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Cenozoic transverse zones and igneous belts in the Great Basin, western United States: Their tectonic and economic implications

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ABSTRACT

Transverse zones are common, albeit controversial, features in the Great Basin, United States, that appear to allow different amounts and types of extension on either side. Most strike east-northeast to east-southeast, at high angles to, and partly contemporaneous with, most other extensional structures and topography. Transverse zones commonly contain faults, joints, and folds of the same strike. Intrusive masses and transverse zones in the Great Basin are closely associated in place and time, suggesting that (1) magma injects along transverse zones, (2) transverse zones form above intrusive masses, and (3) transverse zones separate domains where extension was primarily accomplished by magma emplacement from domains where extension was accomplished by faulting. Alignments of intrusive masses form igneous belts that generally parallel transverse zones. However, most igneous belts also contain associated northerly striking extensional faults that formed at the same time as the belt. Igneous belts are younger from north (Eocene) to south (Miocene) through the Great Basin, and within some igneous belts, individual intrusions young toward the west in the western Great Basin and toward the east in the eastern Great Basin. Most transverse zones and igneous belts are interpreted to form parallel to the extension direction. Transverse zones and igneous belts play a major role in the spreading of brittle crust and act together with the brittle-ductile transition in the spreading process.

Transverse zones and igneous belts have economic importance because they commonly influence ground-water flow and are potential conduits for ground water and hydrocarbons. Hot springs and hydrothermally altered rock may be concentrated along transverse zones because of the common association of long-lived faults, which open pathways for ground water, and synchronous magma bodies. Hydrothermal solutions driven by magmatic heat commonly transport metals and localize ore deposits in transverse zones and igneous belts, a major reason for the great abundance of mineral deposits in the Great Basin. Another significant feature of transverse zones is that some coincide with segment boundaries in active faults and thus may influence the size and location of earthquakes.

INTRODUCTION

The Great Basin (Fig. 1), United States, is a tectonically active region of internal drainage characterized by north- to north-northeast–trending basins and ranges (Fig. 2) defined by late Cenozoic (middle Miocene to Quaternary) basin-range faults. These basins and ranges record major extension, but this is only the younger of two episodes of extensional tectonism and associated magmatism that began in the Eocene in the northern part of the Great Basin and the early Miocene in the southern

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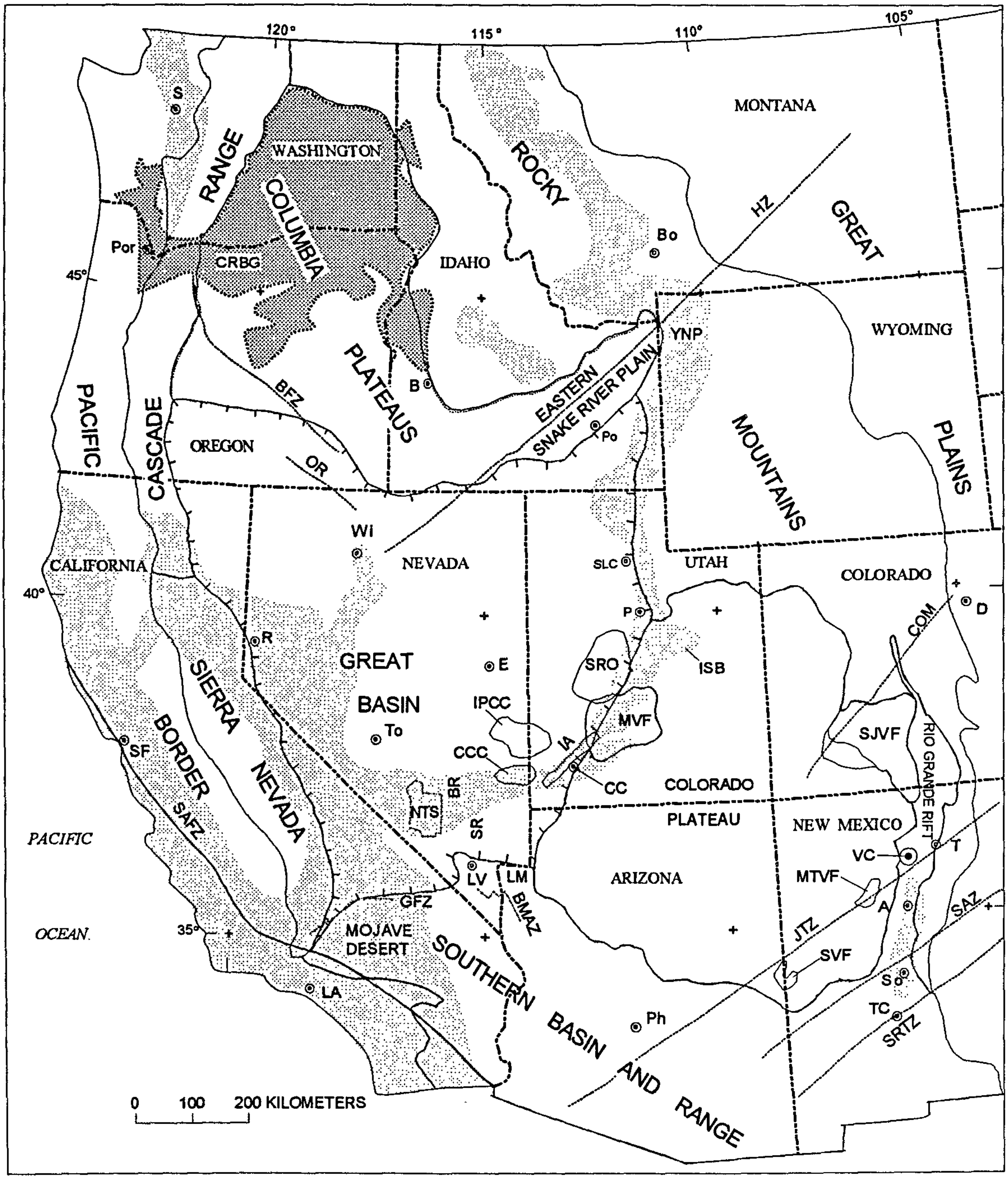


Figure 1. Map of the western United States showing physiographic provinces (from Eaton, 1980; Kuntz et al., 1992; Chapin et al., 1978). The Great Basin is hachured. Seismicity (stippled pattern) is generalized from Smith (1978, Fig. 6-6). A, Albuquerque; B, Boise; Bo, Bozeman; BR, Belted Range; CC, Cedar City; D, Denver; E, Ely; LA, Los Angeles; LM, Lake Mead; LV, Las Vegas; NTS, Nevada Test Site; P, Payson; Ph, Phoenix; Po, Pocatello; Por, Portland; R, Reno; S, Seattle; SLC, Salt Lake City; SF, San Francisco; So, Socorro; SR, Sheep Range; T, Taos; TC, Truth or Consequences; To, Tonopah; Wi, Winnemucca; YNP, Yellowstone National Park. Geologic features: BFZ, Brothers fault zone; BMAZ, Black Mountains accommodation zone; CCC, Caliente caldera complex; COM, Colorado mineral belt (Cunningham et al., 1994); CRBG, Columbia River Basalt Group (check pattern); GFZ, Garlock fault zone; HZ, Humboldt zone (Eaton et al., 1975; Rowan and Wetlaufer, 1981); IA, Iron axis; IPCC, Indian Peak caldera complex; ISB, Intermountain seismic belt; JTZ, Jemez transverse zone (Lipman, 1980, Fig. 14.5); MVF, Marysvale volcanic field; MTVF, Mount Taylor volcanic field; OR, Orevada rift; SAFZ, San Andreas fault zone; SAZ, Socorro accommodation zone; SJVF, San Juan volcanic field; SRO, Sevier River oval; SRTZ, Santa Rita transverse zone; SVF, Springerville volcanic field; VC, Valles caldera.

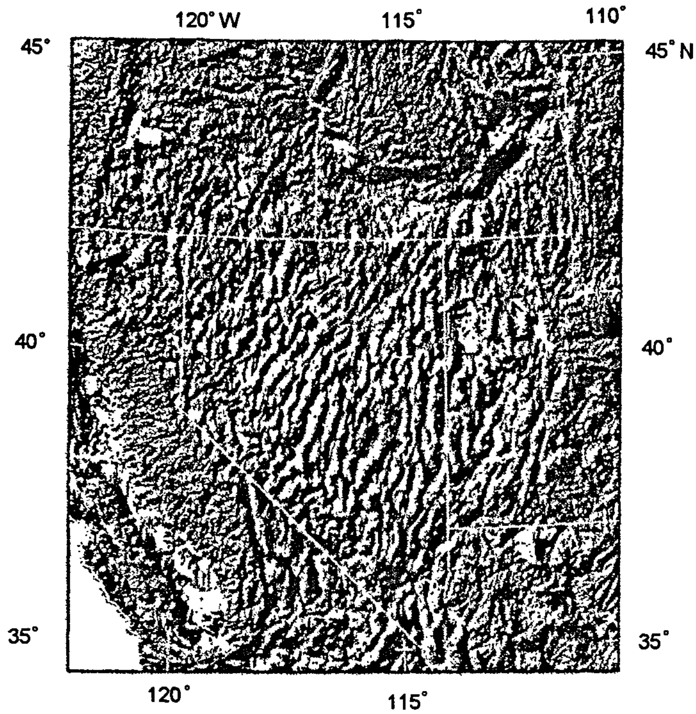


Figure 2. Topography of the Great Basin and adjacent areas, showing the basins and ranges of Nevada and adjacent states and the lowland of the Snake River Plain (from Thelin and Pike, 1991).

part. The older episode, of middle Cenozoic age (Eocene to middle Miocene), had comparable if not greater amounts of extension. Overall extension of the two episodes may have totaled 100% (Hamilton, 1988a). Young strata and basin-range faults commonly obscure middle Cenozoic extensional structures, but detailed geologic mapping reveals that these structures are abundant yet heterogeneous in type, distribution, orientation, and age from place to place. Middle and late Cenozoic extensional structures include east-west faults, transverse zones, and igneous belts, but evidence for the origin of these east-west features is controversial. The extension direction in the Great Basin under which all the extensional structures formed was generally east-northeast during the Eocene to middle Miocene and changed progressively with time to east-southeast (Zoback et al., 1981) and east during middle Miocene to Quaternary time.

The regional (long wavelength) gravity field, topography, and some other parameters of the Great Basin, but not its exposed geology, are marked by general bilateral symmetry about a north-south axis (Figs. 3 and 4; Eaton et al., 1978; Eaton, 1982). This axis partly coincides with the aeromagnetic "quiet zone" (Stewart et al., 1977) characterized by rocks of low magnetic intensity and magnetic relief (Figs. 3 and 5). The genetic relationship of the axis to the quiet zone is unclear (Blakely, 1988). Geologic features west of the axis that may reflect processes that formed the bilateral symmetry include several north-northwest–striking rifts (Stewart et al., 1975; Mabey et al., 1978; Zoback and Thompson, 1978; McKee and Noble, 1986; Blakely, 1988; Blakely and Jachens, 1991). The most prominent of these is the northern Nevada rift (Fig. 3), which is filled with 17–14 Ma basalt and rhyolite dikes and lava flows. The strike of this rift, like normal faults and dikes in tension fractures, defines the older (middle Cenozoic) east-northeast direction of extension (Zoback and Thompson, 1978). McKee and Noble (1986), Blakely (1988), Blakely and Jachens (1991), Anderson et al. (1994), and Zoback et al. (1994) described the rift and projected it to southern Nevada. The northern Nevada rift is cogenetic with the McDermitt caldera at its northern end and with the Columbia River Basalt Group (Figs. 1 and 3; Zoback et al., 1994).

Cenozoic faults are generally younger eastward and westward from the axis toward the margins of the Great Basin, where most current seismicity is concentrated (Smith, 1978; Eaton, 1982; Rogers et al., 1991). The seismicity on the eastern and southern margin of the Great Basin forms the arc-like Intermountain seismic belt (Fig. 1; Smith, 1978), whereas most seismicity on the western margin is in the Walker Lane belt and Owens Valley area (Fig. 3). In addition, active faults occupy the northerly trending Central Nevada seismic belt (Fig. 3; Wallace, 1984).

This chapter describes transverse zones and associated igneous belts in the Great Basin that result from synchronous magmatism and extension and that were identified primarily by geologic mapping (e.g., Steven et al., 1990; see citations in Rowley et al., 1998). Mapping of the Marysvale volcanic field, Utah, which is mostly in the High Plateaus transition zone between the Basin and Range province and the stable part of the Colorado Plateau (Figs. 1 and 6), revealed two sequential episodes (one mostly in the middle Cenozoic and the other in the late Cenozoic) of magmatism and one episode of extensional faulting (late Cenozoic). Mapping in the eastern Great Basin, including the Iron Springs mining district and its associated northeast-trending belt of intrusions called the Iron axis (Figs. 1 and 6), disclosed much the same picture as in the Marysvale field, except that locally we found suggestions of an older episode (middle Cenozoic) of extensional faulting. Later mapping, centered on the Caliente caldera complex of Nevada and Utah (Figs. 1 and 6), demonstrated that here the middle Cenozoic extensional episode was more significant than that in the late Cenozoic. Our work in the Marysvale and Caliente areas also identified east-striking "lineaments" and igneous belts. For preliminary interpretations of these features, see Rowley et al. (1998).

The transverse zones and igneous belts were projected outside the areas of detailed mapping on the basis of topography, geology, mining districts (Hilpert and Roberts, 1964; Roberts, 1964; Bonham, 1991; Tingley, 1992), geothermal areas (Blackett, 1994; Garside, 1994), and aeromagnetic data (Fig. 5; Zietz et al., 1976, 1978; Stewart et al., 1977; Hildenbrand and Kucks, 1988a; V. Bankey, 1995, written commun.). Aeromagnetic maps in which the magnetic field has been reduced to the north magnetic pole (Hildenbrand and Kucks, 1988b) are most valuable, because anomalies are shifted to correspond more accurately to geologic sources. Bouguer gravity maps (Cook et al., 1989; Saltus, 1988a; V. Bankey, 1992, written commun.) are also useful, although less so than aeromagnetic maps, because gravity data are more profoundly influenced by later basin-range faults. Where available, isostatic

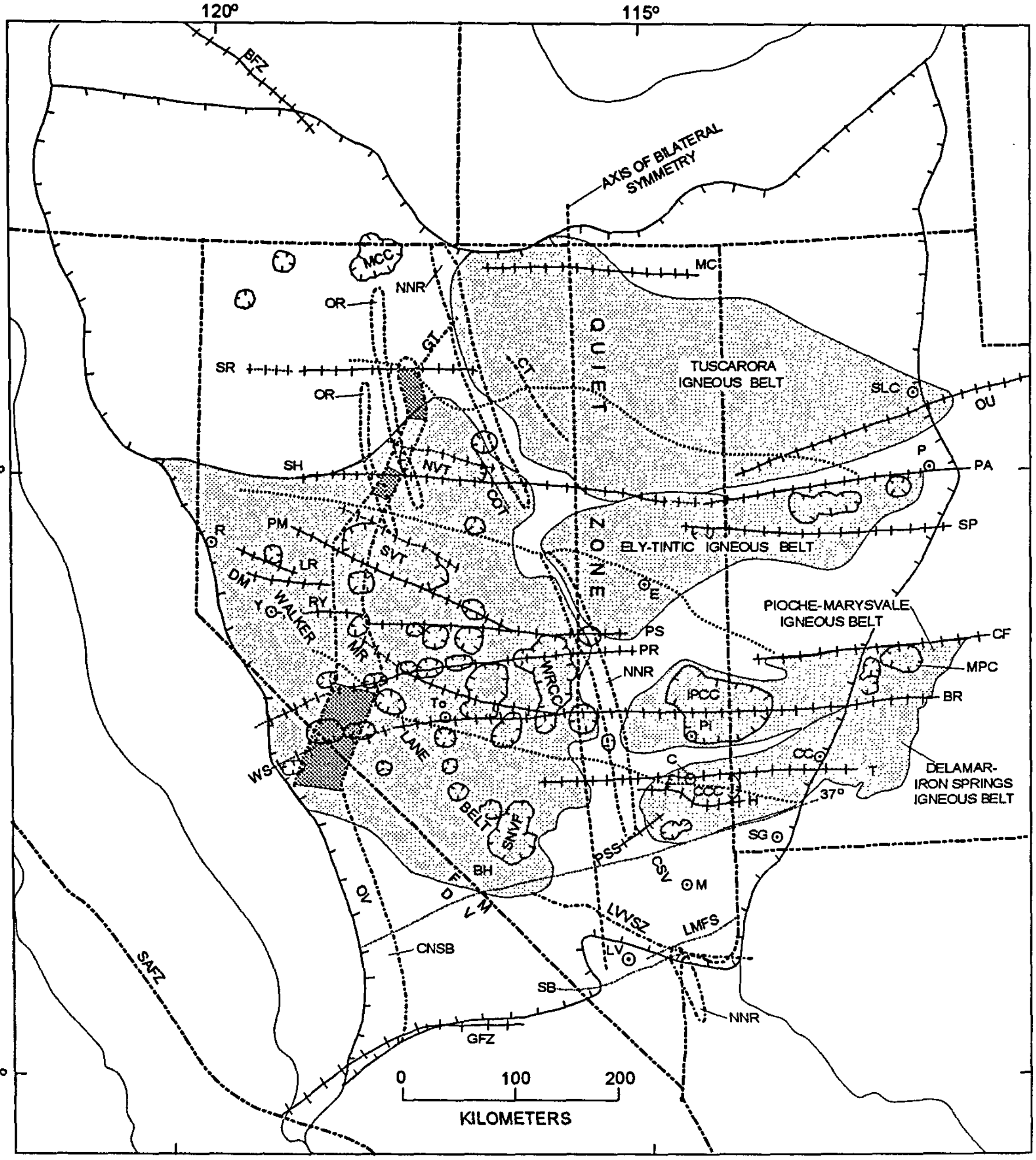


Figure 3. Map of the Great Basin (hachured) part of the Basin and Range province, showing transverse zones, igneous belts, and other features discussed in the text. BH, Bullfrog Hills; C, Caliente; CC, Cedar City; CSV, Coyote Spring Valley; DV, Death Valley; E, Ely; FM, Funeral Mountains; LV, Las Vegas; M, Moapa; OV, Owens Valley; P, Payson; Pi, Pioche; R, Reno; SG, St. George; SLC, Salt Lake City; To, Tonopah; Y, Yerington. Transverse zones are shown by crossed lines and dashed where approximately located: BR, Blue Ribbon; CF, Cove Fort; H, Helene; MC, Mountain City; MR, Monitor Range; OU, Oquirrh-Uinta; PA, Payson; PM, Pirouette Mountain; PR, Pancake Range; PS, Prichards Station; RY, Rawhide-Yerington; SH, Sou Hills; SP, Sand Pass; SR, Sonoma Range; T, Timpahute; WS, Warm Springs. Other probable transverse zones shown with the same crossed line: BFZ, Brothers fault zone; DM, Desert Mountains structural and volcanic zone; GFZ, Garlock fault zone; LR, Lahontan Reservoir structural and volcanic zone; PSS, Pahranagat shear system. Igneous belts are stippled. Calderas in the Great Basin (after Best et al., 1989b) are shown by light hachures and include the following: CCC, Caliente caldera complex; IPCC, Indian Peak caldera complex; MCC, McDermitt caldera complex; MPC, Monroe Peak caldera; NVT, northern volcano-tectonic trough of Burke and McKee (1979); SNVF, southwest Nevada volcanic field; SVT, southern volcano-tectonic trough of Burke and McKee (1979); WRCC, Williams Ridge–Hot Creek Valley caldera complex. Three "transverse boundaries" of Stewart (1980, this volume) are shown by dotted lines. Heavy north-south dashed line is the Great Basin axis of general bilateral symmetry of Eaton et al. (1978, Fig. 3-8). Other geologic features are shown by dash-dot lines: CT, Carlin trend; COT, Cortez trend; CNSB, Central Nevada seismic belt (Rogers et al., 1991); GT, Getchell trend; LMFS, Lake Mead fault system; LVVSZ, Las Vegas Valley shear zone; NNR, three well-defined segments of the northern Nevada rift; OR, two other rifts west of the northern Nevada rift; SAFZ, San Andreas fault zone; SB, southern boundary of amagmatic corridor (Anderson and Barnhard, 1993); 37°, lat 37°N discontinuity. Within the Central Nevada seismic belt, crosshatched areas show the following seismic gaps, from north to south: Sonoma Range, Stillwater, and White Mountain.

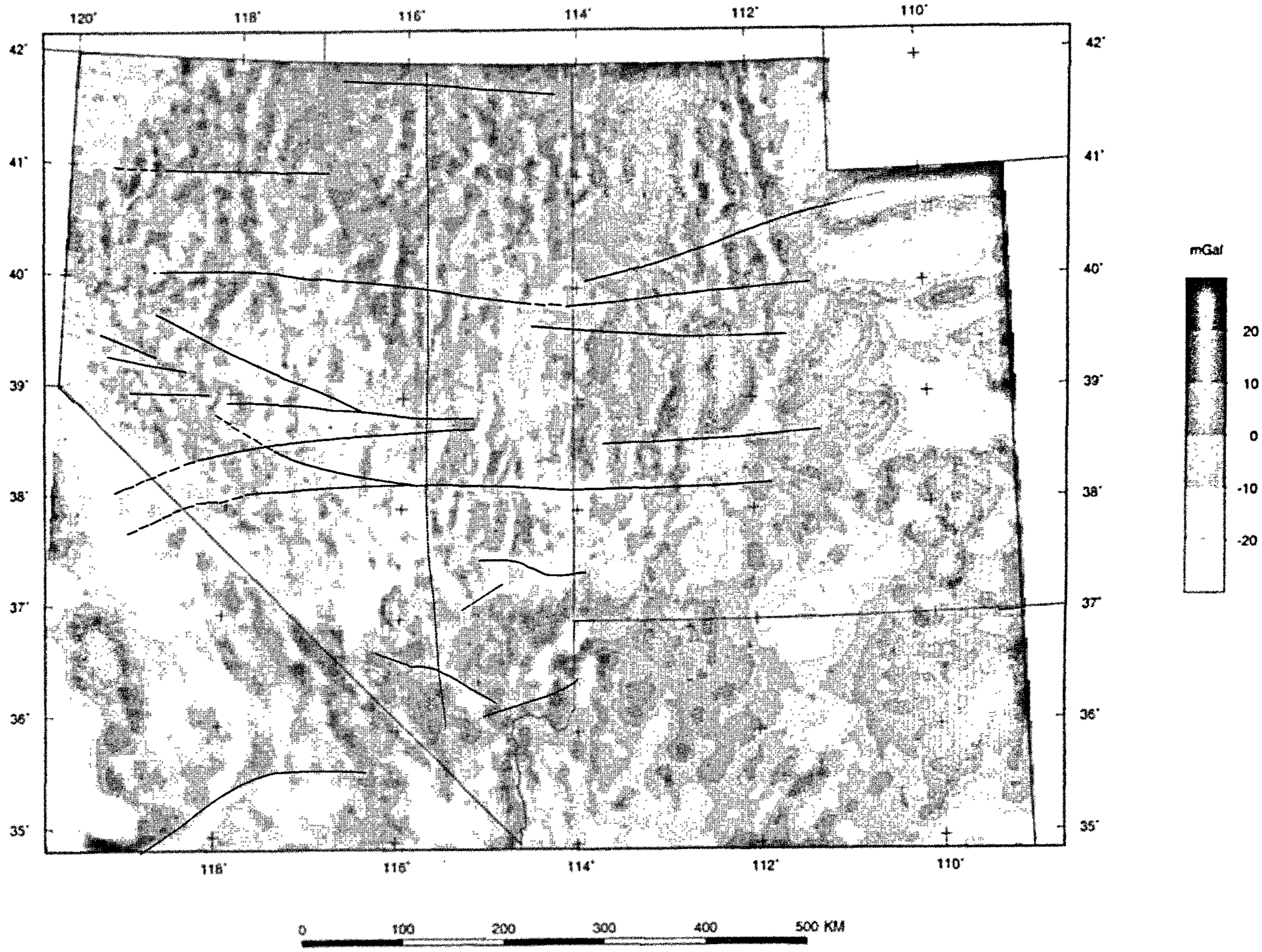


Figure 4. Isostatic residual gravity anomaly map of Nevada, Utah, and adjacent areas, on which are superimposed transverse zones from Figure 3. Contour interval 10 mGal. Data from Saltus and Jachens (1995); grids available from National Geophysical Data Center, Boulder, Colorado. Nevada data modified from Saltus (1988b). Dashed north-south line is the Great Basin axis of general bilateral symmetry of Eaton et al. (1978).

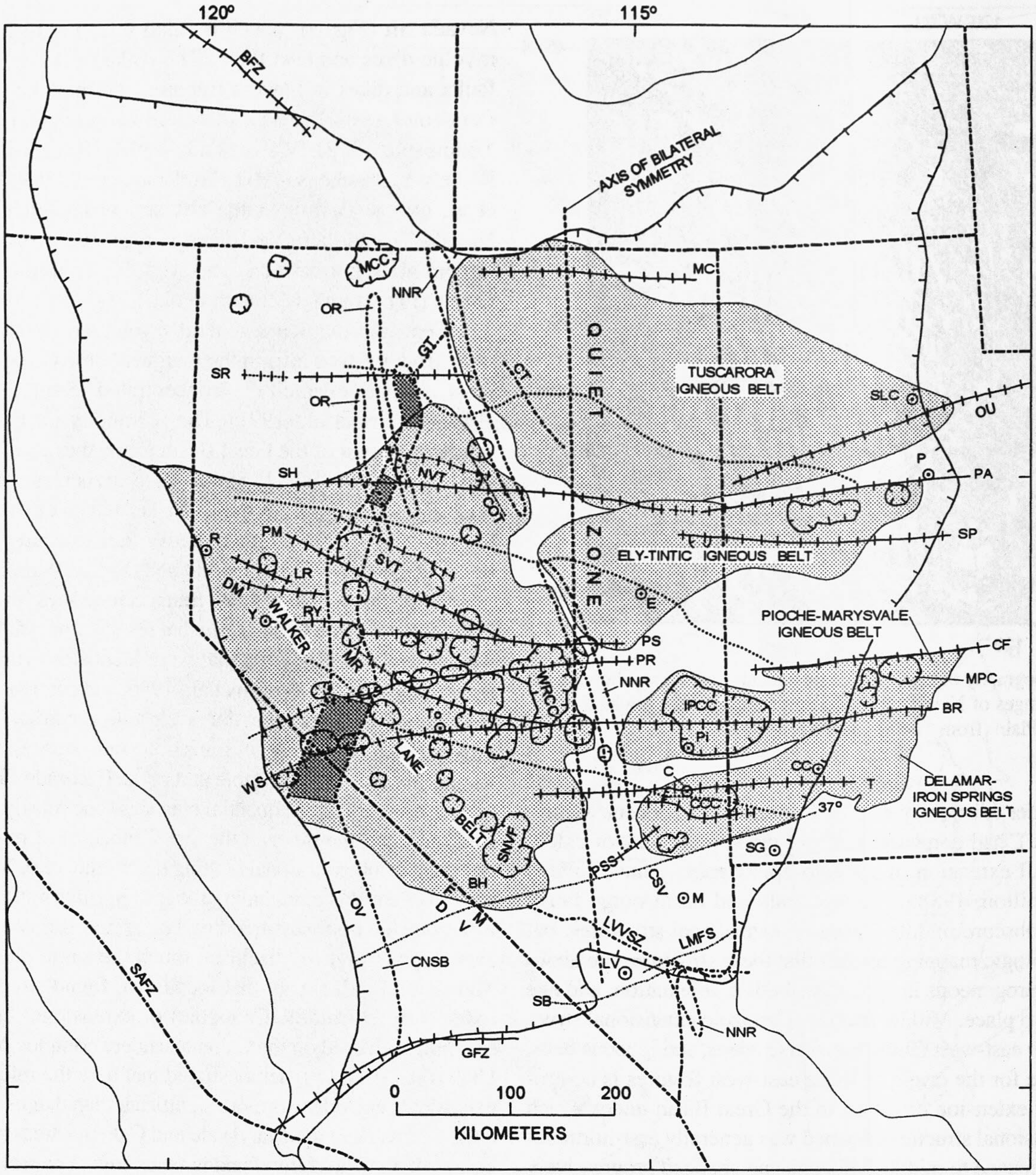
residual gravity maps (Fig. 4; Saltus, 1988b; Saltus and Jachens, 1995; V. Bankey, 1994, written commun.) better represent shallow density distributions (Blakely and Jachens, 1991). A gravity map of pre-Tertiary basement rocks (Jachens and Moring, 1990; Saltus and Jachens, 1995), which removes the effects of basin-range faulting, is even more valuable (Blakely and Jachens, 1991).

Terminology

I use Great Basin as the northern subprovince of the Basin and Range physiographic province (Fig. 1). "Basin-range" is applied to the late Cenozoic high-angle faults that produced the present topography (Fig. 2) or to the responsible deformational episode, following Gilbert (1928) and Mackin (1960). "Spreading" and "extension" refer to the means by which brittle and ductile crust extends by faulting, fault-block rotation, and magmatism.

Following Slemmons (1967) and Stewart (1980), I use the nongenetic term "transverse zone" for features commonly oriented at a high angle, or transverse, to coeval extensional structures and topography. Transverse zones have been referred to by other names, notably "lineaments" (e.g., Ekren et al., 1976) and "transverse structures" (Duebendorfer and Black, 1992; Rowley et al., 1998). Transverse zones commonly consist of belts, some as wide as 25 km, of high-angle oblique-slip faults, but other zones are expressed by normal-slip and strike-slip faults, folds, joints, and zones of twisting. Most if not all transverse zones are boundaries that separate domains of different style, amount, or rate of strain. "Igneous belts" are linear belts of intrusions, volcanic centers, and associated extensional faults that are commonly bounded by transverse zones.

"Transfer faults" (or transfer zones) and "accommodation zones" are in many ways synonymous with transverse zones.



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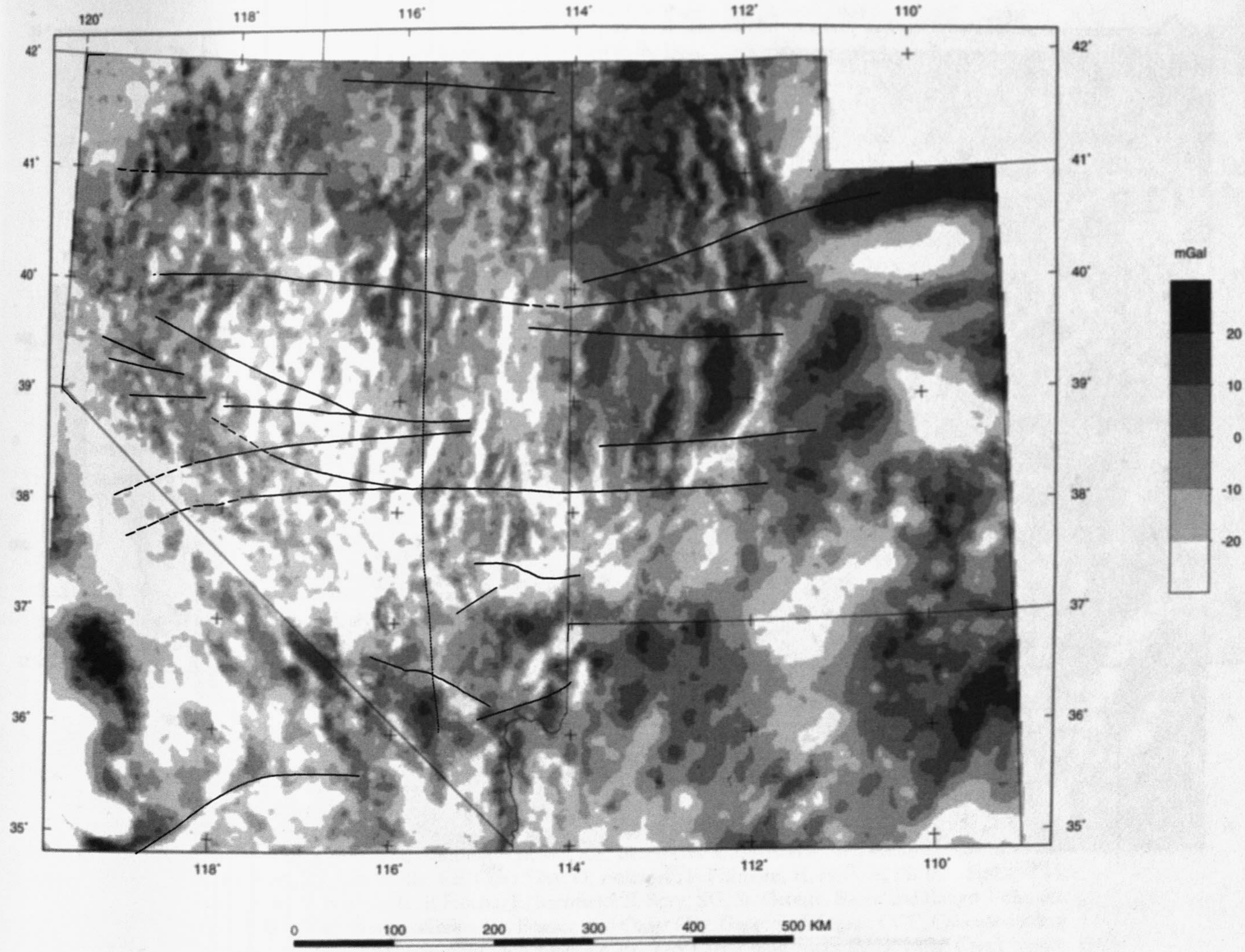

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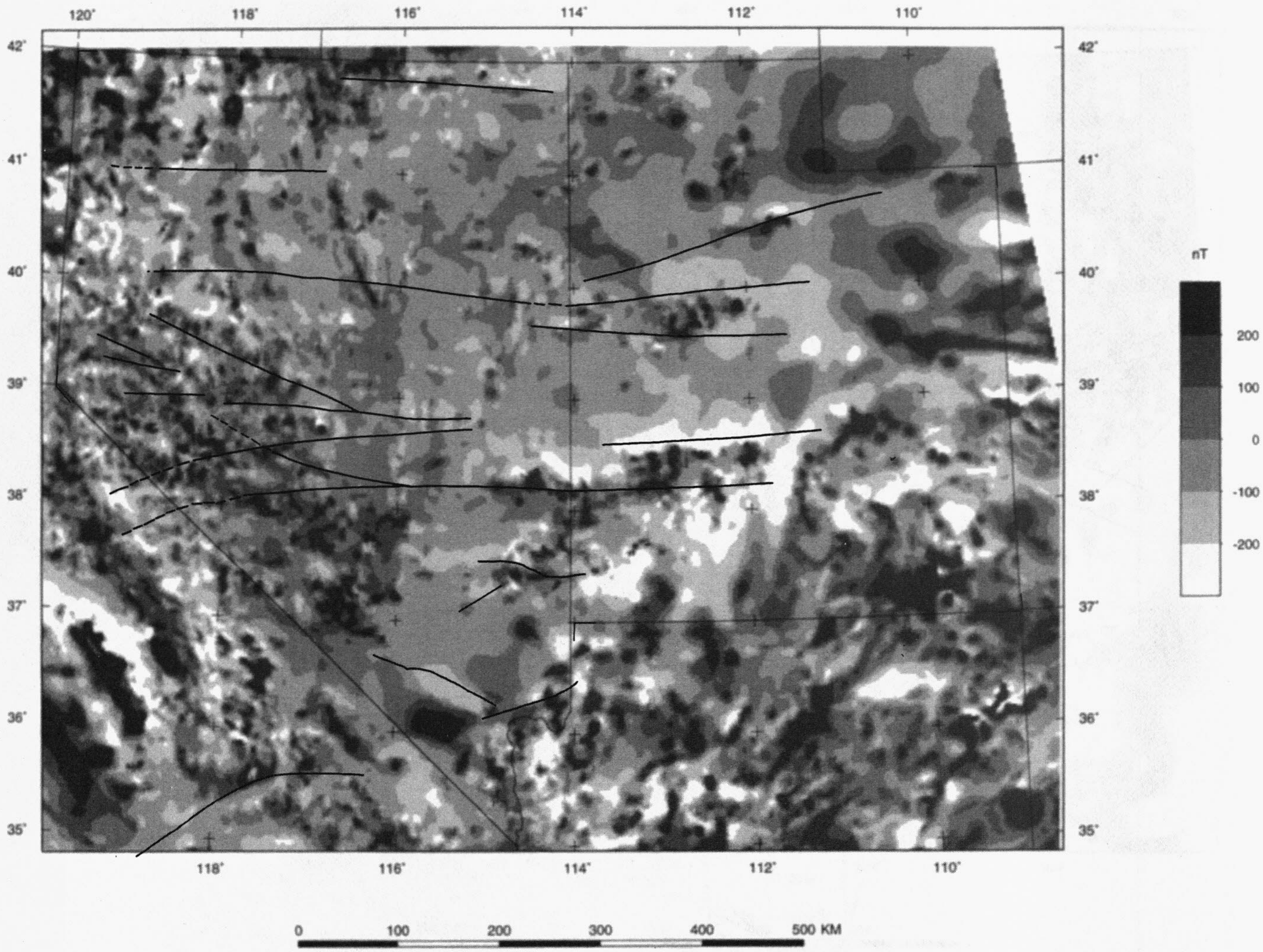

Figure 5. Aeromagnetic anomaly map of Nevada, Utah, and adjacent areas, on which are superimposed transverse zones from Figure 3. Contour interval 100 nT. Data from Magnetic Map of North America (Committee for the Magnetic Map of North America, 1987), also available on CD-ROM (Hittleman et al., 1990). Nevada data modified from Hildenbrand and Kucks (1988a).

Transfer faults are transversely oriented strike-slip or oblique-slip faults that accommodate changes in the style or magnitude of extension across them (Gibbs, 1984). As such, they are a subspecies of transverse zones in which all accommodation is by faults. Accommodation zones may be parallel, oblique, or orthogonal to the extension direction (e.g., Faulds and Varga, this volume). Accommodation zone, first used as a general term (e.g., "zone of accommodation" by Davis and Burchfiel, 1973) in much the same way as I use transverse zone, was restricted in definition by Bosworth (1985, 1986), Rosendahl (1987), and Faulds et al. (1990) to a tilt-block domain boundary. For example, the Black Mountains accommodation zone (Fig. 1) of Arizona and Nevada accommodates a reversal in structural polarity (i.e., reversal in dominant tilt direction of fault blocks and dip direction of normal faults) through twisting or torsional strain (Faulds et al., 1990, 1992) and may correspond to a regional rupture barrier (Faulds, 1994; Faulds and Varga, this volume). Accommodation zones are commonly associated with igneous centers (Faulds et al., 1994; Faulds and Varga, this volume). As I use it, transverse zone is a more general term than accommodation or transfer zones, because it includes examples of both.

Some transverse zones are similar to transform faults (Freund, 1974). Both develop when amounts of strain differ on opposite sides of the structure (Davis and Burchfiel, 1973; Duebendorfer and Black, 1992). The actual sense of strike-slip motion in both structures may be opposite to apparent offset of extended terranes. Transform faults commonly are subparallel to the world's lines of latitude (McKenzie and Parker, 1967), as are most large transverse zones. Intrusions may be emplaced along both types of structures. In the case of transforms, these have been called "leaky" and are considered to be due to changes in spreading directions that create transtension (Menard and Atwater, 1969).

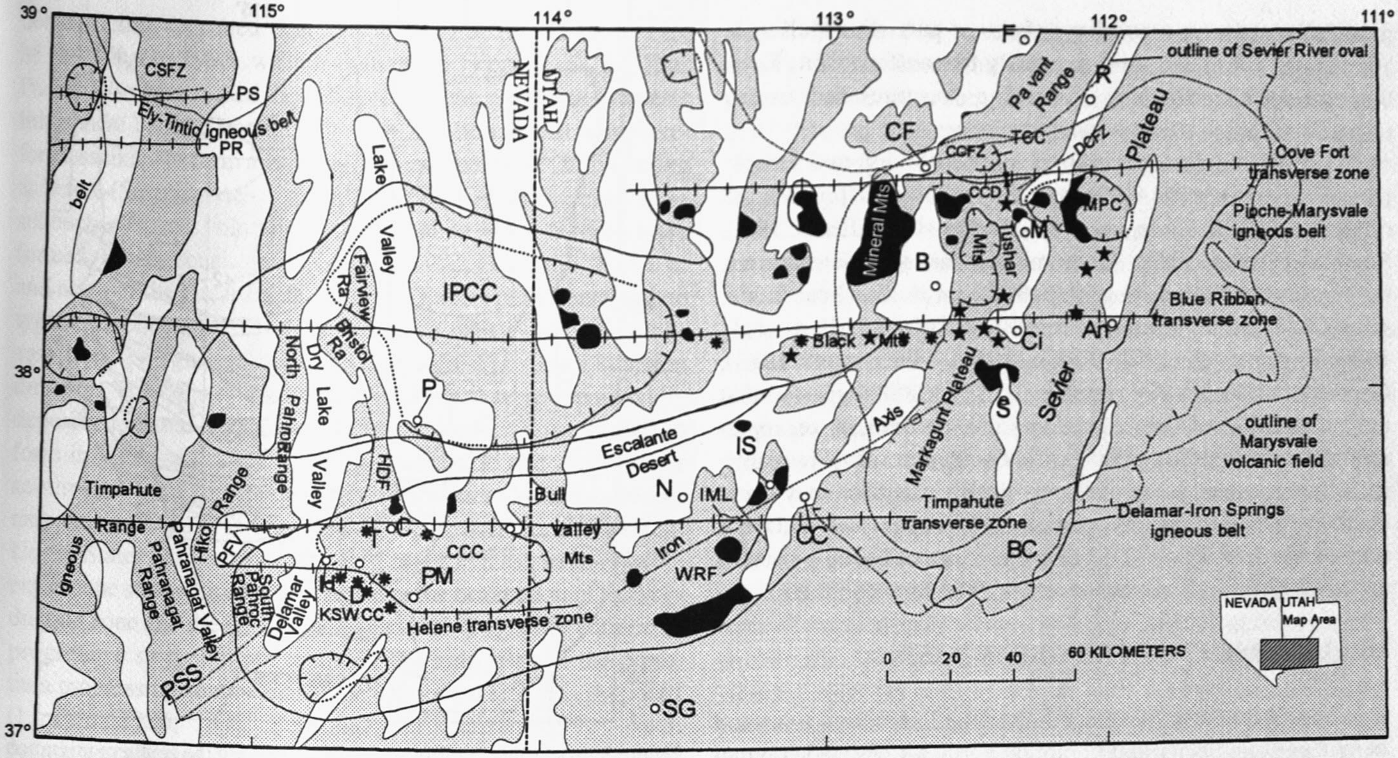

Figure 6. Map showing the geographic setting of the Pioche-Marysvale and Delamar–Iron Springs igneous belts, transverse zones (crossed lines), calderas, volcanic fields, and other geologic features in the southeastern Great Basin. Outline of High Plateaus transition zone between the Basin and Range and the Colorado Plateau is shown by widely spaced hachures on dashed fine lines. Calderas are shown by closely spaced hachures on fine lines; stratovolcano centers shown by stars; rhyolite domes shown by asterisks. Ranges are shaded; valleys underlain by alluvium are undecorated; Tertiary intrusions are shown in dark gray. An, Antimony; B, Beaver; BC, Bryce Canyon National Park; C, Caliente; CC, Cedar City; Ci, Circleville; CF, Cove Fort; D, Delamar; F, Fillmore; H, Helene; IS, Iron Springs; M, Marysvale; N, Newcastle; P, Pioche; R, Richfield; S, Spry; SG, St. George. Basin and Range–Colorado Plateau boundary is west of Fillmore, Beaver, and Cedar City. Geologic features: CCC, Caliente caldera complex; CSFZ, Currant Summit fault zone; DCFZ, Dry Creek fault zone; CCD, Clear Creek downwarp; CCF, Cove Creek fault; HDF, Highland detachment fault; IML, Iron Mountain laccolith; IPCC, Indian Peak caldera complex; KSWCC, Kane Spring Wash caldera complex; MPC, Monroe Peak caldera; PM, Pennsylvania mining district; PR, Pancake Range transverse zone; PS, Prichards Station transverse zone; PSS, Pahranagat shear system; PVF, Pahroc Valley fault; T, Taylor mine; TCC, Three Creeks caldera; WRF, Woolsey Ranch fault.

Alternatively, underlying intrusions may control the location of some transverse zones (Faulds et al., 1994). Both transverse and transform zones are long-lived features that may remain active for tens of millions of years, whereas few faults are active for more than a few million years (Sibson, 1989). Transform and transverse zones differ from each other in that transforms are generally found in oceanic lithosphere, although major exceptions such as the San Andreas fault zone (Fig. 1) are well known. Thus, the term "continental transform fault" has been used informally for transverse zones by some workers. Transforms have large amounts of strike-slip displacement, in places hundreds of kilometers. Although some transverse zones in the Great Basin might have ten or more kilometers of strike-slip displacement (e.g., Wernicke et al., 1988), most segments typically have little or no detectable strike-slip offset. I conclude that slip on most transverse zones is oblique or normal. Perhaps the most significant difference is that transforms separate lithospheric plates and are parallel to the relative velocity vector between two plates (McKenzie and Parker, 1967), whereas transverse zones occur within lithospheric plates. Thus transforms are larger features. The San Andreas transform (McKenzie and Parker, 1967; Atwater, 1970) has been imaged by COCORP seismic reflection data to a depth of at least 20 km (Cheadle et al., 1986) and may extend to the base of the crust (Zandt, 1981). The Garlock fault zone (Fig. 1), probably a transverse zone, has been imaged by COCORP data to a depth of at least 7 but less than 18 km (Serpa, 1990). Furthermore, transforms terminate at extensional structures (e.g., spreading centers), shortening structures (e.g., fold belts), or another transform, so that the strain is accounted for at both ends (Freund, 1974); transverse zones do not do this. Transform faults conserve crust along planes of pure slip (McKenzie and Parker, 1967) and serve primarily to transfer strain. Thus, they are not extensional or shortening structures and, in this respect, are similar to some transverse zones.

I use "brittle-ductile transition zone" for an inferred subhorizontal zone that marks the base of the seismogenic (upper) part of the crust (R. E. Anderson, 1971; Tocher, 1975; Eaton, 1980; Smith and Bruhn, 1984). Along mid-oceanic spreading centers, the brittle-ductile transition is shallow, but in the Great Basin, seismicity indicates that it now is at 10–15 km depth (Benz et al., 1990; Rogers et al., 1991). Eaton (1980), Miller et al. (1983), Gans and Miller (1985), and Gans et al. (1985) suggested that major detachment faults and metamorphic core complexes represent subhorizontal flow at the brittle-ductile transition. In reality, many zones analogous to the brittle-ductile transition may occur at other strong competency contrasts in the upper crust, and these may be the sites of small detachment faults or bedding-parallel zones of shear (R. E. Anderson, 1994, 1995, oral commun.).

DEVELOPMENT OF THE GREAT BASIN

The Great Basin developed after the Late Cretaceous and early Cenozoic Sevier and Laramide northeast-directed contractional events. The Laramide event probably resulted from rapid subduction beneath western North America of the shallowly east- to northeast-dipping Farallon-Vancouver plates of oceanic lithosphere, which extended as far east as Colorado (Severinghaus and Atwater, 1990). After culmination of the Laramide event after about 50 Ma (Severinghaus and Atwater, 1990), erosion of resulting orogenic highlands led to lacustrine and alluvial deposition in adjacent basins and generation of a widespread surface of low relief. Extension that formed the Great Basin began north of the current Great Basin in the Eocene, about 10 m.y. after the end of the Laramide event, and swept southward so that in the southern Great Basin it began in the Miocene, about 40 m.y. after the end of the Laramide event (Fig. 7; Christiansen and Yeats, 1992).

The Great Basin is best distinguished by its extensional faults, but the volume of igneous rocks is no less remarkable and is also closely tied to its origin. Spreading of some areas of the Great Basin was partly accommodated by passive shallow intrusions as well as the more commonly considered mechanisms of faulting, block rotations, and pervasive fracturing (Rowley et al., 1998). For example, the emplacement of basaltic dikes at spreading centers can accommodate extension to the exclusion of normal faulting (Lachenbruch and Sass, 1978). The emplacement of basaltic dikes can increase σ_3 so that normal faults will not form (Parsons and Thompson, 1991). Faults that die out along strike into eruptive centers of the same age may be explained by a similar mechanism (e.g., Pierce and Morgan, 1992; Faulds et al., 1994; Sawyer et al., 1994). Our mapping, for example, shows that extensional faults of middle Cenozoic age are absent in the Marysvale field (also noted by Gans et al., 1989, p. 48) and the Iron axis, in striking contrast to many middle Cenozoic faults in the Caliente area. This may indicate that most of the east-west

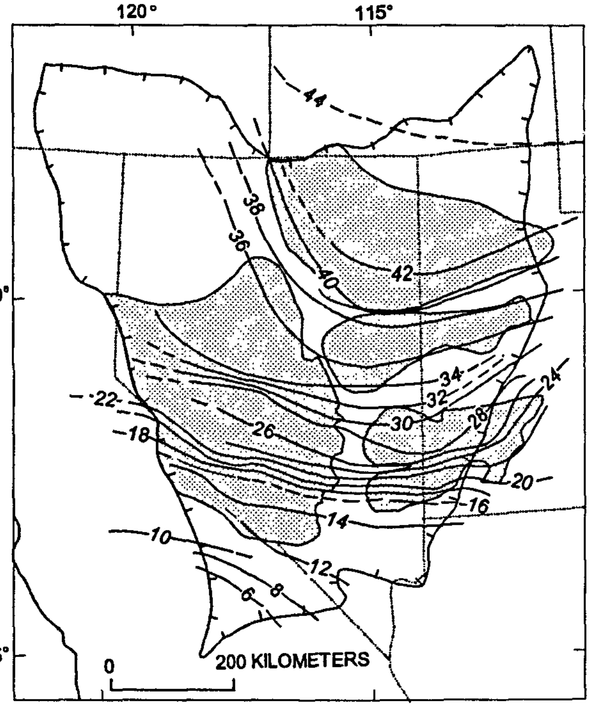

Figure 7. Isochrons, in millions of years, showing the southern migration of calc-alkaline magmatism and inferred coeval extension, and after about 18 Ma, bimodal (rhyolite) magmatism and inferred coeval extension (from Stewart and Carlson, 1976; Stewart et al., 1977; Luedke and Smith, 1981; Best et al., 1989b; Christiansen and Yeats, 1992; Luedke, 1993, 1994; Brooks et al., 1995; L. W. Snee, 1994, written commun.). Each isochron shows the southern limit of eruptive centers of the age shown; centers that are younger than this age may occur north of the isochron.

least-principal stress in the crust at Marysvale and in the Iron axis was taken up by passive emplacement of middle Cenozoic shallow intrusions of all compositions rather than by faults (e.g., Faulds et al., 1994; Rowley et al., 1998). I apply this field-based conclusion, however, only to highly extended areas and some transtensional fault zones in the Great Basin.

Magmatism and synchronous extensional faults occur together in only some areas of the Great Basin (Gans et al., 1989; Best and Christiansen, 1991; Faulds et al., 1995). Gans et al. (1989) developed a theory that voluminous magmatism, triggered by mantle-derived magma bodies, thermally weakened the crust and resulted in catastrophic failure of the upper crust by extensional faults. Although I agree with Gans et al. (1989) that faulting and magmatism commonly occur together in the Great Basin, I conclude that faulting is not due to magmatism or vice versa; both are products of deviatoric stresses and thus they may or may not occur in the same place. These deviatoric stresses were in turn derived from either subduction (in middle Cenozoic) or transform motion (in late Cenozoic).

During the middle Cenozoic, east-northeast–directed subduction continued beneath western North America, but at a reduced rate from that of the Laramide event (Severinghaus and Atwater, 1990). The flat subducting slab shortened and perhaps detached and steepened (Thompson and Zoback, 1979, p. 175; M. L. Zoback, 1995, written commun.) because, as the East Pacific Rise approached the subduction zone, subducted oceanic lithosphere became younger, thinner, and hotter, and was therefore less likely to form a coherent slab extending as far to the east as before (Severinghaus and Atwater, 1990). As above most other subduction zones, voluminous calc-alkaline igneous rocks were formed (Lipman et al., 1972; Lipman, 1992; Hamilton, 1995). In and north of the Great Basin, however, the eruptive centers were typically concentrated in linear, east-northeast– to east-southeast–trending igneous belts formed parallel to the extension direction. The belts of igneous centers, which were separated by depositional areas containing outflow volcanic facies, started to form in Washington and Idaho in the Eocene and are younger southward (Stewart et al., 1977). This pattern of migration of magmatism (Fig. 7) is one of several proposed for the western United States (Anderson, 1989; Best et al., 1989b). In an effort to explain the strong discordance between the orientation of the subduction zone and the igneous belts and their migratory patterns, a progressive steepening or lateral foundering (first in the north, then progressing southward) of the subducted slab was proposed (Lipman, 1980, 1992; Best and Christiansen, 1991). Such foundering allowed a southward-moving wedge of asthenosphere (Hamilton, 1995) to move upward and to generate mafic magmas, apparently by supplying heat to dewater the slab and partially melt the slab and mantle lithosphere (Lipman, 1980; Best and Christiansen, 1991). Perhaps the subducted slab shortened, beginning in the Eocene, in a complicated pattern that included breaking and foundering of pieces in different directions, leading to eventual destruction of the slab by the middle Miocene.

During the late Cenozoic (mostly middle Miocene to Quaternary), the crust continued to extend, although the plate tectonic setting differed radically from that of the middle Cenozoic. Except under the Cascade Range, the subducted slab had given way to transform movement in a broad zone, extending from the San Andreas fault to the Walker Lane belt (Fig. 3; Atwater, 1970; Eaton, 1980; Severinghaus and Atwater, 1990). As a result, the Great Basin underwent oblique extension (Hamilton and Myers, 1966). A change to a bimodal petrologic regime (Christiansen and Lipman, 1972), in places beginning before 20 Ma but generally younger than 17 Ma, coincided with the start of the transition from convergent to transform motion. The first product of late Cenozoic extension appears to have been broad warping (Rowley et al., 1981; Liberty et al., 1994). Extension culminated in basin-range faulting generally after 10 Ma, but earlier in the southern Basin and Range.

TRANSVERSE ZONES

Transverse zones ("lineaments") have long been recognized in the Great Basin by mining and petroleum geologists (e.g., Butler et al., 1920; Fuller, 1964; Hilpert and Roberts, 1964; Roberts, 1964, 1966), because the zones locally coincide with intrusions and associated mining districts. However, few zones were mapped in detail and evidence for their existence consisted mostly of linear trends that connected mining districts, faults, and potential-field anomalies. Butler et al. (1920, p. 102) predicted that "detailed work in the Basin Range of Utah will doubtless show the east-west faults to be rather abundant." Hamilton and Myers (1966) and Davis and Burchfiel (1973) recognized the significance of transverse faults that accommodated changes in the amounts of extension. On the basis of the tilts of fault blocks and distribution of young faults, Slemmons (1967) suggested the presence of three west-northwest–striking transverse zones across Nevada. Stewart (1980) noted that these zones correspond with tilt-domain boundaries, refined their location, and traced them into adjacent states. Thenhaus and Barnhard (1989, this volume) adopted them and called them "extensional accommodation zones." Aeromagnetic maps of Utah and Nevada by Zietz et al. (1976, 1978) confirmed that these trends had their roots in intrusive rocks. Descriptions of transverse zones followed when Ekren et al. (1976) described five zones (the longest 320 km) on the basis of detailed and reconnaissance geologic mapping in southern Nevada. They noted that the combination of different ages of plutons, some at least as old as Late Cretaceous, and faults demonstrate that many transverse zones are long lived. Transverse zones commonly mark terminations or interruptions of magnetic and other geophysical anomalies (Zietz et al., 1976, 1978; Hildenbrand and Kucks, 1988a, 1988b; Saltus, 1988a, 1988b; Cook et al., 1989; V. Bankey, 1985, written commun.) and commonly offset or terminate basins and ranges. The transverse zones have also been noted in several regional studies (Rowley et al., 1978; Stewart et al., 1977; Eaton et al., 1978; Mabey et al., 1978; Eaton, 1979; Wernicke et al., 1982; Stewart, 1983; Duebendorfer and Black, 1992; Stoeser, 1993). Eaton et al. (1978) and Mabey et al. (1978) interpreted them as transform faults.

Schwartz and Coppersmith (1984) suggested that the north-striking Wasatch fault zone of Utah, which marks the Basin and Range–Colorado Plateau boundary for nearly 400 km (Fig. 1), and other major basin-range fault zones consist of fault segments that rupture during earthquakes and are separated along strike by discrete boundaries. Zoback (1983) first noted that many boundaries correlate with east-west "transverse zones" defined by detailed gravity data. Some boundaries, which do not rupture, may be long lived, perhaps as long as the 10 m.y. history of the fault zone (Wheeler and Krystinik, 1992), and others are transient. For historically active fault zones, the term "seismic gap" (R. E. Wallace, 1978) has been applied to along-strike aseismic portions of the fault. Wheeler (1989) proposed that, through time, persistent boundaries result in local transverse bedrock ridges, some of which correspond to the transverse zones of Zoback (1983) and the "salients" of Gilbert (1928). Schwartz and Coppersmith (1984), Wheeler (1989), Wheeler and Krystinik (1988, 1992), Bruhn et al. (1992), and Machette et al. (1992) described as many as 10 segments with interceding persistent and transient boundaries along the Wasatch fault zone. Thenhaus and Barnhard (1989) also suggested that some boundaries are due to regional transverse zones.

Marysvale volcanic field

The main eruptive centers of the Marysvale volcanic field are bounded on their north and south sides by two transverse zones (Figs. 3, 6, and 8). The southern zone, the Blue Ribbon–Warm Springs transverse zone, is the best documented and longest in the Great Basin and is therefore described first and used as a model for the others. Its western part is the 220-km-long Warm Springs transverse zone ("lineament") of Ekren et al. (1976), named for a large east-striking fault zone, tufa mound, and hot-spring cluster at Warm Springs (Fig. 8). The zone may be traced by reconnaissance and detailed mapping from the western Great Basin boundary in California, east across the Walker Lane belt into central Nevada. Stewart's (1985) east-striking Coaldale fault zone (Fig. 8), which has significant late Mesozoic right-lateral offset and later Tertiary offset, is in part the same feature as the Warm Springs transverse zone where it crosses the Walker Lane belt. Ekren et al. (1976) defined the Warm Springs zone on the basis of prominent interruptions of magnetic anomalies and topographic features, many warm springs (Garside, 1994), intrusions, volcanic centers, east-northeast to east-striking faults, tectonic and landslide breccia, and local terminations of volcanic units. The zone may pass through young volcanic centers and geothermal areas of the Long Valley caldera–Mono basin area of California (Fig. 8) and along at least six other Cenozoic calderas (Ekren et al., 1976; Best et al., 1989b), including the probable syntectonic caldera (terminology of Rowley and Anderson, 1996) of the Candelaria Hills (Fig. 8; Speed and Cogbill, 1979; Robinson and Stewart, 1984; Hardyman and Oldow, 1991). The transverse zone passes through many mining districts, including Tonopah (Tingley, 1992). It includes east- to northeast-striking, late Miocene to Quaternary faults in and east of the Lone Mountain area (Fig. 8; Stewart and Carlson, 1978; G. L. Dixon, 1996, oral commun.).

The Blue Ribbon transverse zone is at least 280 km long and 25 km wide (north-south) in western Utah and eastern Nevada, and it extends westward another 100 km across the aeromagnetic quiet zone to join the Warm Springs transverse zone (Rowley et al., 1978). Thus the Blue Ribbon–Warm Springs transverse zone is 600 km long, and crosses virtually the entire Great Basin and transition zone. Hurtubise (1994) confirmed the connection across the quiet zone between the Warm Springs and Blue Ribbon zones through mapping of the Silver King lineament, which contains east-striking faults, interruptions of topography, a caldera, and mining districts. He noted a change of structural polarity across it, which he recognized as an accommodation zone. The Blue Ribbon zone is identified on the basis of interruptions and alignments of magnetic anomalies, east-trending range fronts (e.g., the northern Black Mountains, Fig. 6) and other topographic features, rhyolite eruptive centers of 26–5 Ma, andesite stratovolcano centers of the 32(?)–21 Ma Mount Dutton Formation (west of Circleville, Fig. 6), altered and mineralized areas, east-striking faults, and hot springs (Rowley et al., 1978, 1994a). It crosses through the center of the huge 32–27 Ma Indian Peak caldera complex (Figs. 1 and 6; Best et al., 1989a). In the eastern part of the zone, intrusions and altered and mineralized rocks are rare south of the zone. East-southeast–striking horsts and grabens in the northern Markagunt Plateau (Fig. 6), along the southern edge of the Blue Ribbon transverse zone, formed after deposition of regional ash-flow tuffs of the 27 Ma Isom Formation and during deposition of both the local ash-flow tuffs of the 26 Ma Buckskin Breccia and eolian sandstone of the 26–25 Ma Bear Valley Formation (J. J. Anderson, 1971, 1988). The Buckskin Breccia and Bear Valley Formation accumulated to thicknesses of 200 to 300 m in the grabens. East of the Markagunt Plateau, en echelon basin-range faults of the Sevier fault zone (at least 2 km of normal offset), which defines the western edge of the Sevier Plateau (Fig. 6), change strike (north-northwest to the north, north-northeast to the south) where they cross the transverse zone (Rowley, 1968), a characteristic feature of transverse zones (cf. Taylor et al., 1991).

The northern side of the Marysvale volcanic field is controlled by the east-striking Cove Fort transverse zone, which is at least 150 km long (Figs. 3 and 6). It was first noted by Cook and Montgomery (1974; incorrectly located at lat 28.6° N, a typographical error for 38.6°), who suggested that it was a transform fault that accommodated 40 km of right-lateral offset of the sharp gravity gradient marking the Basin and Range–Colorado Plateau boundary (Fig. 4) (Cook et al., 1989, 1990). Stewart (1983) identified it on the basis of aeromagnetic anomalies and called it the Black Rock lineament. It has strong aeromagnetic expression (Fig. 5) on both regional (Zietz et al., 1976; Blank and Kucks, 1989; V. Bankey, 1992, written commun.) and local (Campbell et al., 1984; Cook et al., 1984; T. A. Steven, 1992, written commun.) scales. The central part of the Cove Fort transverse zone includes the east-striking Cove Creek fault (Steven and Morris, 1983), along which is the Cove Fort geothermal area (Ross and Moore, 1994). The Cove Creek fault is known to have down-to-the-south offset that predates the 23 Ma Osiris Tuff, the second largest ash-flow sheet in the Marysvale field. The tuff is derived from the Monroe Peak caldera (Steven et al., 1984, 1990), the northern side of which may be controlled by the transverse zone (Fig. 6). Just east of the Cove Creek fault, the east-striking Clear Creek downwarp between the Tushar Mountains and Pavant Range (Fig. 6) exhibits at least 2 km of long-lived (from about 20 to 10 Ma) structural relief (Anderson and Barnhard, 1992). The eastern end of the downwarp terminates against the northeast-striking Dry Creek fault zone (Fig. 6; Anderson and Barnhard, 1992), which is a transfer zone. North of the downwarp, the 27 Ma Three Creeks caldera (Fig. 6), source of the largest ash-flow sheet (Three Creeks Tuff Member of the Bullion Canyon Volcanics) in the Marysvale field (Steven et al., 1984), may be partly controlled by the transverse zone. Other features of the transverse zone are intrusions south of it (Steven et al., 1990), the Kimberly gold district south of the downwarp, and hydrothermally altered rocks, faults, and joints along it. The zone also bounds the Sevier River oval of Steven (1985, written commun.) on the south (Figs. 1 and 6). The oval is a north-northeast–striking oval uplift about 115 km long (Rowley et al., 1998) marked by high gravity (Fig. 4; Thompson and Zoback, 1979; V. Bankey, 1993, written com-

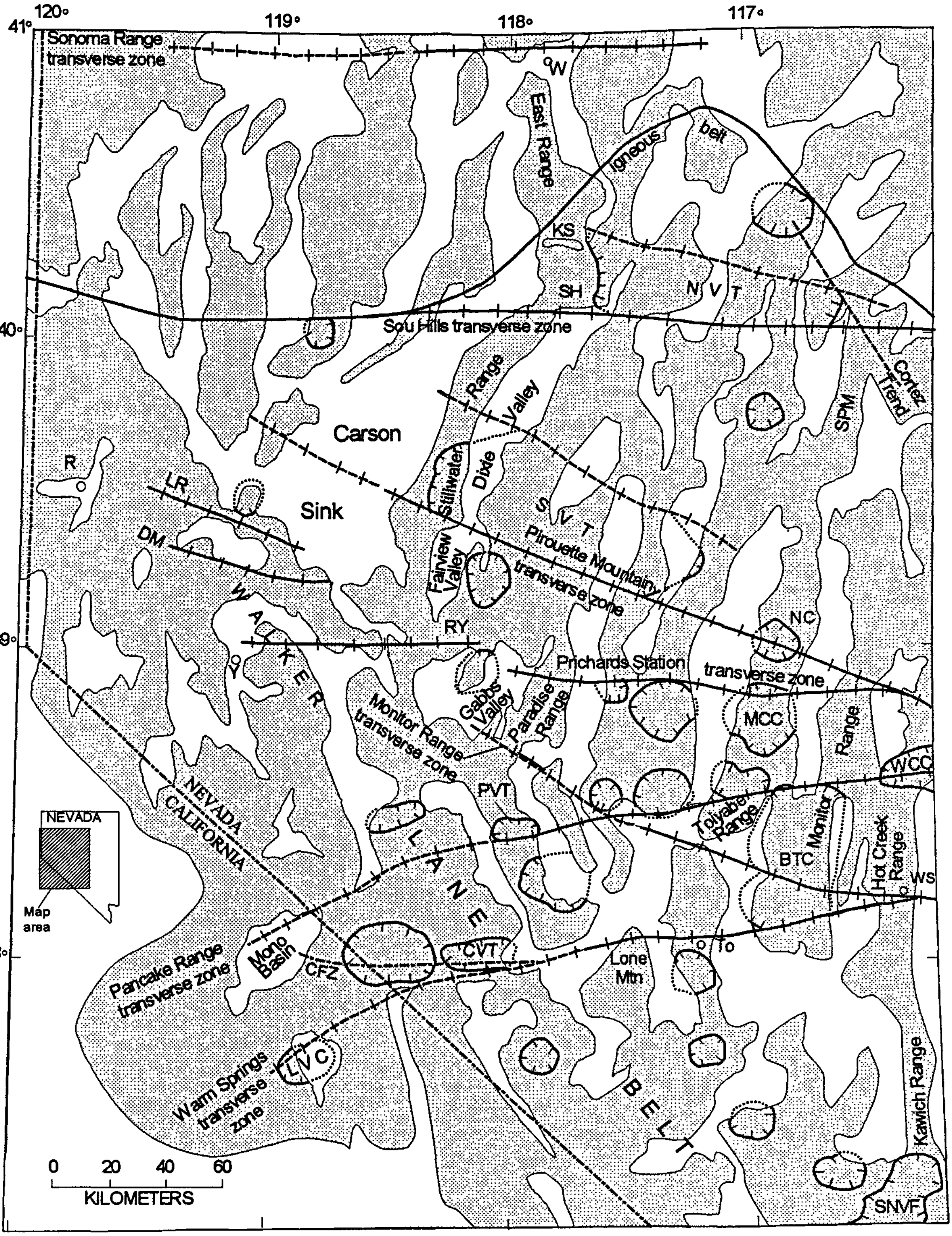

Figure 8. Map showing the geographic setting of igneous belts, transverse zones (crossed lines), calderas, and other geologic features in the western Great Basin. Calderas are shown by closely spaced hachures on fine lines. Ranges are shaded; valleys are undecorated. R, Reno; SH, Sou Hills; SPM, Simpson Park Mountains; To, Tonopah; W, Winnemucca; WS, Warm Springs, Y, Yerington. Geologic features: BTC, Big Ten Peak caldera; CFZ, Coaldale fault zone; CVT, Candelaria volcano-tectonic trough; DM, Desert Mountains structural and volcanic zone; KS, Kennedy stock; LVC, Long Valley caldera; LR, Lahontan Reservoir structural and volcanic zone; MCC, Mount Jefferson caldera complex; NC, Northumberland caldera; NVT, northern volcano-tectonic trough of Burke and McKee (1979); PVT, northern Pilot Mountains volcano-tectonic trough; RY, Rawhide-Yerington transverse zone; SNVF, Southwest Nevada volcanic field; SVT, southern volcano-tectonic trough of Burke and McKee (1979); WCC, Williams Ridge–Hot Creek Valley caldera complex.

mun.). Steven interpreted it to be a late Miocene or younger, incipient (little eroded) metamorphic core complex.

Caliente caldera complex

Mapping indicates that east-striking transverse zones define the northern and southern sides of the large, 23–13 Ma Caliente caldera complex (Figs. 1 and 6; Rowley et al., 1995). The northern zone, the east-striking Timpahute "lineament" of Ekren et al. (1976), is recognized from interruptions and alignments of aeromagnetic anomalies, east-trending range fronts (e.g., Timpahute Range, Bull Valley Mountains) and other topographic features, many east-striking faults of mostly strike-slip displacement, folds, intrusions, silicic feeders and plugs, and an apparent east-trending zone of recent earthquakes (Figs. 3 and 6). The transverse zone underlies several small mining districts (Tingley, 1992) and several hot springs (Blackett, 1994; Garside, 1994). It separates areas of contrasting structural styles, stress configurations (Ekren et al., 1976, 1977), and magnitudes of extensional strain (Hudson et al., 1993; Axen, 1994; Rowley et al., 1992). Ekren et al. (1976) observed that both right-lateral and left-lateral faults have been mapped along the zone, which is another attribute of transverse zones, and they were the first to conclude that it controlled the northern side of the Caliente caldera complex. West of the Caliente caldera complex, east-striking strands of the Timpahute transverse zone include the major Pahroc Valley fault (Fig. 6; R. B. Scott et al., 1992), which has subhorizontal striae with both left- and right-lateral indicators, as well as faults of apparent normal and oblique offset (Scott and Swadley, 1992; Swadley and Rowley, 1994). Ash-flow units of the Needles Range Group (31–28 Ma; Best et al., 1989a) were interpreted by R. B. Scott (1992, oral commun.) to have ponded against a north-facing scarp made by these and other faults. Farther west along the transverse zone, a prominent east-striking Tertiary fault in the Pahranagat Range (Fig. 6) has about 1.5 km of left-lateral displacement (Tschanz and Pampeyan, 1970). Farther westward, east-striking faults are abundant in the Timpahute Range (Ekren et al., 1977). To the east, the Timpahute transverse zone can be mapped at least to the Iron axis, for a total length of at least 220 km. The evidence includes east-striking lower to middle Miocene lava flows along the northern Bull Valley Mountains, an east-striking graben at least 20 km long that contains about 300 m of coarse clastic sedimentary rocks of 17–12 Ma, and the Newcastle geothermal area (e.g., Siders et al., 1990; Siders, 1991).

The southern side of the Caliente caldera complex is defined by the Helene transverse zone, named for a major east-striking fault that underlies the ghost mining town of Helene (Figs. 3 and 6; Best et al., 1993; Scott et al., 1996). The fault contains a large rhyolite dike, and this and other dikes of the same strike are inferred to be derived from an underlying granitic intrusion that controlled the mineralization of the major Delamar epithermal gold district (Fig. 6). Rhyolite lava flows fed by the dikes locally spilled into the caldera complex and intertongue with the 18.2 Ma Hiko Tuff intracaldera fill (Rowley et al., 1995). The transverse zone contains many other major east-striking faults, rhyolite dikes, plugs, and volcanic domes, an andesite breccia pipe, hydrothermally altered and mineralized rocks, and two small epithermal gold districts (Taylor mine and Pennsylvania district, Fig. 6; Rowley et al., 1992) along the 40 km length that I have mapped so far. It likely defines the southern margin of the Caliente caldera complex for at least another 30 km to the east. East-striking faults mapped in and west of the aeromagnetic quiet zone by Ekren et al. (1977) suggest that the transverse zone continues west at least across the South Pahroc Range, for a total length of 110 km.

Other zones in southern Nevada

Ekren et al. (1976) described two other, mostly east-striking, transverse zones north of the Warm Springs transverse zone. From north to south, these are the Prichards Station and Pancake Range "lineaments" (Figs. 3, 6, and 8). The Prichards Station transverse zone is defined by interruptions of ranges and aeromagnetic anomalies and by east-striking faults of large displacement, one of partly normal sense but most of left-lateral sense. This transverse zone bounds the sides of at least four calderas and the Mount Jefferson and Williams Ridge–Hot Creek Valley caldera complexes (Figs. 3 and 8; Best et al., 1989b) and underlies many mining districts (Tingley, 1992) and several hot springs (Garside, 1994). Its central part was mapped in detail and is expressed by large, east-striking, left-lateral faults, Tertiary thrusts, and wide zones of breccia (Dixon et al., 1973; Ekren et al., 1974). Near its eastern end, it includes the major east-striking Currant Summit fault zone, which may have 2.4 km of left-lateral displacement and may be a "structural discontinuity separating two areas which have developed differently in response to a different stress situation" (Fig. 6; Moores et al., 1968, p. 1716). The transverse zone is about 230 km long and probably continues farther west, across a gap of 40 km, and connects with the Rawhide-Yerington lineament of Bingler (1971), which is about 80 km long and crosses much of the Walker Lane belt (Figs. 3 and 8). The Rawhide-Yerington transverse zone is recognized by a string of intrusions, east-striking faults, and major metallic ore deposits (Bingler, 1971). Just east of the transverse zone, a seismic gap in historic earthquakes may be controlled by east-west structures (Doser, 1986).

The Pancake Range transverse zone is characterized by interruptions of ranges and aeromagnetic anomalies, east-striking faults, and differences in structural style (Ekren et al., 1976). It is about 330 km long; its western end probably strikes west-southwest and continues into the eruptive center of Mono basin, California, after crossing the Walker Lane belt (Figs. 3 and 8). Ekren et al. (1976) interpreted the transverse zone to pass through the Williams Ridge–Hot Creek Valley caldera complex (Fig. 8) and along the sides of at least five other calderas (Best et al., 1989b), including a probable syntectonic caldera (Rowley and Anderson, 1996) in the northern Pilot Mountains (Fig. 8; Hardyman and Oldow, 1991). The transverse zone passes through many mining districts (Tingley, 1992) and several warm springs (Garside, 1994). Detailed mapping (Moores et al., 1968; Ekren et al., 1973; Quinlivan and Rogers, 1974) shows that large strike-slip faults of apparent left-lateral displacement mark its eastern part.

A possible transverse zone was suggested by J. H. Stewart (1996, written commun.) to extend from Warm Springs west-northwest along the southern side of the Toiyabe Range (where it crosses the Pancake Range transverse zone), then on to Gabbs Valley, for a total length of 160 km (Figs. 3 and 8). In most places it is not exposed but follows faults of the same strike across the Hot Creek Range (Fig. 8; Whitebread and John, 1992) and has strong gravity expression (Saltus, 1988a, 1988b). In the Monitor Range, it has been interpreted either to follow faults of the same strike (Whitebread and John, 1992) or to follow the margin of the Big Ten Peak caldera (Fig. 8), where it is inset into an older caldera to the south (Ekren et al., 1976, Plate 1). Thus, I call this zone the Monitor Range transverse zone.

Two west-northwest–striking transverse zones characterized by faults and volcanic rocks are present north of the Rawhide-Yerington transverse zone. They were called the Lahontan Reservoir structural and volcanic zone (on the north) and the Desert Mountains structural and volcanic zone (on the south) (John et al., 1993). Each is 60 km long and 10 km wide, contains rhyolite flows and domes of 14–8 Ma (John et al., 1993), includes many faults of the same strike (Greene et al., 1991) and possibly the same age, and crosses the northern Walker Lane (Figs. 3 and 8).

The exact traces of the Pancake Range and Warm Springs zones across the north-northwest–trending Walker Lane belt are poorly defined. Despite late Cenozoic right-lateral slip on north-northwest–striking faults along the Walker Lane, many discontinuous east-striking features are in the belt, some of which are tentatively assigned to these two transverse zones. Although not yet confirmed by mapping, the two zones and the faults of the Walker Lane are partly coeval and likely offset each other. The area of intersection of the Walker Lane belt with the Lahontan Reservoir, Desert Mountains, Rawhide-Yerington, Prichards Station, Pancake Range, Monitor Range, and Warm Springs transverse zones is one of the major mineral belts in the world (John et al., 1989; Hardyman and Oldow, 1991). Slemmons (1967) and Stewart (1980) projected the southern of their three transverse zones into this complex area, which also coincides with the White Mountain seismic gap (Fig. 3; R. E. Wallace, 1978, 1984) of the Central Nevada seismic belt.

The Pahranagat shear system is a northeast-striking fault zone about 50 km long and 20 km wide, the faults of which may have 16 km of cumulative left slip (Figs. 3 and 6; Tschanz and Pampeyan, 1970; Ekren et al., 1977). Liggett and Ehrenspeck (1974) suggested that it was a continental transform fault that accommodated differential extension and passed at both ends into north-striking basin-range faults. Detailed mapping in the eastern part of the system (Scott et al., 1990, 1993; Page et al., 1990; Swadley and Scott, 1990; G. L. Dixon, 1995, oral commun.) confirms these views. The Pahranagat shear system is a relatively youthful transfer zone that contains basalt feeders.

The east-northeast–striking Garlock fault zone is a transverse zone (Slemmons, 1967); the kinematics were well described by Davis and Burchfiel (1973), who called it an "intracontinental transform structure." It had at least 60 km of left-lateral displacement since the early Cenozoic (Louie and Qin, 1991) and defines the southwestern edge of the Great Basin (Figs. 1 and 3). Hamilton and Myers (1966) noted it as a province boundary with a transform-like nature because basin-range structure north of its west end is missing in the Mohave Desert to the south. Davis and Burchfiel (1973) and Wernicke (1992) concluded that it separates crustal blocks of different behavior and that displacement drops to zero at its eastern end, where extension on the northern side equaled that of the southern side. Rogers et al. (1991, p. 165) noted that the Garlock now appears to be locked and has high potential for a large earthquake. COCORP seismic reflection data were interpreted to indicate that the Garlock fault intersects a detachment fault at about 9 km depth and is therefore a tear fault in the hanging wall of a master detachment (Cheadle et al., 1986; Louie and Qin, 1991) rather than the major through-going fault proposed by Davis and Burchfiel (1973). Serpa (1990), however, reinterpreted the seismic line across the fault and concluded that a detachment fault at 9 km depth stops at the Garlock and is confined to the area under the Mohave Desert.

The west-northwest–striking, right-lateral Las Vegas Valley shear zone and the northeast-striking, left-lateral Lake Mead fault system, along with other major faults, form the southern margin of the Great Basin sector of the Basin and Range (Fig. 3). These transverse faults are surrounded by complex late Cenozoic structures that defy simple explanations. Fleck (1970) suggested that the Las Vegas Valley shear zone is a transform fault that accommodated different amounts of extension and magmatism. Davis and Burchfiel (1973) and Liggett and Childs (1977) offered similar interpretations. Duebendorfer and Black (1992) evaluated the slip budget along the Las Vegas Valley shear zone in terms of its variable displacements (as much as 48 km, almost half its length), evidence of opposite strike-slip movement along strike, and abrupt termination along strike. They considered the shear zone to be closely related to variable distribution of extension in adjacent blocks and to form "one of the largest and best-studied transverse structures in the Basin and Range." In investigations along the large-displacement (about 65 km) Lake Mead fault system, Anderson (1973) and Bohannon (1984) also considered a transform origin for it. Faulds et al. (1990) suggested that the Lake Mead fault system and Las Vegas Valley shear zone are "intracontinental transform faults associated with an axis of extension in the Colorado River extensional corridor." Duebendorfer and Black (1992), Wernicke et al. (1982, 1988), and others, however, interpreted that the primary structures in the area are detachment faults, and the Las Vegas Valley shear zone and Lake Mead fault system are tear faults in the hanging wall of a detachment. Ron et al. (1986) and Campagna and Aydin (1994), however, concluded that the Las Vegas Valley and Lake Mead structures are fundamental structures that control the geometry of some normal faults in the region.

An alternative view of the Lake Mead fault system and the Las Vegas Valley shear zone is that these structures bound the northern side of an area of tectonic escape, in which upper crustal rocks moved westward on a current of lower-crustal ductile rocks (Anderson, 1973, 1990; Anderson et al., 1994). In this model, the westward ductile crustal flow occurs below a shallow (10 km or greater; Anderson et al., 1994) brittle-ductile transition zone and includes north-south shortening and emplacement of the synextensional Wilson Ridge pluton. The Las Vegas Valley shear zone and Lake Mead fault system parallel the flow direction (Anderson et al., 1994).

An east-trending discontinuity or boundary at about lat 37° N in the southern Great Basin extends between northern Death Valley, California, and St. George, Utah (Fig. 3; Eaton, 1975). The 37° discontinuity is defined by a strong north-facing gravity gradient (Saltus and Jachens, 1995), moderate seismicity, a trough in the Moho discontinuity, and different seismic velocities (Eaton, 1975, 1979; Eaton et al., 1978). Eaton (1980) drew his boundary between the northern and southern Basin and Range on this discontinuity. The discontinuity is expressed by a southward decrease in topography of about 800 m (Saltus and Thompson, 1995) and changes in the geometry of basin-range topography (Best and Hamblin, 1978). The discontinuity coincides with the northern edge of the amagmatic corridor (Anderson and Barnhard, 1993), which is at the southern edge of the igneous belts shown in Figure 3. The discontinuity also marks the southern edge of volcanic rocks and mining districts (Stewart et al., 1977), the northern edge of exposed Precambrian rocks older than 800 Ma, and the locus of abrupt northward thickening of the Phanerozoic section (Eaton, 1975). Seismicity along the discontinuity is the southwestern tail of the Intermountain seismic belt (Fig. 1). Because earthquake focal mechanisms display left-lateral motion, Suppe et al. (1975) concluded that the discontinuity may be an incipient strike-slip fault zone whose westward expression is the Garlock fault zone. Saltus and Thompson (1995), however, attributed the changes across the discontinuity to asthenospheric buoyancy related to the southern edge of the initial plume head of the Yellowstone hotspot (e.g., Pierce and Morgan, 1992; Zoback et al., 1994) to the north. In contrast, I suggest that the 37° discontinuity may be a transverse zone that is both profound and enigmatic, has a deep-crustal expression, and includes parts of the Las Vegas Valley shear zone and Lake Mead fault system.

Northeastern Great Basin

Linear, more or less east-striking faults are well known in the northeastern Great Basin. Stewart et al. (1977) and Stewart (1983) showed three east-striking lines in northeastern Nevada, just south of the Nevada-Idaho State line, drawn mostly on the basis of geophysical anomalies (Zietz et al., 1976; Hildenbrand and Kucks, 1988b; Saltus, 1988b). These anomalies are due to east-trending Cretaceous (?) intrusions (Stewart and Carlson, 1978) and by Tertiary rhyolite domes, basaltic centers, and small mining districts (Tingley, 1992). Short east-striking faults are present locally, but most of the anomalies are covered by younger volcanic rocks. Stewart's lines probably define parts of a transverse zone, which I call the Mountain City transverse zone for the town that it passes through; it is 180 km long (Figs. 3 and 9).

Zoback (1983) interpreted gravity anomalies in Utah to indicate many short, east-striking "transverse zones" that truncate ranges and basins, and she suggested that they represent faults. Cook and Montgomery (1975) and Cook et al. (1990) suggested a "lineament," which they considered to be a transform fault, at lat 40.6° N on the basis of a deflection of the sharp gravity gradient that marks the Basin and Range–Colorado Plateau boundary south of Salt Lake City (Fig. 4; Cook et al., 1989). Cook and Montgomery (1974) and Cook et al. (1990) considered it to strike east, on line with the crest of the Uinta Mountains (Uinta arch), but they were not able to constrain its strike by other gravity anomalies. A better expression of this "lineament" is a prominent 100-km-long, east-northeast–striking line that marks the northern side of aeromagnetic anomalies (Zietz et al., 1976) of the "Oquirrh-Uinta mineral belt" of Hilpert and Roberts (1964). I rename this east-northeast line and its western extension the Oquirrh-Uinta transverse zone (Figs. 3 and 9). It contains many geophysical anomalies, faults of the same strike, plutons, and major mining districts such as Ophir, Bingham, and Park City (Fig. 9). Its eastern end ("Oquirrh-Uinta mineral belt") is better defined than other parts because its middle part passes along the southern side of the Great Salt Lake Desert (Fig. 9), where geophysical anomalies are generally concealed under thick basin-fill deposits. The western part of the zone is defined by aeromagnetic anomalies of the same trend as its eastern part. As thus proposed, it extends about 230 km west-southwest, from 50 km east of the Wasatch front to the northern Deep Creek Range (Fig. 9), which contains the Gold Hill mining district. In the Gold Hill district, faults of the same strike as the transverse zone are mapped, as are Jurassic and late Eocene intrusions (Moore and Sorenson, 1979; Moore and McKee, 1983). Crittenden et al. (1973, Fig. 1) summarized the isotopic ages (41–25 Ma) and field relations of intrusions that mark the eastern part of the transverse zone. The point where the transverse zone crosses the north-striking Wasatch fault zone is marked by the Little Cottonwood stock (32–25 Ma), and here the Wasatch fault is offset 6 km in a left-lateral sense by the east- to east-northeast–striking Deer Creek fault, which belongs to the transverse zone (Fig. 9). Zoback (1983) noted east-northeast–striking normal faults and interruptions of gravity anomalies along and west of the Deer Creek fault, where seismic data along east-northeast–striking faults are interpreted to represent a nonconservative barrier containing oblique (right lateral and normal) offset (Bruhn et al., 1992). As used here, the Oquirrh-Uinta transverse zone is the more conservatively drawn part of the "Uinta-Gold Hill trend" of Erickson (1976), the "Uinta trend" of Moore and McKee (1983), and the "Bingham–Gold Hill mineral trend" of Stein et al. (1989), all of which extend from the Uinta Mountains, through the Gold Hill mining district, into Nevada. Erickson (1976) noted that deposition of many Proterozoic and Paleozoic sedimentary units along the line seem to have been con-

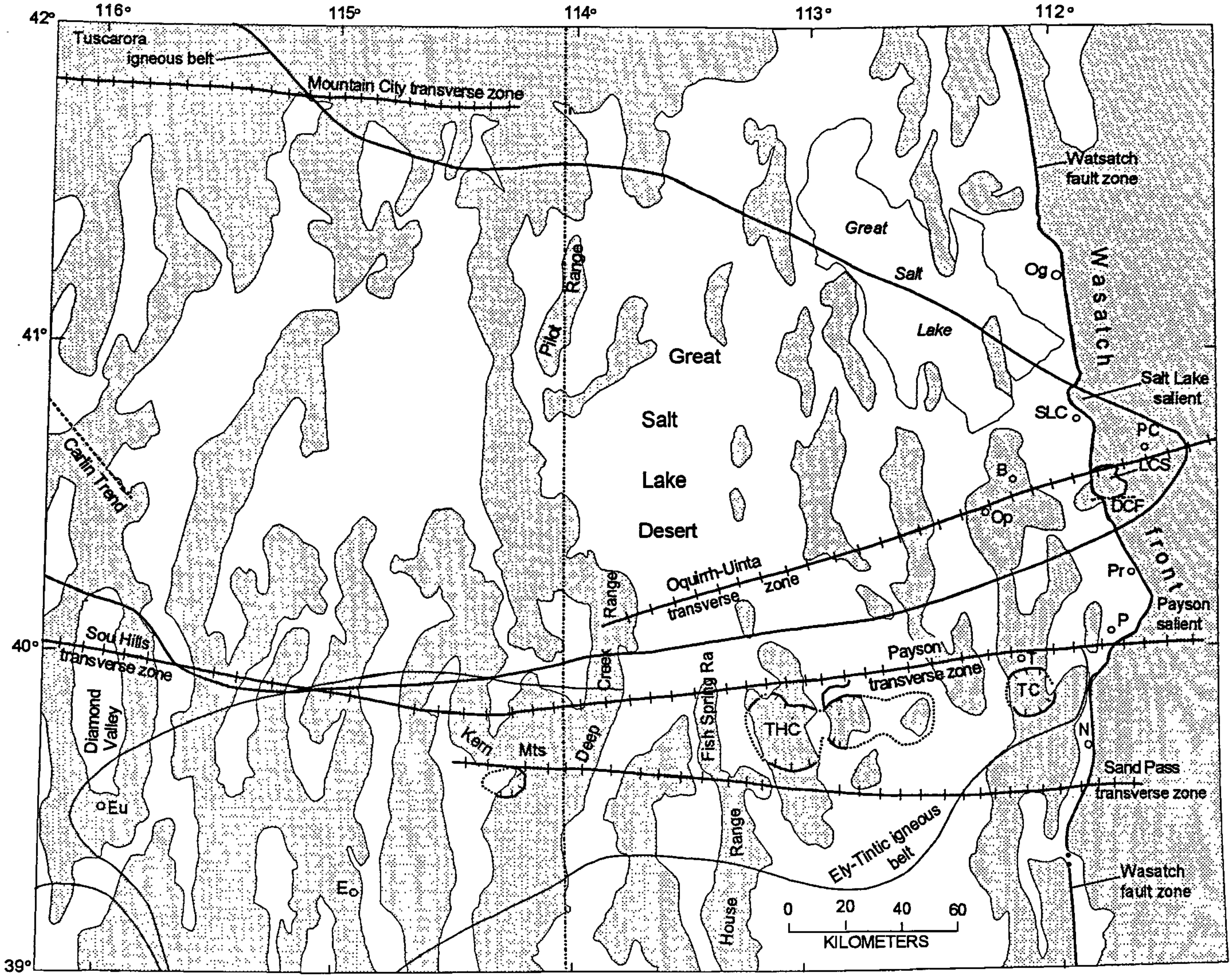

Figure 9. Map showing the geographic setting of igneous belts, transverse zones (crossed lines), calderas, and other geologic features in the northeastern Great Basin. Calderas are shown by closely spaced hachures on fine lines. Ranges and Colorado Plateau are shaded; valleys are undecorated. B, Bingham; E, Ely; Eu, Eureka; N, Nephi; Og, Ogden; Op, Ophir; P, Payson; PC, Park City; Pr, Provo; SLC, Salt Lake City; T, Tintic. Geologic features: DCF, Deer Creek fault; LCS, Little Cottonwood stock; TC, Tintic caldera; THC, Thomas caldera.

trolled by coaxial east-trending topography and structures. Abundant evidence suggests control of the Uinta Mountains by east-striking structures (Wheeler and Krystinik, 1988).

Zoback (1983) suggested that another deflection of the Wasatch fault zone, at the Salt Lake salient (Fig. 9) just north of Salt Lake City, represents the influence of another east-striking "transverse zone." This zone also is marked by gravity and aeromagnetic anomalies (Mabey, 1992), a west-trending basement ridge and fault zone (Bruhn et al., 1992), and interruptions of gravity gradients (Cook et al., 1989). By the same evidence, it could be extended to the west at least 100 km, but it is poorly constrained and therefore not shown in Figure 3.

Farther south, in the Ely-Tintic igneous belt (see following discussion), Fuller (1964, Fig. 2) suggested an east-west "fracture" marked by aeromagnetic anomalies (later shown by Zietz et al., 1978) extending eastward from Ely, Nevada (Fig. 1), for about 70 km. A faulted middle Cretaceous stock, which controls the Robinson mining district, is present at Ely, but the "fracture" does not appear to be geologically expressed otherwise (Stewart and Carlson, 1978).

Cook and Montgomery (1974) and Cook et al. (1990) suggested a "lineament," which they proposed was a transform fault, at lat 40° N, on the basis of a crudely defined east-west line that separates gravity anomalies and deflects, at Payson, Utah, the sharp gravity gradient that marks the Basin and Range–Colorado Plateau boundary (Cook et al., 1989). Deflections of gravity anomalies can be traced another 50 km to the east. Zoback (1983), Wheeler and Krystinik (1988, 1992), Mabey (1992), and Machette et al. (1992) called attention to this, the Payson salient (Fig. 9), as an example of a complex boundary between segments of the Wasatch fault zone. The lineament is here called the Payson transverse zone and is at least 250 km long (Figs. 3 and 9). The zone marks the northern side of part of the Ely-Tintic igneous belt, which is delineated by aeromagnetic anomalies (Zietz et al., 1976, 1978; Hildenbrand and Kucks, 1988b). The zone passes westward to an Eocene intrusion in the southern Deep Creek Range (Hintze, 1980), then perhaps 20 km into Nevada. Some ranges are terminated along it and others north of it appear to be dragged in a left-lateral sense. It bounds the northern side of several calderas (Stoeser, 1993) and passes through the northern Tintic mining district and several hot springs (Fig. 9; Blackett, 1994).

Stoeser (1993) suggested four west- to west-northwest–striking "lineaments" on the basis of interruptions of gravity gradients in the eastern Ely-Tintic igneous belt. Three have little support other than the gravity data, but the fourth, which Stoeser called the Sand Pass "lineament," defines the southern side of the Ely-Tintic belt. Stoeser traced it from the Utah-Nevada border to 25 km west of the Basin and Range–Colorado Plateau boundary (Figs. 3 and 9). He concluded that it is commonly coaxial with faults, aligns with gravity interruptions and range fronts, separates a difference in structural polarity between the Fish Springs Range to the north and the House Range to the south (Fig. 9), is underlain by intrusions at Sand Pass (separating the Fish Springs and House Ranges), and coincides with several hot springs (Blackett, 1994). Stoeser (1993) suggested that the transverse zone separates the west-dipping Sevier Desert detachment to the south from another detachment (his Tintic Valley detachment) of similar vergence to the north that was presumed to underlie the eastern part of the Ely-Tintic igneous belt. Deflection of the sharp gravity gradient at the Basin and Range–Colorado Plateau boundary (Cook et al., 1989) and notable differences in the geology (Hintze, 1980) north and south of its projected trace suggest that it continues for at least another 25 km east of the boundary. It also continues west-northwest from the Utah-Nevada border and is coaxial with the west-northwest–trending Kern Mountains of extreme eastern Nevada, which contain the Eagle mining district and a composite Late Cretaceous and Oligocene batholith of the same trend (Fig. 9; Best et al., 1974). West of the state border it has mild aeromagnetic expression and passes along the northern side of the caldera source of the 34 Ma Kalamazoo Tuff (Fig. 9; Best et al., 1989b; Gans et al., 1989). The Sand Pass transverse zone is at least 230 km long.

Northwestern Great Basin

The northwestern edge of the Great Basin is defined by the west-northwest–striking, right-lateral Brothers fault zone (Fig. 1), which appears to be a transverse zone. At least for the episode of extension 14–6 Ma, it has been oriented parallel to the extension direction (Hart and Carlson, 1987). The rocks north of the fault zone are relatively unextended, whereas the rocks to the south are extended (Lawrence, 1976). The Brothers fault zone is a fundamental feature, despite its "relatively limited visible offset," because of the concentration of igneous vents along it (Lawrence, 1976). South of the Brothers fault zone, a 200-km-long, similarly striking belt of 19–7 Ma igneous centers, called the Orevada rift (Fig. 1; Rytuba and Conrad, 1981) may have an origin similar to that of the Brothers zone. McKee and Noble (1986) suggested that the Brothers zone and Garlock fault served as transforms accommodating extension within the Great Basin.

Two possible east- to east-southeast–trending tuff depocenters in western Nevada, called volcano-tectonic troughs by Burke and McKee (1979), may reflect transverse zones. Burke and McKee (1979) suggested that the apparent troughs are bounded by faults that were active during volcanism and tuff deposition. The isostatic gravity map and basement gravity map (Saltus, 1988b; Jachens and Moring, 1990) show both troughs as east-southeast–trending gravity lows that cross several ranges, as would be expected for caldera complexes. Stewart (1983) called attention to the troughs and their parallelism with transverse zones. I conclude that, like the Caliente caldera complex, the troughs represent caldera complexes defined by transverse zones on their northern and southern sides; Rowley and Anderson (1996) called them syntectonic calderas. The eastern end of the northern caldera (Figs. 3 and 8) is underlain by a 2.4-km-thick graben filled with the 32 Ma Caetano Tuff. The central part of the syntectonic caldera is underlain by the 24 Ma Fish Creek Mountains caldera (Burke and McKee, 1979). The East Range west of the syntectonic caldera contains the east-elongated, 30 Ma Kennedy stock and associated mining district; the emplacement of the stock was likely controlled by structures that controlled the syntectonic caldera (Fig. 8; A. R. Wallace, 1978). Perhaps the Kennedy stock is an intracaldera pluton eroded below the floor of the caldera after the East Range was uplifted along basin-range faults. If so, the northern syntectonic caldera is at least 110 × 25 km in size.

The southern side of the northern syntectonic caldera is on line with, and 20 km east of, the Stillwater seismic gap (Fig. 3; R. E. Wallace, 1978; Wallace and Whitney, 1984) of the Central Nevada seismic belt. The gap also contains the Sou Hills geothermal area (Garside, 1994). Fonseca (1988) and Wheeler and Krystinik (1992) concluded that the Sou Hills area overlies a persistent, probably nonconservative, barrier to fault propagation, the Sou Hills "transverse zone," which "may be related to the structure that controls the volcano-tectonic trough" (Fonseca, 1988). Aeromagnetic (Hildenbrand and Kucks, 1988b), isostatic residual gravity (Saltus, 1988b), and basement gravity (Jachens and Moring, 1990) anomalies suggest that the west-northwest–striking Sou Hills transverse zone extends as much as 100 km west of the Sou Hills (Figs. 3 and 8). Many mining districts line up along the feature (Tingley, 1992). East of the trough, the transverse zone crosses, with no apparent effect, the north-northwest–trending, sediment-hosted, disseminated gold deposits (Bonham, 1991) of the Cortez trend (Fig. 3; Bagby and Berger, 1985). About 30 km farther east-southeast, the transverse zone follows an east-striking "transverse element" interpreted to be a tear fault and imaged by seismic reflection data (Effimoff and Pinezich, 1981) under Diamond Valley (Fig. 9), east of which is a string of sediment-hosted precious-metal deposits (Bonham, 1991) and other mining districts (Tingley, 1992). Other evidence for the transverse zone includes a series of anomalous east-northeast–striking faults, some of which cut Pleistocene alluvium in the northwestern Simpson Park Mountains (Fig. 8; Stewart and Carlson, 1978). As here defined, the transverse zone has a total length of 320 km. Although not well defined across the quiet zone, it probably connects eastward with the Payson transverse zone, and if so, it crosses nearly the entire Great Basin and High Plateaus. The northern side of the northern syntectonic caldera may be controlled by a parallel transverse zone, as suggested by geophysical anomalies of the same trend and by interruptions of anomalies (Hildenbrand and Kucks, 1988b; Saltus, 1988b).

Burke and McKee's (1979) 100 × 40 km southern "trough" (Figs. 3 and 9) was filled with as much as 5 km of tuff during synchronous faulting at about 30–24 Ma. The central part of the area has not been mapped in detail, and some of the tuffs may be relatively thin outflow sequences between calderas (Hardyman et al., 1988; D. A. John, 1995, oral commun.). John (1995) mapped the Stillwater caldera complex in the Stillwater Range (Fig. 8), in the western end of the syntectonic caldera, and found three calderas ranging in age from 29 to 25 Ma, some of which subsided along synchronous faults. As with the northern syntectonic caldera, the west-northwest projection of the southern side of the southern syntectonic caldera is occupied by a seismic gap that geographically separates historically active faults of the Central Nevada seismic belt. This gap was called the Pirouette Mountain boundary (Fig. 3) by Wheeler (1989) and Wheeler and Krystinik (1992), who considered it to be a persistent, probably nonconservative segment boundary. Wheeler and Krystinik (1992) noted that gravity data show the boundary along a largely buried bedrock ridge that separates Dixie and Fairview Valleys (Fig. 8). The northern side of the ridge and the gap pass westward into an accommodation zone (different structural polarity across it) that passes through the central part of the Stillwater caldera complex (John, 1995, Fig. 2; M. R. Hudson, 1995, written commun.). I suggest that all these features are part of a west-northwest–striking transverse zone, called here the Pirouette Mountain transverse zone, that defines the southern part of the southern syntectonic caldera. Isostatic residual gravity (Saltus, 1988b), mining districts (Bonham, 1991; Tingley, 1992), and hot springs (Garside, 1994) help define this transverse zone and an additional inferred one on the northern side of the syntectonic caldera. If the Pirouette Mountain transverse zone were projected east-southeast through the Northumberland caldera (Fig. 8) and gold district (33 Ma; Best et al., 1989b), it terminates at the Prichards Station transverse zone. The Pirouette Mountain zone is at least 180 km long and might extend another 50 km to the west-northwest through a series of geothermal areas (Garside, 1994) in and west of the Carson Sink.

Thenhaus and Barnhard (1989) proposed the Sonoma Range seismic gap (Fig. 8) at the northern edge of the 1915 Pleasant Valley fault scarps. They drew the western end of the northern transverse zone of Slemmons (1967) and Stewart (1980) through the northern part of this seismic gap. I here refer to this as the Sonoma Range transverse zone (Figs. 3 and 8). In the seismic gap, ranges are terminated along an east-west line that passes through Winnemucca (Fig. 1). East-trending geophysical anomalies, interruptions of anomalies (Hildenbrand and Kucks, 1988b; Saltus, 1988b), mining districts (Bonham, 1991; Tingley, 1992), hot springs (Garside, 1994), and the southern end of precious-metal deposits of the north-northeast Getchell trend (Fig. 3; Bagby and Berger, 1985) are along the zone, which is about 200 km long.

IGNEOUS BELTS

Butler et al. (1920) recognized that the mining districts of the Great Basin could be lined up into "uplifts" or "mineral belts" that trend generally east-northeast in the eastern half of the Great Basin and northwest or west-northwest in the western half. Data about these belts were compiled by many workers (Hilpert and Roberts, 1964; Roberts, 1964, 1966; Shawe and Stewart, 1976; Stewart et al., 1977; Bagby, 1989). Rowley et al. (1979) used the term "igneous belt," rather than mineral belt, because virtually all the mineral deposits plotted by Shawe and Stewart (1976) were localized by intrusive rocks, which are cupolas on batholiths and stocks that largely underlie the belts (Steven et al., 1984). The shape of some of the belts (Fig. 1), but not their overall trend, was later modified to reflect aeromagnetic anomalies, which delineate the intrusive rocks (Stewart et al., 1977; Rowley et al., 1978; Zietz et al., 1976, 1978; Mabey et al., 1978). Outflow deposits, primarily ash-flow tuffs, are commonly between igneous belts but do not generally result in aeromagnetic anomalies.

Most aeromagnetic anomalies in igneous belts are due to middle Cenozoic calc-alkaline intrusions. Most belts have the same trend as the middle Cenozoic east-northeast extension direction (Zoback et al., 1981). A significantly smaller volume of the belts is due to high-silica rhyolites of the upper Cenozoic bimodal sequence. In contrast, basalts of the upper Cenozoic bimodal sequence are not in belts. Exceptions to the east-northeast trend of the middle Cenozoic igneous belts include the west-northwest volcano-tectonic troughs of Burke and McKee (1979) and the east-west Caliente caldera complex and Marysvale igneous centers.

Calc-alkaline igneous belts in and north of the Great Basin are generally younger from north to south (Fig. 7). The Eocene (54–43 Ma) Challis igneous belt ("Challis magmatic belt" of Christiansen and Yeats, 1992; Luedke, 1994) spanned northeastern Washington, most of Idaho, southwestern Montana, and northwestern Wyoming, all north of the eastern Snake River Plain. In the Great Basin, the oldest calc-alkaline igneous rocks are about 43 Ma (Brooks et al., 1995). The youngest calc-alkaline igneous rocks in the Great Basin are 23–17 Ma in the Caliente caldera complex (Rowley et al., 1995). Eruption of the high-silica rhyolites of the bimodal suite generally seems to have begun at about 20 Ma throughout the Great Basin. However, in the southeastern Great Basin, high-silica (bimodal) rhyolites also are younger southward, from 17–13 Ma in the Caliente caldera complex (Rowley et al., 1995) to 15–13 Ma in the Kane Springs Wash caldera complex (Fig. 6; Scott et al., 1995a, 1995b). In the southwestern Great Basin, high-silica rhyolites are younger westward and southwestward, from 15–11 Ma in the southwestern Nevada volcanic field of the Nevada Test Site (Figs. 1 and 3; Sawyer et al., 1994) to about 6 Ma in California (Luedke and Smith, 1981).

An important aspect of igneous belts in the Great Basin is that their southward migration was accompanied by generally north-striking normal faults (Anderson, 1989; Armstrong and Ward, 1991; Seedorff, 1991; Gans et al., 1989; Dilles and Gans, 1995; Scott et al., 1995b). Thus, the western end of the Pioche-Marysvale igneous belt (Fig. 6) contains extensional faults and unconformities that indicate deformation of the same age (31–27 Ma) as the volcanic rocks (Bartley, 1989; Scott et al., 1995c). In and near the Caliente caldera complex, the main episodes of magmatism and extension coincided at about 22–16 Ma (Rowley et al., 1992, 1995). In the Kane Springs Wash caldera complex (Fig. 6), the main episodes of magmatism and extension coincided at 17–14 Ma (Scott et al., 1995a, 1995b). In the Las Vegas area, the main episodes of magmatism and extension coincided at about 14–10 Ma (Anderson et al., 1994).

Pioche-Marysvale belt

The Pioche-Marysvale igneous belt extends from the aeromagnetic quiet zone of Nevada east-northeast through the northern and central Marysvale volcanic field (Figs. 3 and 6). It was called the Beaver uplift by Butler et al. (1920), the Wah Wah–Tushar belt by Hilpert and Roberts (1964), and the Pioche mineral belt by Shawe and Stewart (1976). Its overall east-northeast trend includes some strongly east-trending aeromagnetic anomalies (Fig. 5; Zietz et al., 1976; Steven et al., 1984; V. Bankey, 1995, written commun.), especially in the westernmost and easternmost parts of the belt. The east trends may reflect the local middle Cenozoic extension direction or dismembered pieces of the overall belt due to later east-striking transverse zones. The aeromagnetic signature of the western part of the belt forks (Fig. 5); the south fork is located east and south of Pioche (Stewart et al., 1977). With inclusion of the south fork, the overall trend of the belt is east-northeast. As Steven et al. (1984) noted, the entire belt, and thus the source of the aeromagnetic anomalies, is underlain by a batholith complex of largely middle Cenozoic calc-alkaline rocks. These igneous rocks are generally younger from west to east (Steven et al., 1984), from the huge Indian Peak caldera complex (as old as 31 Ma; Best et al., 1989a) to the Monroe Peak caldera (as young as 22 Ma; Steven et al., 1984). The igneous belt also includes much less abundant high-silica rhyolite (bimodal) rocks, which are not consistently younger eastward. These rhyolites include the Blawn and Steamboat Mountain Formations in and near the Indian Peak caldera complex, which are 23–12 Ma (Best et al., 1987), and to the east, the Mount Belknap Volcanics of the Marysvale field, which are 22–14 Ma (Rowley et al., 1994a, 1998). Still younger rhyolites range from 6 Ma to Pleistocene.

Middle Cenozoic extensional faults in the Pioche-Marysvale igneous belt are sparse and widely distributed and appear to be the same age as the intrusive rocks. In the North Pahroc Range (Fig. 6) and the western part of the belt, two angular unconformities (Scott et al., 1995c) are constrained in age between outflow ash-flow tuffs of the 30 Ma Wah Wah Springs Formation and 28 Ma Lund Formation of the Needles Range Group (Best et al., 1989a) and between the Lund and 27 Ma Isom Formations (Best et al., 1989a). Scott et al. (1995c) suggested that development of the unconformities followed deformation on north-striking faults. In the Fairview Range (Fig. 6) to the northeast, Bartley (1989) noted several angular unconformities of 29–27 Ma, as also constrained by ash-flow tuffs. In the southern Marysvale volcanic field, just west of the town of Antimony (Fig. 6), an angular unconformity formed after the 26 Ma Kingston Canyon Tuff Member (Rowley et al., 1994a) of the Mount Dutton Formation was tilted along a north-striking fault, after which the 23 Ma Osiris Tuff (Rowley et al., 1994a) was deposited (Rowley, 1968).

Delamar–Iron Springs belt

The Delamar–Iron Springs igneous belt, which is slightly younger than the Pioche-Marysvale belt, extends from the aeromagnetic quiet zone east-northeastward to the southern Marysvale volcanic field (Figs. 3 and 6). It was named by Shawe and Stewart (1976), who considered it to be a subbelt to the Pioche-Marysvale belt and extended it only as far east as the Iron Springs mining district, at and near Iron Springs (Fig. 6). Laccoliths in the Iron Springs district (Mackin, 1960; Rowley and Barker, 1978) are part of a string of intrusions that make up the northeast-trending Iron axis, from the southern Bull Valley Mountains to the Markagunt Plateau (Fig. 6; Blank et al., 1992). These intrusions are the upper part of a batholith that extends from the Iron axis west to at least the central Escalante Desert (Williams, 1998) and east across the northern Markagunt Plateau (Fig. 6; Blank and Kucks, 1989; Blank et al., 1992; Rowley et al., 1998). Most intrusions in the Iron axis and Markagunt Plateau intruded to less than 1 km of the paleosurface. Thus, the Delamar–Iron Springs igneous belt is comparable in scale to the Pioche-Marysvale belt. It extends at least as far east as the Spry intrusion (Anderson and Rowley, 1975), which is a large batholith in the subsurface, as indicated by a large gravity low near and south of Spry (Fig. 6; Rowley et al., 1998). The igneous belt is not as well delineated by aeromagnetic data as the Pioche-Marysvale belt, perhaps because the middle parts of it underlie thick basin-fill sediments in the Escalante Desert and other valleys west of Cedar City (Fig. 6).

As with the Pioche-Marysvale belt, the Delamar–Iron Springs belt is underlain by a forked aeromagnetic ridge in its western part (Fig. 5); the northwestern fork is underlain by the

Caliente caldera complex and the southwestern fork is underlain by the generally younger Kane Springs Wash caldera complex. The Caliente caldera complex consists of numerous inset calderas that range in age from at least 23 to 13 Ma (Rowley et al., 1992, 1995), unusually long lived for a caldera complex. Although it has long been recognized that many calderas in the Great Basin are elongated east-west due to subsequent extension (e.g., Best et al., 1989b), the Caliente complex is highly elongated (80 km east-west vs. 35 km north-south) because recurring faulting and magmatism extended it parallel to the middle Cenozoic extension direction. It also differs from many other caldera complexes in that it is bound on its north and south sides by the Timpahute and Helene transverse zones, respectively. Thus, it is a volcano-tectonic trough in the sense of Burke and McKee (1979) and a syntectonic caldera in the sense of Rowley and Anderson (1996).

The part of the Delamar–Iron Springs belt that includes the Caliente caldera complex and areas to the east was studied paleomagnetically by Hudson et al. (1993, this volume), who called it the Caliente-Enterprise zone and interpreted it as a broad transfer zone. This and adjacent areas north and south of the caldera complex are cut by north-northwest–striking faults, some shown by Ekren et al. (1977) and many others shown by my mapping; most of these faults have oblique-slip vectors. Hudson et al. (this volume) documented locally profound, counterclockwise, vertical-axis rotation of rocks as young as 14 Ma along the north-northwest–striking faults in and near the caldera complex and along west-northwest–striking faults east of the complex. They suggested that the vertical-axis rotation resulted from a wide zone of horizontal ductile shear between the two transverse zones that was enhanced by synchronous magmatism and accommodated by the north-northwest–striking oblique-slip faults. Although I am a coauthor of the Hudson et al. report (this volume), I instead conclude that the north-northwest–striking faults resulted from a regional stress field involving a north-south σ_1 and an east-west σ_3. Axen et al. (1993) suggested that the Timpahute transverse zone is a tear fault that bounds the southern side of a terrane underlain by the east-dipping, low-angle Stampede fault, which they interpret to be prevolcanic and Tertiary. Although I formerly interpreted the Stampede fault similarly (Rowley et al., 1992, 1994b), it more closely resembles Mesozoic attenuation faults mapped by Nutt and Thorman (1992, 1994) and Nutt et al. (1994), and, if so, it long predates and would have little bearing, except as a crustal flaw, on the siting of the Timpahute zone or Caliente caldera complex. Axen (1994) hypothesized that an east-northeast–trending area between the northern Sheep Range (north of Las Vegas, Fig. 1) and Cedar City is an accommodation zone. In contrast to Hudson and Axen, I think of the Caliente complex and areas to the east as an east-trending segment of the Delamar–Iron Springs belt bounded by two east-striking transverse zones.

The northerly striking middle Cenozoic faults identified mostly in the central to western parts of the Delamar–Iron Springs belt are generally coeval with adjacent intrusions and volcanic source areas, and both faulting and magmatism are younger southward in the belt. Faults are sparse, however, in the batholiths of the Iron axis and southern Marysvale field. The Woolsey Ranch fault in the Iron axis, however, is a northwest-striking left-lateral transfer fault (Blank and Mackin, 1967) that allowed spreading of the east flank of Iron Mountain laccolith during its emplacement at about 22 Ma (Fig. 6). Gravity slides were shed during rapid emplacement of the shallow intrusive domes of the Iron axis (Blank and Mackin, 1967; Blank et al., 1992; Hacker, 1995). Local high-angle faults in the Iron axis as well as coarse syn- and post-intrusion, middle Miocene fanglomerates also recorded uplift and deformation at the same time as magmatism (Mackin, 1960; Cook, 1960; Siders and Shubat, 1986). In the northern Markagunt Plateau, different types of low-angle faults may be temporally and genetically related to a batholith complex (Davis and Rowley, 1993), including (1) a pre-22 Ma gravity slide that apparently produced the Markagunt megabreccia (Anderson, 1993); (2) the low-angle, 22.5–20 Ma, extensional Red Hills shear zone (Maldonado, 1995); and (3) 30–20 Ma thin-skinned thrust faults (Lundin, 1989; Merle et al., 1993). Faults synchronous with calc-alkaline intrusions, however, are generally sparse in the eastern Delamar–Iron Springs igneous belt. Similar to the Marysvale volcanic field, this may be related to emplacement of shallow intrusions, which relieved most of the middle Cenozoic deviatoric stress.

In the western part of the igneous belt, middle Cenozoic faults that accompany both calc-alkaline and bimodal magmatism are abundant. This is likely due in part to the relatively deep level of intracaldera intrusions. In and near the Caliente complex, the north-northwest–striking high-angle oblique (right lateral and normal) faults are abundant, and conjugate north-northeast–striking high-angle oblique (left lateral and normal) faults are present locally. In addition, my mapping shows that the east-striking faults of the transverse zones were active at the same time as the oblique faults, as was the north-striking west-verging 15 Ma Highland Peak detachment fault (Fig. 6; Axen et al., 1988; Rowley et al., 1992, 1994b). All these faults formed during the main extensional episode in this part of the Great Basin when σ_3 was oriented east-west. This episode is well constrained by dikes, such as the porphyry of Meadow Valley Wash, that intruded many of the major faults. About a dozen dikes have been dated as 21–16 Ma by $^{40}Ar/^{39}Ar$ geochronology (e.g., Rowley et al., 1992). In addition, mapping of growth-type faults (Bowman, 1985; Michel-Noël et al., 1990) shows that some faults remained active for 1 m.y. or more, after which offset was transferred to a parallel fault of the same geometry and slip. Upward fanning of intracaldera strata, in which younger rocks dip progressively less, is common. Detailed mapping and dating constrain the main extensional episode as 21–12 Ma in most areas. Locally, however, extension probably began as early as 25 Ma (Rowley et al., 1992, 1995). The maximum rate of tilting took place at about 20–17 Ma.

In the Kane Springs Wash caldera complex (17–14 Ma), detailed mapping and dating demonstrate similar but somewhat younger faults. On the basis of dips of strata, the main episode of extension began at 17 Ma, peaked at 15–13.5 Ma, and continued to about 12 Ma (Scott et al., 1995a, 1995b, 1996). In the areas of both the Caliente and Kane Springs Wash caldera complexes, north-striking basin-range faults blocked out some ranges in a separate, younger episode, beginning after 12 Ma. The southward migration of middle Cenozoic pre-basin range faulting and synchronous magmatism spanned the transition from calc-alkaline to bimodal magmatism, which in these areas was at about 17 Ma. In the Caliente caldera complex, pre-17 Ma tuffs are calc-alkaline, mostly low-silica rhyolites, whereas post-17 Ma tuffs are bimodal, high-silica rhyolites. In the Kane Springs Wash caldera complex, all rocks are bimodal, consisting of high-silica rhyolite and trachyte tuffs and basalt flows.

Tuscarora belt

The northernmost igneous belt in the Great Basin is here called the "Tuscarora igneous belt," following Christiansen and Yeats (1992, Fig. 19), who named it the Tuscarora magmatic belt and showed it as a broad west-northwest-trending swath extending as far south as Ely (Figs. 3, 9). Igneous centers in the belt are small and scattered (Brooks et al., 1995) and andesitic lava flows are more potassic eastward, with dacite and rhyolite prominent in later phases and in more eastern parts (Christiansen and Yeats, 1992). Because of its small, separated vent areas, the belt does not show up well as aeromagnetic anomalies (Zietz et al., 1976, 1978), except for the narrow (20 km), 100-km-long east-northeast–trending belt of anomalies of the Oquirrh-Uinta transverse zone which contains large igneous centers. The Tuscarora belt was called the Bingham-Park City uplift by Butler et al. (1920) and Oquirrh-Uinta mineral belt by Hilpert and Roberts (1964) and Shawe and Stewart (1976). Christiansen and Yeats (1992) assigned the igneous rocks of the Tuscarora belt to an age of 43–37 Ma, which postdates the Challis igneous belt. Brooks et al. (1995) determined $^{40}Ar/^{39}Ar$ dates on the belt north of about lat 40° N, which they called the northeast Nevada volcanic field, and constrained its age as 43–39 Ma. The northern Deep Creek Range probably is part of the belt because igneous rocks there have K-Ar dates of 44–37 Ma (Moore and McKee, 1983) and an $^{40}Ar/^{39}Ar$ date of 39.2 Ma (Brooks et al., 1995). Part of the southern Deep Creek Range, where Brooks et al. (1995) dated volcanic rocks as 39.5 Ma, may also be included. In Nevada, Roberts (1966) showed a poorly defined east-northeast–trending "mineral belt," the Cortez-Uinta axis, that crosses the quiet zone and connects most of Nevada's iron deposits; I include it in the Tuscarora igneous belt. The Tuscarora igneous belt also includes metamorphic core complexes, which individually have complex histories that include early and middle Cenozoic and Cretaceous igneous rocks and cooling ages (Saltzer and Hodges, 1988; Wells et al., 1990; McGrew and Snee, 1994).

Magmatism in the Tuscarora belt took place during "major tectonic extension in at least part of the belt" (Christiansen and Yeats, 1992). Miller (1991) noted latest Eocene extensional faults in the Pilot and Deep Creek Ranges (Fig. 9) on the Utah-Nevada border and probable extensional faulting in other nearby areas. There is a middle Eocene angular unconformity between the volcanic rocks (39 Ma at their base) of the belt and the underlying lacustrine sedimentary rocks of the lower Eocene Elko and White Sage Formations (Brooks et al., 1995; Potter et al., 1995). These sedimentary rocks contain volcanic material (Brooks et al., 1995), so deformation appears to be synchronous with early volcanism in the belt.

Ely-Tintic belt

Butler et al. (1920) coined the name Deep Creek–Tintic uplift, and Hilpert and Roberts (1964) used the name Deep Creek–Tintic mineral belt for a west-northwest–trending belt of mining districts in Utah that crosses into Nevada centered at lat 40° N (Figs. 3 and 9). Shawe and Stewart (1976), however, showed the Deep Creek–Tintic belt trending just north of east. Its counterpart in Nevada was called the Cherry Creek mineral belt (Roberts, 1964, 1966). Aeromagnetic anomalies (Zietz et al., 1976) show that the Deep Creek–Tintic belt is an east-west– to east-northeast–trending belt in Utah (Stewart et al., 1977; Bagby, 1989). The belt is not as well defined by aeromagnetic anomalies in Nevada (Zietz et al., 1978), and thus Stewart et al. (1977) and Bagby (1989) showed it considerably wider (north-south) so as to include the Ely mining district (Fig. 9). With accumulation of additional aeromagnetic data and better processing of these data (Hildenbrand and Kucks, 1988a, 1988b), the belt is better constrained. I call it the Ely-Tintic igneous belt (Figs. 3 and 9).

Stoeser (1993) showed an east-northeast–trending string of calderas and other igneous rocks underlying the eastern 100 km of the Ely-Tintic igneous belt. From the Thomas caldera on the west to the Tintic caldera on the east (Fig. 9), these areas are marked by the most prominent short-wavelength aeromagnetic anomalies in the igneous belt and consist of calc-alkaline rocks having $^{40}Ar/^{39}Ar$ dates of 37–34 Ma (Stoeser, 1993). High-silica rhyolites of the bimodal sequence of about 21–16 Ma (Luedke, 1993; Stoeser, 1993) are superimposed on the calc-alkaline rocks, as in the Pioche-Marysvale igneous belt. Gans et al. (1989) mapped calc-alkaline igneous rocks in the western Ely-Tintic igneous belt and found that most vents and intrusions have K-Ar ages of 36–33 Ma. In the northern edge of their map area, which I interpret to be the southwestern edge of the Tuscarora igneous belt, they compiled K-Ar ages of 40–35 Ma. Gans et al. (1989) concluded that extensional faulting began at least by 36 Ma and continued during and after magmatism began. In the eastern part of the Ely-Tintic belt, field relationships do not well constrain the age of faulting, although pre-basin-range faults were considered likely by Stoeser (1993).

Western Great Basin

Cenozoic igneous belts west of the northern Nevada rift are poorly defined due to complex geology and, in some areas, cover by widespread young basalts. In the northwestern Great Basin (Washington, Oregon, northeastern California, and western Nevada), tectonostratigraphic terranes (John et al., 1993, Fig. 5) include the Cascades magmatic arc (Fig. 1) and, on its east, basalts of its back arc (Christiansen and Yeats, 1992). Calc-alkaline rocks in the southwestern Great Basin (mostly south of lat 40° N), however, are similar to those discussed elsewhere in this chapter except that large volcanic centers, notably calderas, are much more highly concentrated than elsewhere (Figs. 3 and 8). The calderas show up as a belt of isostatic residual gravity lows (Fig. 4; Saltus, 1988b; Jachens and Moring, 1990; Saltus and Jachens, 1995). Most of these rocks are 30–20 Ma and form a convex-southward arc (Stewart and Carlson, 1976; Stewart et al., 1977; Best et al., 1989b). The eastern arm of this arc crosses the aeromagnetic quiet zone and swings into Utah as the Pioche-Marysvale igneous belt, which also exhibits profound isostatic residual gravity lows (Fig. 4; Saltus and Jachens, 1995; V. Bankey, 1993, written commun.). Sullivan et al. (1991) projected this belt farther eastward through the laccoliths of the Colorado Plateau to the San Juan volcanic field (Fig. 1) and called it the Reno–San Juan magmatic zone.

In the Walker Lane belt, aeromagnetic anomalies that mostly reflect Mesozoic and Tertiary calc-alkaline intrusions are more abundant and widespread than in the eastern Great Basin, but they are arranged mostly in short, narrow, northwest-trending belts of positive anomalies separated by equally short and narrow lows. In contrast, the anomalies in the eastern Great Basin consist of small-wavelength anomalies from Tertiary intrusions separated by broad areas of mostly long-wavelength negative anomalies (Fig. 5). The northwest-trending anomalies in the Walker Lane belt may reflect dismemberment of igneous belts by strike-slip faults; therefore, the anomalies do not define igneous belts as in the eastern Great Basin. That the Walker Lane belt had a middle Cenozoic history of deformation that was synchronous with magmatism (Ekren et al., 1980; Hardyman and Oldow, 1991) complicates the interpretation. Anomalies extending east from the Walker Lane through and east of Tonopah, however, are the closest to those in igneous belts of the eastern Great Basin. These anomalies may represent a relatively undeformed igneous belt. Nonetheless, because of its complex geologic history, I group the igneous rocks of the western Great Basin into one large unnamed mass (Figs. 3 and 8). This mass probably includes the western ends of the several belts that extend eastward into Utah. The similarity of igneous rocks on both sides of the Great Basin is indicated by their similar ages, which are also generally younger southward (Fig. 7). As in the eastern Great Basin, bimodal high-silica rhyolites are also in the south.

Pre-basin-range faults in the Kawich Range (Fig. 8) and northern Belted Range (Fig. 1) are represented by northwest- and northeast-striking normal and oblique faults of 27–18 Ma (Ekren et al., 1968). Strike-slip faulting in the Walker Lane belt began about 26–25 Ma (Stewart, 1988). John et al. (1989) and Hardyman and Oldow (1991) documented 24–19 Ma normal and strike-slip faulting and mineralization in the central Walker Lane belt that continued until at least 15 Ma (John and Hudson, 1990). The culmination of extension in the Yerington mining district (Fig. 8) resulted in 17–11 Ma, north-striking normal faults (Proffett, 1977; Hardyman and Oldow, 1991). Dilles and Gans (1995) refined the timing on these normal faults as between 14 and 12.5 Ma for the main phase of extension in the area.

The tie between pre-basin-range faults and synchronous calc-alkaline and bimodal magmatism in the western Great Basin is less certain than the evidence for older extension. In outflow tuffs between the two volcano-tectonic troughs of Burke and McKee (1979), transtensional deformation and vertical-axis rotation were concurrent with deposition of ash-flow tuffs of 33–23 Ma (Hudson and Geissman, 1991). In the Walker Lane, northwest-striking strike-slip faults were synchronous with, and localized sources for, calc-alkaline tuffs and flows of 27–22 Ma (Ekren et al., 1980). John et al. (1989) and John and Hudson (1990) concluded that 22–15 Ma extensional faults in the Paradise Range (Fig. 8) bordering the central Walker Lane belt coincided with the change from calc-alkaline volcanism to bimodal volcanism and associated mineral deposits. Seedorff (1991) found southward-sweeping magmatism and extensional faulting throughout the Great Basin, including its western part, but he concluded that calc-alkaline magmatism generally preceded pre-basin-range faulting and that most bimodal rocks generally preceded main phase basin-range extension, except for later basalts that coincided with waning extension. Later work in and east of the central Walker Lane belt indicated a close association between magmatism and extensional faults, both of which swept southwestward at 27–22 Ma (Dilles and Gans, 1995). West of the Walker Lane belt, 17–11 Ma normal faults formed as calc-alkaline magmatism was ceasing in the area (Proffett, 1977). The age of this faulting was better constrained by Dilles and Gans (1995) as 14–12.5 Ma, and they found a close association between magmatism and extensional faults. In the Nevada Test Site and adjacent areas, high-silica rhyolite tuffs of the bimodal episode (15–7.5 Ma) coincided with peak extension in surrounding areas, although locally coincident faults and caldera eruptions are not common (Sawyer et al., 1994; Minor, 1995).

TRANSVERSE ZONES AND IGNEOUS BELTS OUTSIDE THE GREAT BASIN

Transverse zones are not unique to the Great Basin, and are probably important parts of the geometry of all extended terrains. However, some transverse zones are in areas that are not highly extended, and others may have formed in contractional regimes parallel to the axis of σ_1. An example of the latter is the Colorado mineral belt (summarized by Cunningham et al., 1994). The Colorado mineral belt strikes northeast for about 500 km, from extreme southwestern Colorado to north of Denver (Fig. 1). It contains most of the major mining districts in Colorado, which are associated with intrusive rocks that define the mineral belt (Cunningham et al., 1994). The strike of the Colorado mineral belt is parallel to the subduction direction during the Laramide deformational event (Atwater, 1970; Lipman, 1980, Fig. 14.7; Severinghaus and Atwater, 1990, Fig. 8), and thus may be genetically linked. The belt, however, also follows Precambrian structural trends (Tweto and Sims, 1963) and has been proposed to be part of a more extensive Precambrian feature called the Colorado lineament (Warner, 1978; Stewart and Crowell, 1992). Cunningham et al. (1994) showed that most of the intrusive rocks and mineral deposits are part of the Laramide deformational event and that virtually all igneous rocks of this age in the general area are along the belt. They also concluded that, contrary to previous work on the belt, some intrusions and mineral deposits along the belt are middle and late Tertiary and include older calc-alkaline and younger bimodal (granitic) types. Tertiary intrusive activity youngs northeastward along the mineral belt at a rate of about 4 cm/yr (Cunningham et al., 1994), a trait common to some igneous belts in the eastern Great Basin. Bimodal intrusions and associated mineral deposits, which are older than in the Great Basin, are present where the mineral belt crosses the Rio Grande rift (Fig. 1) and appear to be related to extension along the rift (Cunningham et al., 1994). Cunningham et al. (1994) recognized the plate tectonic significance of the belt and suggested that Laramide activity along the belt represents "a leaky transform fault in the subducted slab." The feature is probably a transverse zone that was initiated during Laramide event contraction and was reactivated during middle and late Tertiary extension.

Eaton et al. (1978), Mabey et al. (1978), Eaton (1979), Lipman (1980, 1992), and Smith and Luedke (1984) emphasized that at least two other northeast-striking belts, the Humboldt zone (also called the Snake River–Yellowstone or Snake River belt) and the Jemez (or Springerville-Raton) zone (Fig. 1), formed at the same time as the Colorado mineral belt and are transform faults. Suppe et al. (1975) called these two belts "aseismic volcanic chains" and interpreted them as hotspot traces. The similar northeast strike of these belts suggests that either they began to form during the Laramide event or they followed structures of the event. The Humboldt zone was proposed by Eaton et al. (1975, 1978), Mabey et al. (1978), Christiansen and McKee (1978), and Rowan and Wetlaufer (1981) to extend about 1,000 km from central Nevada, through the eastern Snake River Plain, Yellowstone National Park, and into Montana, and to be controlled by Precambrian crustal flaws of the same strike (Rowan and Wetlaufer, 1981). In Nevada, the Humboldt zone is characterized by high heat flow (Lachenbruch and Sass, 1978; Blakely, 1988). On the basis of teleseismic P-wave velocities in the upper mantle, the zone is characterized by low velocities at great depth and thus is a deep structure (Humphreys and Dueker, 1994). Eaton et al. (1975) and Mabey et al. (1978) proposed that the zone, and its late Cenozoic volcanism in the eastern Snake River Plain and Yellowstone Park, were controlled by two "lineaments" of transform-type kinematics. Pierce and Morgan (1992, p. 32) disputed this idea and ascribed the origin of the eastern Snake River Plain to a migrating hotspot due to a mantle plume. Zoback et al. (1994) and others endorsed this hypothesis. Lipman (1980, 1992), however, disputed it largely on the grounds that magmatism along and south of the Brothers fault zone, in stark contrast, migrated toward the west-northwest. I agree with Eaton, Mabey, Lipman, and their coworkers and propose that the eastern Snake River Plain developed along two transverse zones formed parallel to the extension direction. The upper Cenozoic volcanic rocks controlled by the transverse zones were erupted during a period of northeast-trending extension (Kuntz et al., 1992; Pierce and Morgan, 1992), in contrast to the generally east-west extension direction in the Great Basin.

The Jemez "zone" (Jemez lineament of Chapin et al., 1978) is a largely aseismic feature marked by a northeast-trending series of upper Cenozoic volcanic centers, including the Springerville and Mount Taylor volcanic fields and the Valles caldera, extending from southwestern Arizona to northeastern New Mexico and adjacent Colorado (Fig. 1; Chapin et al., 1978; Lipman, 1980; Smith and Luedke, 1984). No age progression exists along the zone, and it follows Precambrian and Laramide trends of the same strike (Lipman, 1980; Cather, 1992). As with the Humboldt zone, it has been imaged as a low-velocity zone in the upper mantle (Humphreys and Dueker, 1994). Where the Jemez zone crosses the late Cenozoic, north-striking Rio Grande rift south of Taos (Fig. 1), it includes a northeast-striking fault, the Embudo fault, of complex displacement that may offset the rift in a right-lateral sense. Muehlberger (1979) discussed the Embudo fault and interpreted it to be a transform fault. I interpret the Jemez zone to be a transverse zone.

Two other northeast-striking transverse zones, the Morenci "lineament" at Socorro and the Santa Rita "lineament" at Truth or Consequences, cross the Rio Grande rift (Fig. 1; Chapin et al., 1978). Both zones localize plutons and mineral deposits of the Laramide event and are oriented parallel to Laramide and Precambrian structures. The Morenci lineament is defined by loci of calderas younger than 32 Ma and separates tilt-block domains of opposite polarity. Although the Morenci lineament is associated with a right step in the Rio Grande rift, its true sense of shearing may be left lateral. Chapin et al. (1978) suggested that it is an "incipient transform fault" that followed Precambrian lines of weakness, which may explain why these transverse zones are oriented at 45°–60° to normal faults of the Rio Grande rift rather than at 90°. Chapin (1989) and Chapin and Cather (1994) renamed part of the Morenci lineament the Socorro accommodation zone. They recorded no strike-slip motion along it and concluded that it is aseismic and is not even a conspicuous shear zone at the surface. Other parallel transverse zones also cross the Rio Grande rift (Chapin, 1989; Cather, 1992; Chapin and Cather, 1994; Lewis and Baldridge, 1994; May et al., 1994; Russell and Snelson, 1994). Cather (1992), Machette and Hawley (1996), and J. W. Hawley (1996, oral commun.) concluded that most of the Rio Grande rift is sliced by Cenozoic transverse zones and that their strikes are inherited from Laramide structures.

Other rifts contain transverse zones that are recognized to be of a "transform" nature. Dickerson and Muehlberger (1994) and Henry (this volume) noted west-striking transverse zones that separate north-northwest–trending rift basins in the Big Bend area of western Texas. The zones generally predate the oldest (23

Ma) basins but were locally active during basin formation. Basalt feeders and hot springs are present locally at intersections between transverse zones and basin-forming faults. Barberi and Varet (1977) described northeast-striking "transverse volcanic structures" in the Afar depression of East Africa. Rosendahl (1987) and D. L. Scott et al. (1992), among others, described northwest-, north-, and northeast-striking "transfer" or "accommodation" zones in the East African rift. Patton et al. (1994) described "transfer zones" in the Gulf of Suez rift. "Transfer" and "accommodation" zones have also been described from passive continental margins and in oceanic lithosphere. For example, Gibbs (1984) described transfer faults from the North Sea extensional basin. Etheridge et al. (1985) and Scott and Second (1994) described Lower Cretaceous "transfer" and "transverse" faults along the submarine passive margins of Australia. Along the Mid-Atlantic Ridge, east-striking Quaternary transfer zones localize undersea basalt and hydrothermal vents (Karson and Rona, 1990). A submarine back-arc basin south of Japan is segmented by east-striking transfer zones that are identified not by discrete faults of the same strike, but instead by termination of, or changes in strike and relative offset of, rift-parallel (generally north-striking) normal faults and by volcanic vents (Taylor et al., 1991). In a study of mid-ocean ridges in general, Mutter and Karson (1992) interpreted that an orthogonal system of normal faults and transfer zones characterize slow-spreading ridges, whereas magmatic processes (nonmechanical deformation) characterize fast-spreading ridges. Scale-model experiments of simulated normal rifting (Serra and Nelson, 1988) and oblique rifting (Mark, 1994) resulted in "transfer zones."

ECONOMIC IMPLICATIONS

Formation of metallic mineral deposits is the most obvious economic implication of transverse zones and igneous belts. Metallic mineral deposits are abundant in the Great Basin. Those in the western half are especially numerous and contain greater volumes of precious metals than those in the eastern half due to their different underlying stratigraphy and structural terranes (Shawe and Stewart, 1976; Bagby, 1989). Mining districts in the eastern Great Basin are aligned along mostly east-northeast–striking igneous belts and transverse zones. Mining districts in the western Great Basin have a nearly random distribution, although Roberts (1964, 1966), Shawe and Stewart (1976), and Bagby (1989) suggested a preferred northwest trend. Regardless of what trends exist, most mineral deposits in the western Great Basin, like those in the eastern part, are due to a combination of intrusions, transverse zones, and faults of both the middle and late Cenozoic episodes.

Geothermal areas are commonly associated with transverse zones and igneous belts. Some hot ground water along them is due to shallow magma, whereas other hot ground water reflects deep ground-water circulation along permeable sheared rock of transverse zones. Some geothermal areas in the Great Basin have been noted herein. In the Rio Grande rift, many thermal springs along the Socorro accommodation zone are ascribed to excellent permeability caused by the abundance and interconnectedness of fractures during faulting along the zone (Chapin et al., 1978).

Control of ground-water flow in the Great Basin is increasingly being linked to transverse zones. In Utah, the Blue Ribbon transverse zone corresponds to a ground-water divide between northward flow toward the Great Salt Lake Desert and southward flow toward the Virgin River (Prudic et al., 1995). Southward flow in the great carbonate-rock aquifer of central to southern Nevada (Dettinger et al., 1995; Prudic et al., 1995; Schmidt and Dixon, 1995) has been greatly influenced by transverse zones. For example, ground water in the White River flow system (eastern part of the carbonate aquifer; Dettinger et al., 1995) is redirected to the west along the Blue Ribbon and Timpahute transverse zones. The southward flow in the aquifer under Lake Valley (north of Pioche; Fig. 6) was modeled to appear to encounter a barrier along the axis of the Blue Ribbon transverse zone and thus to turn west between the north-trending Fairview and Bristol Ranges (M. Johnson, 1994, oral commun.; Brothers et al., 1996). The ground water flows west for 10 km, then continues south under Dry Lake Valley west of the Bristol Range (Fig. 6; Page and Ekren, 1995; Page et al., 1995). The ground water was further modeled as appearing to make a second 90° bend at the Timpahute transverse zone, where it passes 20 km westward from southern Dry Lake Valley, through the northern South Pahroc and Hiko Ranges, to beneath Pahranagat Valley (Fig. 6; M. Johnson, 1994, oral commun.; Brothers et al., 1996). Under the Pahranagat Valley, which is a remarkably well watered valley of springs and lakes, the ground water continues southward until it encounters the Pahranagat shear system (Ekren et al., 1977). Two large natural lakes, Upper and Lower Pahranagat Lakes, have formed north of the shear system due to damming created by the shear system (Prudic et al., 1995). The faults within the shear system here are plugged by basalt dikes and thus form a barrier to ground-water flow (Ekren et al., 1977; G. L. Dixon, 1994, oral commun.). Ground water of the White River flow system is diverted southwestward along the shear system before continuing southeastward beneath Coyote Spring Valley and then issuing at Muddy River springs 15 km northwest of Moapa (Fig. 3; Dettinger et al., 1995; Schmidt and Dixon, 1995) to form the Muddy River. In other parts of the carbonate aquifer, levels of ground water suggest that both the 37° discontinuity of Eaton (1975) and the Las Vegas shear zone act as barriers to southward ground-water flow (Prudic et al., 1995). In addition, the western part of the Timpahute transverse zone is both a barrier and ground-water divide (R. J. Laczniak, 1996, oral commun.).

In New Mexico, ground-water resources along transverse zones have received considerable attention. The Socorro accommodation zone contains some of the largest springs in the area and is a zone of ground-water movement (Chapin, 1989). Chapin and Cather (1994) concluded that transverse zones in the Rio Grande area act as barriers to the general southward ground-water flow and "constrict and alter ground-water flow patterns." Anderholm (1983) noted high chloride content in ground water in the southern Socorro Basin of the Rio Grande area and ascribed it to upward movement of deep-basin ground water or geothermal fluids along a transverse zone. Coons and Kelly (1984) recorded constriction of ground-water flow in the Española basin, causing an increased flow velocity, as a result of the Embudo fault zone (Muehlberger, 1979) of the Jemez transverse zone. J. W. Hawley (1996, oral commun.) considers that transverse zones form important aquifers throughout the Rio Grande rift.

Hydrocarbons are locally concentrated along transverse zones, which also influence sedimentation (see Faulds and Varga, this volume). High-quality seismic reflection surveys of the submarine extensional Bass basin and Gippsland basin between Australia and Tasmania, for example, document orthogonal sets of synchronous Lower Cretaceous faults, west-northwest–striking normal faults and north-northeast–striking transfer faults (Etheridge et al., 1985). Five oil fields produce from anticlines above a single transfer fault in the Gippsland basin that was reactivated in the late Tertiary (Etheridge et al., 1985).

Fluid circulation is a common thread for these resources. Most mineral deposits and geothermal resources are due largely to ground water that circulates in convection cells surrounding magma sources (e.g., Podwysocki et al., 1983; Cunningham et al., 1984), whereas other geothermal fluids are hot due to their relatively rapid circulation from great depths to the surface along permeable fracture zones. Whether due to enhanced permeability from faulting, barriers to ground-water flow from clay-like gouge, or juxtaposition of rocks of contrasting permeability, transverse zones are important controls of fluid migration. Furthermore, because transverse zones are long lived and commonly localize (or are localized by) intrusions, the resulting hot water passes through fractures that are continually reopening, permitting mineral deposits to grow. Thus, transverse zones are important targets for mineral deposits, geothermal energy, ground-water supplies, and perhaps hydrocarbons.

DISCUSSION

The dominant northerly-trending elements of Great Basin landscape (Fig. 2), caused by late Cenozoic basin-range faults, mask structures of other strikes. Yet upon careful inspection, even of the topography, some range fronts are seen to trend anomalously eastward and neighboring ranges commonly terminate at the same latitude. Aeromagnetic and gravity maps add significantly more substance to such initial impressions. In addition, isostatic residual gravity maps of the Great Basin, from which thicknesses of basin fill can be determined (Jachens and Moring, 1990), reveal that presumably simple, long, linear, narrow basins are segmented along strike into deep and shallow parts (Saltus and Jachens, 1995). In the discussion, I have attempted to provide field evidence for the transverse igneous belts and structural zones. Interpretation of these trends, however, is more controversial.

Transverse zones

From field relations, I interpret that most transverse zones are boundaries that separate domains of different types, amounts, and rates of extension. Some transverse zones, however, may be structural zones of finite width that control emplacement of intrusive masses and deformation of rocks within them. Other zones, such as the early Cenozoic Laramide parts of the Colorado mineral belt, may form under compression. Transverse zones, by definition, strike generally east, at high angles to the dominant middle and late Cenozoic structural grain. This implies that most large transverse zones in the Great Basin formed parallel to the extension direction. D. L. Scott et al. (1992) suggested that orthogonal sets of transfer and normal faults represent an ideally extending system. I amplify this to suggest that transverse zones are a necessary part of oblique and normal extension. Some small transverse zones, as well as many accommodation zones, however, may strike at low angles or even parallel to the grain. The orientation of some other transverse zones, such as those in the Rio Grande rift, may be due to the influence of basement structures. Furthermore, not all east-striking faults in the Great Basin are due to transverse zones; some represent minor tear faults in the hanging walls of detachment or thrust faults.

On the basis of fault-slip data and east-striking dikes and faults, Best (1988) suggested, in stark contrast to my views, that some of the Oligocene to Miocene extension in the Great Basin resulted from a horizontal least-principal stress axis (σ_3) that trended north. Bartley (1989, 1990), Gans (1990), Bartley et al. (1992, 1994), Overtoom et al. (1993), Overtoom and Bartley (1994), and Brown and Bartley (1994) made the same interpretation for some east-striking, middle Cenozoic faults that they mapped along the Blue Ribbon transverse structure. Stewart (1983), Hardyman and Oldow (1991), and Seedorff (1991) had a similar interpretation for other transverse zones and volcanotectonic troughs. Tingey et al. (1991) also advocated a middle Cenozoic north-south σ_3 on the basis of nearly east striking 24 Ma dikes in the northern High Plateaus transition zone. In contrast, I suggest that east-, east-northeast–, and east-southeast–trending structures and intrusions were controlled by transverse zones and igneous belts, rather than by a north-trending σ_3. Because I interpret that most transverse zones formed parallel to extension, I would not refer to them as rifts (Bartley et al., 1992) or spreading centers (Bartley, 1989).

Studies of seismic gaps suggest that persistent boundaries represent a type of "geometric barrier," where fault ruptures start or stop (Aki, 1979). Ruptures may terminate at transverse zones if the zones consist mostly of gouge, breccia, and severely fractured rock that has no shear strength, that deforms ductilely, and that will not transmit strain parallel to the extension direction (Aki, 1979). The system of fractures and shears in a transverse zone, being generally orthogonal to those of north-striking faults such as the Wasatch fault zone, provide a "nonconservative barrier" (King and Nabelek, 1985) to the rupture, in the same sense that this barrier is provided by a constraining bend in a strike-slip fault (King, 1983). Transverse zones may represent nonconservative fault-rupture barriers, aligned parallel to the extension direction (Thenhaus and Barnhard, 1989). In other words, a transverse zone is a soft "process zone" of intensely faulted rock that absorbs and disperses the energy of incoming rupture tips along many small faults (King and Nabelek, 1985). Mutual interlocking of small fractures of different trends results in asperities on the main fault that must be broken in order for rupture to continue through the process zone (Wheeler and Krystinik, 1988, 1992). Transverse zones commonly terminate ranges and commonly are wide and poorly exposed, suggesting that they contain large volumes of easily eroded fault-damaged rock. Magma bodies localized along a transverse zone would tend to locally decrease shear strength until they cooled and solidified. Seismic gaps may result either because such low-strength rock will not break and generate earthquakes, or the grain of such sheared rock, orthogonal to north-striking active basin-range faults, impedes shearing.

Igneous belts

I interpret east-northeast to east-southeast igneous belts in the Great Basin to have been originally bounded by transverse zones. However, most well-exposed transverse zones strike closer to east, nearly parallel to the current extension direction and at about 20° to the trend of most igneous belts. This suggests that few igneous belts are younger than middle Cenozoic, but that many transverse zones are. Thus, the young transverse zones cut across the igneous belts except where the belts also are young or where the middle Cenozoic extension direction also trended east.

Most individual igneous belts grew parallel to the extension direction, as reflected by isochrons cutting obliquely across the igneous belts (Fig. 7). Thus, igneous rocks of a single belt, including their associated faults, are generally younger westward in the western Great Basin, whereas those in belts in the eastern Great Basin are generally younger eastward. In the region south of the Brothers fault zone (Fig. 1), rhyolite eruptive centers are younger to the west-northwest, from 16 Ma at the McDermitt caldera (Zoback et al., 1994) to 10 Ma in southeastern Oregon and to Quaternary in central Oregon (MacLeod et al., 1976). The Orevada rift also is younger to the west-northwest, from 19 Ma in the McDermitt area to 7 Ma (Rytuba and Conrad, 1981) in south-central Oregon. North of the Garlock fault zone, extension generally migrated westward (Wernicke et al., 1988), from middle Miocene in the Nevada Test Site (e.g., Minor, 1995), to late Miocene in the Bullfrog Hills and Funeral Mountains of California, to Pliocene-Holocene in the Death Valley in California (Fig. 3; Hamilton, 1988b). R. A. Thompson (1994, written commun.) noted west and northwestward migrations of 8–3 Ma volcanic rocks and of post-6.5 Ma faults in the Death Valley area.

In contrast, on the eastern Snake River Plain, rhyolite calderas migrated to the east-northeast, from 16 Ma at the Nevada-Oregon-Idaho border to Quaternary at Yellowstone National Park (Fig. 1; Pierce and Morgan, 1992; Rowley et al., 1998). Steven et al. (1984) noted that igneous rocks in at least the eastern two-thirds of the Pioche-Marysvale igneous belt are generally younger (from 35 to 23 Ma) eastward. Nealey et al. (1994) summarized an eastward migration with time within Quaternary basalt fields in the western Colorado Plateau in Utah and Arizona.

A model for the origin of igneous belts is provided by the eastern Snake River Plain (Figs. 1 and 2; Rowley et al., 1998). The plain geologically is part of the Great Basin, and basin-range topography continues north of it (Eaton et al., 1978; Kuntz et al., 1992). The eastern Snake River Plain is oriented east-northeast, parallel to the extension direction in and north of the plain (Kuntz et al., 1992). Most rocks at the surface of the plain are Quaternary basalt lava flows that erupted from vents along north-northwest–striking linear fissures. Basin-range faults north of the plain also strike north-northwest, and basin-range faults south of the plain strike both north and north-northwest, also perpendicular to the extension direction (Kuntz et al., 1992, Fig. 11). It is clear that northeast crustal spreading was primarily accommodated by magmatism within the plain and by normal faulting to the north and south of the plain (Kuntz et al., 1992). I suggest that the boundaries between these two types of extension, along the northern and southern sides of the plain, were somehow controlled by northeast-striking transverse zones.

CONCLUSIONS

Geologic mapping in the Great Basin has identified complex structural zones that accommodated different amounts and types of extension on either side. Some of these zones were called transform faults, but this name is better restricted to features that separate lithospheric plates. Some zones were identified as accommodation zones or transfer faults. I apply the nongenetic term transverse zone as a general term for all transversely oriented igneous and structural zones, of which transfer zones and several types of accommodation zones are subsets.

The type examples that I use for transverse zones are the Blue Ribbon, Timpahute, and Helene zones, parts of which I mapped in detail. Parts of these transverse zones contain abundant faults of different types that strike parallel to the overall zone. These faults show strike, oblique, and normal slip. Where strike-slip displacement is indicated in transverse zones in the Great Basin, the most common relative movement is left lateral (Ekren et al., 1976), but locally both left and right slip are demonstrated along the same fault or transverse zone. The amount of strike-slip displacement along transverse zones is, however, generally small. For example, zones that cross the northern Nevada rift (mostly 17–14 Ma) show little strike-slip offset. Transform faults that separate plates in the Pacific Ocean basin also are largely left lateral (Fox and Gallo, 1989), yet they commonly exhibit huge amounts of offset. Some parts of transverse zones contain no faults that parallel the overall zone, but they commonly contain other features that may be subtle, including folds, joints, evidence of twisting, and changes in strike or offset of north-northwest– to north-northeast–striking faults. Intrusions also are a common feature of transverse zones, but whether they control the zone (Faulds et al., 1994; Faulds, 1996; Faulds and Varga, this volume) and/or the zone controls them, is not known. Some transverse zones are boundaries that separate an area of intrusions (an igneous belt) on one side from an area of faults on the other side, because shallow intrusions are an effective type of extension in the same way that faults are. Transverse zones are longer lived than typical faults. Their long history of fracturing, which opens pathways for ground water, and heat pumps along them, such as intrusions that drive the water through the pathways, locally result in metallic mineral deposits. The combination of igneous belts and transverse zones is part of what makes the Great Basin one of the world's most productive mineralized areas.

I do not understand many aspects of transverse zones. I conclude that the strike of most igneous belts and large transverse zones in the Great Basin is parallel to the extension direction that existed at the time they formed. The sum total of the products of extension is a generally orthogonal network of structures and intrusions. These observations lead to the suggestion that transverse zones and igneous belts had a major role in the spreading of brittle crust. Dip slip along low-angle faults and the brittle-ductile transition requires lateral (strike-slip) transfer along high-angle faults in the same way that strike slip along high-angle faults requires movement along horizontal or low-angle structures. Because transverse zones and igneous belts are major vertical structures, they are interpreted to act in concert with the brittle-ductile transition to allow spreading.

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