw, H. R., and Jackson, E. D., 1973, Linear island chains in the Pacific: Result of thermal plumes or gravitational anchors?: Jour. Geophys. Research, v. 78, p. 8634–8652.

Slemmons, D. B., 1967, Pliocene and Quaternary crustal movements of the Basin and Range province, U.S.A.: Jour. Geoscience, v. 10, p. 91–103.

Smith, R. B., 1978, Seismicity, crustal structure, and intraplate tectonics of the interior of the western Cordillera, in Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

Smith, R. B., and Sbar, M. L., 1974, Contemporary tectonics and seismicity of the western United States with emphasis on the Intermountain seismic belt: Geol. Soc. America Bull., v. 85, p. 1205–1218.

Snyder, W. S., Dickinson, W. R., and Silberman,

M. L., 1976, Tectonic implications of space-time patterns of Cenozoic magmatism in the Western United States: Earth and Planetary Sci. Letters, v. 32, p. 91–106.

Stewart, J. H., 1971, Basin and Range structure: A system of horsts and grabens produced by deep-seated extension: Geol. Soc. America Bull., v. 82, p. 1019–1044.

——1978, Basin-range structure in western North America—A review, in Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

Stewart, J. H., and Carlson, J. E., 1976, Cenozoic rocks of Nevada—Four maps and a brief description of distribution, lithology, age, and centers of volcanism: Nevada Bur. Mines and Geology Map Sheet 52.

——1978, Generalized maps showing distribution, lithology, and age of Cenozoic igneous rocks in the Western United States, in Smith, R. B., and Eaton, G. P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geol. Soc. America Mem. 152 (this volume).

Stewart, J. H., Walker, G. W., and Kleinhampl, F. J., 1975, Oregon-Nevada lineament: Geology, v. 3, p. 265–268.

Stewart, J. H., Moore, W., and Zietz, I., 1977, East-west patterns of Cenozoic igneous rocks, aeromagnetic anomalies, and mineral deposits, Nevada and Utah: Geol. Soc. America Bull., v. 88, p. 67–77.

Stock, Chester, and Bode, F. D., 1935, Occurrence of lower Oligocene mammal-bearing beds near Death Valley, California: Natl. Acad. Sci. Proc., v. 21, p. 571–579.

Stone, G. T., 1967, Petrology of upper Cenozoic basalts of the western Snake River Plain, Idaho [Ph.D. dissert.]: Boulder, Univ. Colorado.

Suppe, J., Powell, C., and Berry, R., 1975, Regional topography, seismicity, Quaternary volcanism and the present-day tectonics of the Western United States: Am. Jour. Sci., v. 275-A.

Swanson, D. A., Wright, T. L., and Helz, R. T., 1975, Linear vent systems and estimated rates of magma production and eruption for the Yakima Basalt on the Columbia Plateau: Am. Jour. Sci., v. 275, p. 877–905.

Taubeneck, W. H., 1970, Dikes of Columbia River basalt in northeastern Oregon, western Idaho, and southeastern Washington, in Gilmour, E. H., and Stradling, Dale, eds., Proc. 2nd Columbia River Basalt symposium: Cheney, Eastern Washington State College Press, p. 73–95.

Thompson, G. A., and Burke, D. B., 1974, Regional geophysics of the Basin and Range province: Ann. Rev. Earth and Planetary Sci., v. 2,

p. 213–237.

Thornbury, W. D., 1965, Regional geomorphology of the United States: New York, Wiley, 609 p.

Tilley, C. E., and Thompson, R. N., 1970, Melting and crystallization relations of the Snake River basalts of southern Idaho, U.S.A.: Earth and Planetary Sci. Letters, v. 8, p. 79–92.

Tschanz, C. M., 1960, Regional significance of some lacustrine limestones in Lincoln County, Nevada, recently dated as Miocene: U.S. Geol. Survey Prof. Paper 400-B, p. 293–295.

Walker, G. W., 1973, Preliminary geologic and tectonic maps of Oregon east of the 121st meridian: U.S. Geol. Survey Misc. Field Studies Map MF-495.

——1974, Some implications of late Cenozoic volcanism to geothermal potential of south-central Oregon: Ore Bin, v. 36, p. 109–119.

Watkins, N. D., and Baksi, A. K., 1974, Magnetostratigraphy and oroclinal folding of the Columbia River, Steens, and Owyhee basalts in Oregon, Washington, and Idaho: Am. Jour. Sci., v. 274, p. 148–189.

Willden, R., 1964, Geology and mineral deposits of Humboldt County, Nevada: Nevada Bur. Mines Bull. 59, 154 p.

Winfrey, W. M., Jr., 1960, Stratigraphy, correlation and oil potential of the Sheep Pass Formation, east-central Nevada, in Guidebook to the geology of east-central Nevada: Salt Lake City, Utah, Intermtn. Assoc. Petroleum Geologists, p. 126–133.

Wright, L. A., 1976, Late Cenozoic fault patterns and stress fields in the Great Basin and westward displacement of the Sierra Nevada block: Geology, v. 4, p. 489–494.

Wright, T. L., Grolier, M. J., and Swanson, D. A., 1973, Chemical variation related to the stratigraphy of the Columbia River Basalt: Geol. Soc. America Bull., v. 84, p.371–386.