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Cenozoic Igneous and Tectonic Setting of the Marysvale Volcanic Field and Its Relation to Other Igneous Centers in Utah and Nevada

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ABSTRACT

The Marysvale volcanic field of southwestern Utah, largely in the High Plateaus transition zone of the Colorado Plateau, lies at the east-northeastern end of the Pioche-Marysvale igneous belt. This belt consists of exposures of mostly Cenozoic volcanic rocks and is underlain by a batholith complex of even greater volume. The volume of volcanic rocks in the Marysvale field totals at least 12,000 km^3. The field consists mostly of middle Cenozoic, intermediate-composition, fundamentally calc-alkaline rocks erupted at 34(?)–22 Ma; associated mineral deposits are mostly of chalcophile elements. Stratovolcano deposits, especially volcanic mudflow breccia and lava flows, dominate; ash-flow tuffs derived from several calderas make up less than 10 percent of the volume of the volcanic rocks. Most of the central and northern part of the field consists of the Bullion Canyon Volcanics of relatively crystal-rich dacite and andesite; source cupolas reached shallow levels, and many intrusions have associated mineral deposits. The southern part of the field is dominated by crystal-poor andesite of the Mount Dutton Formation; postulated source cupolas are deep and unexposed, and associated mineral deposits are sparse. Shallow laccoliths that are unrelated to and south of the deep Mount Dutton sources were emplaced into lower Tertiary sedimentary rocks at the same time. These laccoliths are comagmatic with laccoliths and stocks of the "Iron Axis," in the Great Basin southwest of the Marysvale field. Large underlying batholiths fed both the southern Marysvale field and Iron Axis laccoliths.

The calc-alkaline igneous rocks of the Pioche-Marysvale igneous belt and the slightly younger Delamar–Iron Springs igneous belt to the south are part of a generally middle Cenozoic igneous sequence that spans much of the Western United States. The igneous rocks probably originated by oblique convergence during subduction of oceanic lithosphere beneath western North America. The overall area underlain largely by the igneous rocks is anomalously wide when compared with igneous areas in other parts of the Pacific rim, apparently because the subducted slab continued at a shallow depth as far to the east as the longitude of the Rocky Mountains. The Pioche-Marysvale and Delamar–Iron Springs igneous belts formed under extension parallel to the subduction direction and were bounded by transverse structures ("lineaments") of the same trend that separate areas of different amounts and types of extension. The generally east-northeast-trending crustal extension of the same age as calc-alkaline magmatism was especially profound in the Basin and Range province, but extensional faults were dominant only in some areas, whereas elsewhere extension was accomplished by passive emplacement of shallow intrusions.

The middle Cenozoic calc-alkaline rocks at Marysvale are overlain by an upper Cenozoic, fundamentally bimodal (rhyolitic and basaltic) volcanic association of rocks ranging in age from at least 23 Ma to Quaternary and intertongued in part with coeval clastic basin-fill sedimentary rocks. The bimodal association is dominated by high-silica rhyolite ash-flow tuff and volcanic domes, potassium-rich mafic rocks, and basalt cinder cones. Rocks of the bimodal association are much less voluminous (about 5 percent of the total volume of the Marysvale field) than the older calc-alkaline rocks, but they host many metallic mineral deposits, mostly of gold, silver, and lithophile elements. Bimodal rocks are also irregularly distributed elsewhere in the Pioche-Marysvale and Delamar–Iron Springs igneous belts. The late Cenozoic volcanism and extension began after subduction had started to diminish, as transform motion became significant on the San Andreas transform fault zone. Regional oblique extension, referred to as the basin-range episode and involving mostly faults under a normal-fault stress regime, was oriented generally east-west, and it began in the Marysvale area and the southeastern Great Basin at about 10 Ma, as faults formed the present topography. Transverse structures continued to be active in the Great Basin but most were oriented east-west, parallel to the new extension direction. The eastern Snake River Plain is a youthful model for volcanism and extension along probable underlying transverse structures oriented parallel to its current northeastern extension direction.

The overall effect of extension and volcanism in the Great Basin during the middle and late Cenozoic has been one of east-west spreading (widening) of the Great Basin, creating a bilateral symmetry. Any single axis of spreading is unlikely; probably there were many north-northwest- to north-northeast-trending axes perpendicular to the then-active extension direction. Extension related to these axes was expressed by faulting, igneous intrusions, or both. Under each axis, heat flow increased and the brittle-ductile transition zone rose. Extensional stress migrated with time, parallel to the extension direction, toward the cooler, more brittle margins of the Great Basin. The transverse structures, oriented parallel to the extension direction of the time, accommodated differing amounts of extension and magmatism north and south of them; they are fundamental features probably passing down to the brittle-ductile transition.

INTRODUCTION

The Marysvale volcanic field of southwestern Utah is one of the largest Cenozoic volcanic fields in the Western United States; it largely straddles the High Plateaus Section of the Colorado Plateaus province, but the western part of the field is in the Great Basin Section of the Basin and Range province (fig. 1). The High Plateaus is geologically a transition zone between the Great Basin and the main, much less deformed part of the Colorado Plateau. The volcanic field overlies Paleozoic to lower Tertiary sedimentary rocks mostly east of the northeast-trending Cordilleran hingeline;

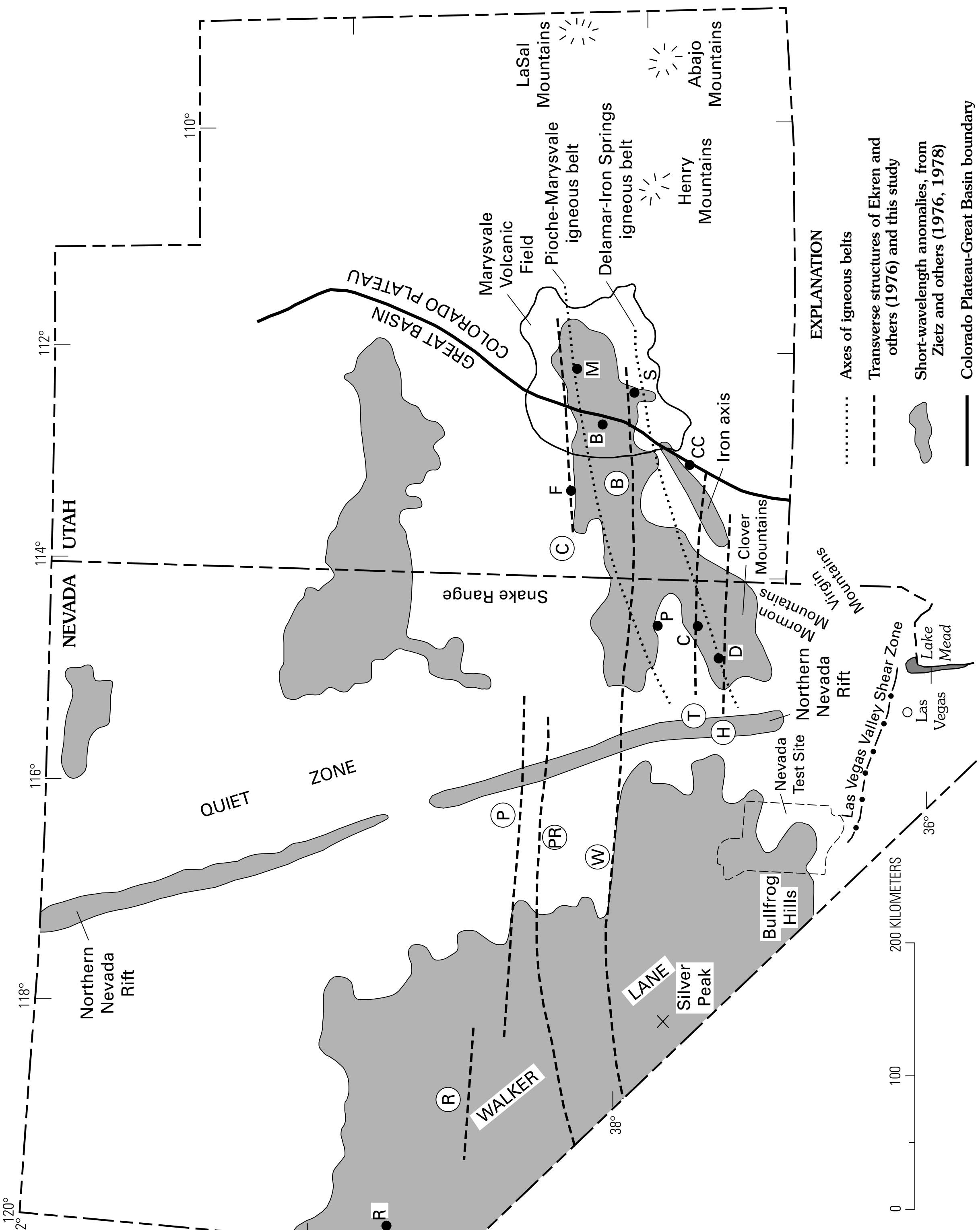

Figure 1. Major igneous and tectonic features of Nevada and Utah. Some additional igneous belts not identified here are shown in Rowley and others (1978, fig. 2). Colorado Plateau–Basin and Range boundary from Fenneman (1931). R, Reno; C, Caliente; P, Pioche; D, Delamar Mountains, F, Frisco; M, Marysvale; B, Beaver; S, Spry; CC, Cedar City. Transverse structures: R, ; P, ; PR, ; B, Blue Ribbon; W, Warm Springs; C, Cove Fort; T, Timpahute; H, Helene.

Sevier compressional deformation formed southeast-directed thrust sheets that underlie at least the northern part of the field, and apparent Laramide upwarps shed coarse clastic material that is now incorporated in lower Tertiary rocks north of the field (R.E. Anderson and Barnhard, 1992).

The volcanic field lies at the eastern end of the east-north-east-trending Pioche-Marysvale igneous belt, the site of extensive, mostly Oligocene and Miocene, volcanism and mineralization above a major batholith complex (fig. 1). Tertiary rocks in the Marysvale field consist of three main sequences: (1) lower Cenozoic (Paleocene to lower Oligocene) fluvial-lacustrine sedimentary rocks (mostly Claron Formation), which overlie in angular unconformity Paleozoic and Mesozoic sedimentary rocks; (2) thick middle Cenozoic (Oligocene and lower Miocene, 34(?) to 22 Ma) intermediate-composition calc-alkaline igneous rocks; and (3) upper Cenozoic (23 Ma to Quaternary), compositionally bimodal volcanic rocks and source intrusions (consisting of basaltic rocks and high-silica rhyolite), which intertongue with and overlie clastic basin-fill sedimentary rocks. Sources of the middle Cenozoic calc-alkaline rocks were mostly stratovolcanoes, shield volcanoes, volcanic domes, and calderas. Some pyroclastic eruptions of the upper Cenozoic rhyolite magma formed calderas; other eruptions formed lava flows and volcanic domes. The upper Cenozoic basaltic lava flows came from fissure eruptions and central vents marked by cinder cones. A total volume of at least 12,000 km^3 of volcanic rocks erupted in the field, mostly between 26 and 23 Ma. The main pulse of lesser-volume bimodal rocks took place at 20–19 Ma (fig. 2), and episodic minor eruptions of rhyolite and/or basalt continued through the remainder of the Cenozoic. Ash-flow tuff makes up less than 10 percent of the volume of the field.

The geologic studies leading to this report began in 1963, when J.J. Anderson and P.D. Rowley, then from the University of Texas and later with the U.S. Geological Survey (USGS), started geologic mapping in the Markagunt and Sevier Plateaus on the southern flank of the Marysvale volcanic field (fig. 3). The USGS began mapping and related studies in the Tushar Mountains and Sevier Valley in the central part of the field in 1975, when T.A. Steven and C.G. Cunningham initiated work on the volcanic rocks and associated mineral deposits. The USGS work was expanded into a broader study of the geology and mineral resources of the whole Richfield 1°×2° quadrangle in 1978. The Marysvale field segment of the quadrangle was studied generally full time by Steven, Cunningham, and Rowley and part time by many others from the USGS and academia. H.T. Morris (USGS), J.J. Anderson (Kent State University and USGS), and M.G. Best (Brigham Young University and USGS) were prominent among the many participants of the Richfield work. K-Ar and fission-track dating by H.H. Mehnert and C.W. Naeser (both USGS) were critical to the study. The overall Richfield investigation resulted in more than 200 open-file and published reports and maps; many of these are cited in Steven and Morris (1987). Most of our conclusions about the Marysvale volcanic field are based on detailed geologic mapping, although field relations that support individual statements are not generally documented here. Our speculative ideas about regional geology that make up the second half of this report are based on our work and on reports by others about nearby areas, and they were written mostly by the senior author; not all coauthors agree with these interpretations. Much of the field data given here come from a series of geologic, geophysical, and mineral-resources maps of the central part of the field that started with Cunningham and others (1983) and from a summary map by Steven and others (1990).

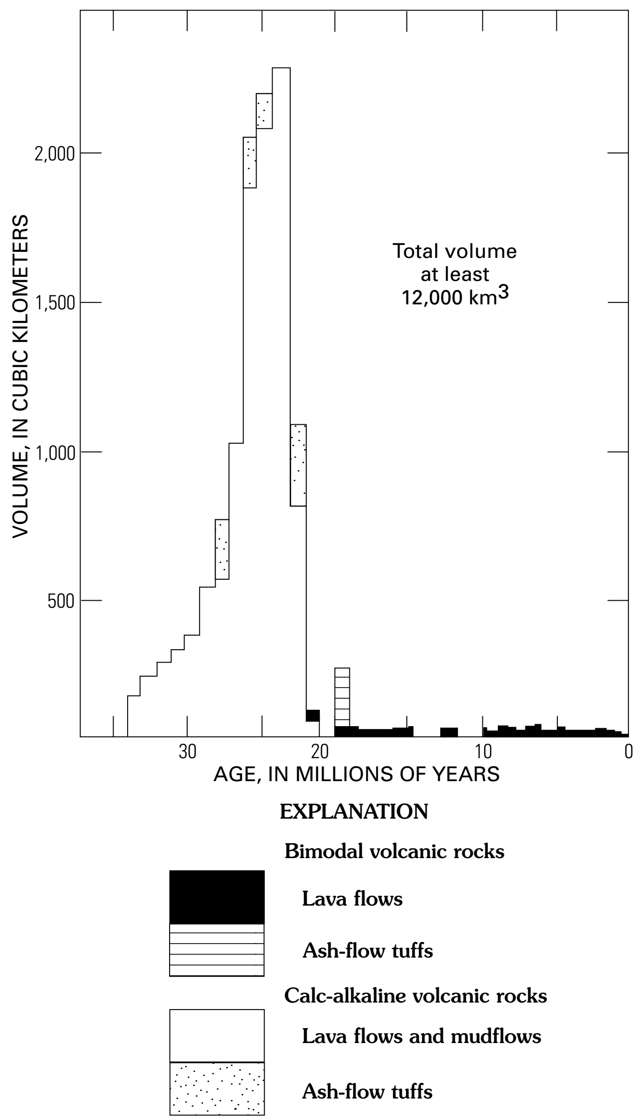

Figure 2. Volumes of volcanic rocks versus age, Marysvale volcanic field, Utah. Each bar represents the estimated volume erupted during a 1-million-year period.

This report summarizes field and other data dealing with the timing and character of Marysvale igneous and tectonic activity and places this activity in a larger igneous and tectonic context that includes laccolith clusters farther east in the Colorado Plateau and other igneous centers farther west in the Great Basin. In a companion report (Cunningham and others, this volume), we summarize chronologic patterns of igneous geochemistry and ore deposits. Isotopic ages given in this report without citing references are mostly from K-Ar and fission-track determinations discussed in various of our reports, most recently in Rowley, Mehnert, and others

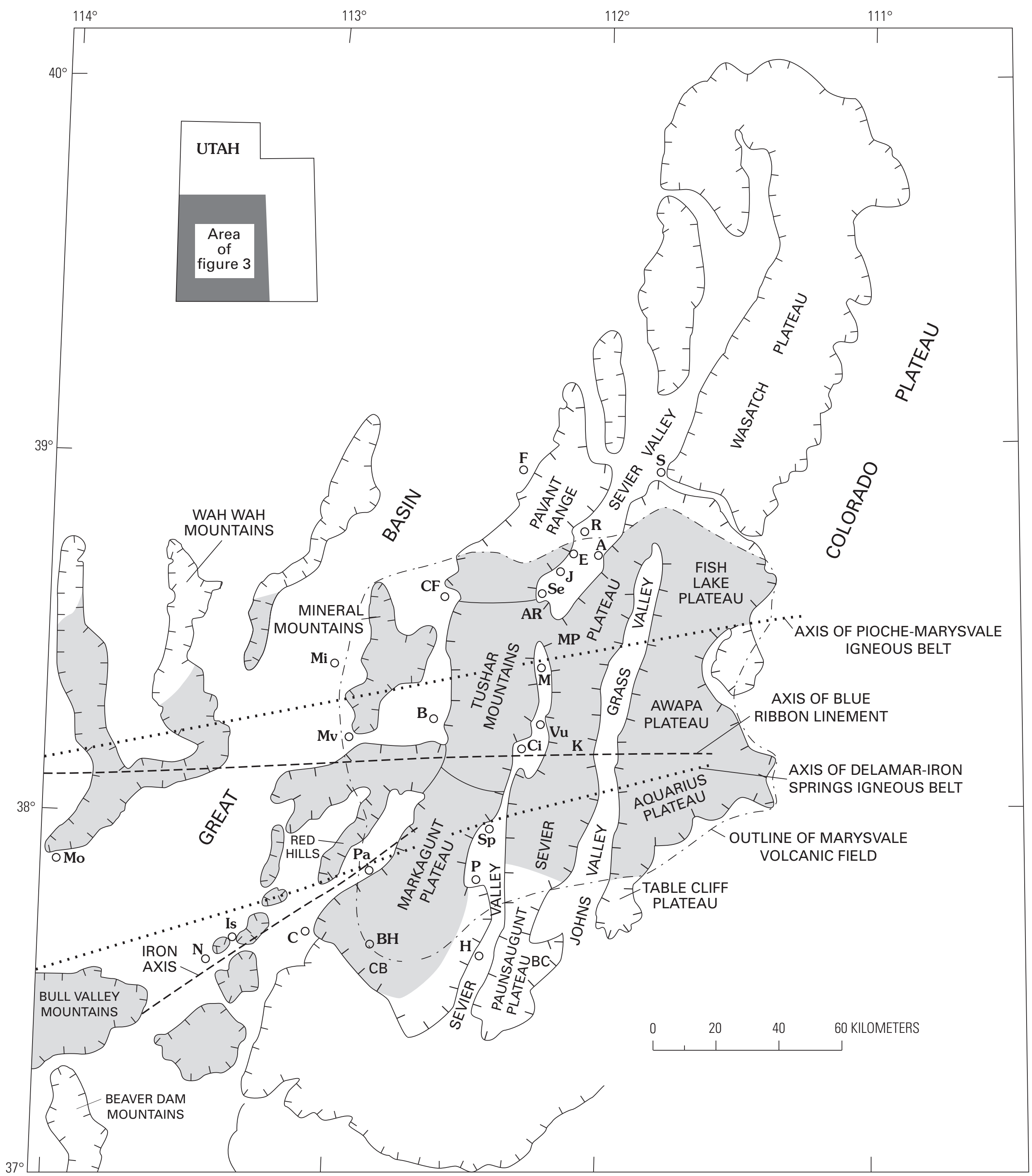

Figure 3. The Marysvale volcanic field and other features of interest in southwestern Utah (after J.J. Anderson and Rowley, 1975, fig. 1, and Steven and others, 1979, fig. 1). Shading shows areas underlain largely by volcanic rocks. The Colorado Plateau–Basin and Range boundary is just west of the towns of Nephi, Fillmore, Cove Fort, Beaver, Parowan, and Cedar City. A, Annabella; B, Beaver; BC, Bryce Canyon National Park; C, Cedar City; CB, Cedar Breaks National Monument; CF, Cove Fort; Ci, Circleville; E, Elsinore, F, Fillmore; H, Hatch; IS, Iron Springs; J, Joseph; Ju, Junction; K, Kingston Canyon; M, Marysvale; Mi, Milford; Mo, Modena; MP, Monroe Peak; Mv, Minersville; N, Newcastle; Ne, Nephi; P, Panguitch; Pa, Parowan; R, Richfield; S, Sevier; Sp, Spry.

(1994) . We are currently compiling data on the igneous isotope geochemistry of the Marysvale volcanic field, and S.R. Mattox and J.A. Walker (Northern Illinois University) are compiling similar data on some of the middle and upper Cenozoic volcanic units of the field (see Mattox, 1992, for instance).

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IGNEOUS ROCKS

MIDDLE CENOZOIC CALC-ALKALINE ROCKS

Middle Cenozoic, fundamentally calc-alkaline volcanism centered near Marysvale resulted in huge volumes of dacitic and andesitic lava flows, ash-flow tuff, flow breccia, and volcanic mudflow breccia from clustered stratovolcanoes and calderas and to a lesser extent from shield volcanoes and volcanic domes. The sequence corresponds to voluminous calc-alkaline igneous rocks common throughout the Western United States that probably formed during subduction of the Farallon/Vancouver plates (as described, for instance, by Lipman and others, 1972). Most rocks in the center of the Marysvale volcanic field contain abundant phenocrysts, commonly plagioclase, biotite, and hornblende. These moderately evolved rocks were derived from shallow magma chambers, and moderate erosion has exposed numerous plutons. Locally the roofs of high-level cupolas of the batholith failed, and calderas (Steven and others, 1984) formed as voluminous ash flows were erupted. The most voluminous stratigraphic unit in the central and northern part of the field (Tushar Mountains) is the Bullion Canyon Volcanics (Callaghan, 1939; Steven and others, 1979, 1984), which formed from magmas erupted from stratovolcano and caldera sources. The thickness of the formation is at least 1,500 m, and its original volume was at least 1,700 km^3. Basal parts of the unit underlie a regional stratigraphic marker, the Wah Wah Springs Formation (>29.5 Ma) of the Needles Range Group, an ash-flow unit derived from the Indian Peak caldera complex on the Utah-Nevada State line (Best, Christiansen, and Blank, 1989). These basal, weathered, and locally altered parts of the Bullion Canyon may be as old as 34 Ma (Willis, 1985; Kowallis and Best, 1990). Above the level of the Wah Wah Springs, the Bullion Canyon includes two ash-flow tuff members: the Three Creeks Tuff Member (27.5 Ma), which was derived from the Three Creeks caldera in the southern Pavant Range and had an estimated volume of 200 km^3 (Steven and others, 1984); and the Delano Peak Tuff Member (about 24 Ma), which was derived from the Big John caldera in the central Tushar Mountains and had an initial volume of 100 km^3. Mineralization associated with some of the shallow plutons that produced the Bullion Canyon Volcanics formed replacement alunite and gold deposits in many local districts in the Bullion Canyon terrain.

In contrast to the crystal-rich, moderately evolved rocks that formed the Bullion Canyon Volcanics, the rocks on the southern side of the Marysvale field (Markagunt Plateau and southern Sevier Plateau) are dominated by crystal-poor andesite and subordinate dacite of the Mount Dutton Formation (J.J. Anderson and Rowley, 1975; Fleck and others, 1975; Mattox and Walker, 1989; Mattox, 1992). Mapping shows that most of these generally pyroxene-bearing, less evolved rocks formed clustered stratovolcanoes whose centers occur along an east-trending belt, the Blue Ribbon transverse structure. (See “Transverse Structures” section, p. xx.) The Mount Dutton has a span of ages similar to those of the Bullion Canyon Volcanics, and the two formations intertongue from base to top across the south-central part of the volcanic field. Some Mount Dutton deposits underlie the Wah Wah Springs Formation (>29.5 Ma), but most do not; isotopic ages of those deposits above the Wah Wah Springs range from 26.7 to 21.2 Ma (Fleck and others, 1975). The total volume of the Mount Dutton Formation was at least 5,000 km^3. Plutons and mineral deposits are rarely exposed in Mount Dutton rocks, as magma chambers were deep and few convecting hydrothermal cells were formed. Members of the Mount Dutton Formation include the Beaver Member (25.6 Ma), which consists of volcanic domes with an estimated volume of 20 km^3 in the eastern Black Mountains, and the Kingston Canyon (25.8±0.4 Ma; volume 20 km^3) and Antimony (25.4±0.9 Ma; volume 50 km^3) Tuff Members, which are thin, densely welded, trachytic ash-flow tuffs that intertongue with the Mount Dutton, Bullion Canyon, and other sequences of mudflows and lava flows. The tuff members are synchronous with and petrologically similar to the tuff of Albinus Canyon (25.3±1.3 Ma; volume 60 km^3) in the northern part of the field and probably will eventually be determined to be identical. Probably all were derived from a common as-yet undiscovered source in the southern Pavant Range or adjacent Sevier Valley; Ekren and others (1984) suggested deep sources for petrologically similar tuffs in Idaho.

The Marysvale volcanic field contains many local calc-alkaline volcanic centers; products from these centers were separated during the mapping from the much more voluminous Bullion Canyon Volcanics and Mount Dutton Formation. Among the oldest of these are three informal units in the northern Tushar Mountains and southern Pavant Range that consist of dacitic flows, breccia, and tuff derived from stratovolcanoes and volcanic domes. These units were included in the Bullion Canyon Volcanics by Steven and others (1990) but they were broken out by Cunningham and others (1983) as the volcanic rocks of Dog Valley, the overlying volcanic rocks of Wales Canyon, and an unnamed dome. They underlie the Three Creeks Member and have been deeply eroded; their total volume was at least 300 km^3. The volcanic rocks of Signal Peak overlie the Three Creeks Member mostly in the northern Sevier Plateau. They consist of an andesite to basaltic andesite volcanic plateau and shield volcano with a volume of at least 400 km^3, not including unmapped rocks in areas to the east. The Bullion Canyon Volcanics and the Mount Dutton Formation intertongue eastward with coeval stratovolcano and shield-volcano units, whose extensions east of long 112° W. are as yet unmapped. These units include the volcanic rocks of Little Table, of Willow Spring, and of Langdon Mountain; where we mapped them in the Sevier Plateau, they have a combined volume of 400 km^3. Stratovolcano rocks of the eastern Marysvale field (Williams and Hackman, 1971; Mattox, 1991a) have a volume of about 2,500 km^3 and undoubtedly include deposits of these three units as well as of the volcanic rocks of Signal Peak and the Mount Dutton Formation. In the southern Tushar Mountains, three younger local units overlie the Osiris Tuff (see below): (1) the formation of Lousy Jim, a trachydacite volcanic dome (22 Ma; 25 km^3); (2) the tuff of Lion Flat, a local rhyolite ash-flow tuff with a volume of 8 km^3, which may represent deposits in and near a small concealed caldera or may represent the tuff-ring deposits (C.G. Cunningham, unpub. data, 1992) of the overlying formation of Lousy Jim; and (3) the andesitic lava flows of Kents Lake (12 km^3).

A series of shallow, calc-alkaline laccoliths were emplaced into lower Tertiary sedimentary rocks in the northern Markagunt Plateau, south of the deep-seated Mount Dutton sources and on the southern flank of the Marysvale field, where the volcanic rocks are relatively thin (J.J. Anderson and Rowley, 1975). The largest of these laccoliths is the Spry intrusion (26–25 Ma; fig. 3), which erupted to form a dacitic vent complex containing lava flows and tuffs that intertongue with the lower part of the Mount Dutton Formation (J.J. Anderson and others, 1990a). Breccia and flows from the complex are called the volcanic rocks of Bull Rush Creek, and a regional ash-flow tuff from the complex is called the Buckskin Breccia; their combined volume is about 60 km^3. A prominent short-wavelength gravity low (Blank and Kucks, 1989; Cook and others, 1989, 1990; Saltus and Jachens, 1995; Blank and others, this volume; Viki Bankey, USGS, unpub. data, 1992) and short-wavelength aeromagnetic anomalies (Zietz and others, 1976; Blank and Kucks, 1989; Viki Bankey, USGS, unpub. data, 1992) underlie and extend 50 km south of the vent complex. These anomalies probably mark the upper part of a source batholith that underlies and fed the laccolith. Another laccolith on the southern flank of the Marysvale field is the Iron Peak intrusion (formerly called Iron Point intrusion by J.J. Anderson and Rowley, 1975) of apparently 21–20 Ma, which fed a series of flows and breccias that intertongue with the upper part of the Mount Dutton Formation (Spurney, 1984). This pluton formed at the northeastern end of the “Iron Axis,” a string of a dozen laccoliths and other plutons in the Great Basin that formed major iron deposits (Mackin, 1947; Blank and Mackin, 1967; Rowley and Barker, 1978; Van Kooten, 1988; Rowley and others, 1989; Barker, 1991; Blank and others, 1992) in the Iron Springs and Bull Valley mining districts west and southwest of Cedar City (fig. 3). Geophysical data suggest that most if not all of the Iron Axis laccoliths (22 Ma) are interconnected at depth, and the nearby younger (20.5-Ma) Pine Valley laccolith (Cook, 1957; D.B. Hacker, Kent State University, and L.W. Snee, USGS, unpub. data, 1992) is also part of the trend. All are interpreted to be fed by a large underlying composite batholith (Blank and Mackin, 1967; Blank and others, 1992), and this batholith may have extended as far northwest as a pluton (Grant and Proctor, 1988) exposed 30 km northwest of Cedar City (H.R. Blank, unpub. data, 1992). Borrowing an idea of H.R. Blank (oral commun., 1991), which was developed in greater detail by Nickelsen and Merle (1991), Merle and others (1993), J.J. Anderson (1993), and Maldonado (1995), we suggest that the Spry and the other plutons in the northern Markagunt Plateau may be cupolas on a large composite batholith that underlies the northern Markagunt Plateau and adjacent areas and led to a variety of deformational features in roof rocks above the batholith. This batholith would be similar to the one that fed the Iron Axis plutons and may even be connected to it. Rocks of the Mount Dutton Formation as well as the Iron Axis and northern Markagunt

Plateau plutons are part of the Delamar–Iron Springs igneous belt, which is south of the Pioche-Marysvale igneous belt (figs. 1, 3).

The Osiris Tuff (23 Ma) intertongues with the tops of the Bullion Canyon Volcanics and the Mount Dutton Formation. Field relations and petrologic studies show that this distinctive rhyolitic to trachytic tuff was erupted from the Monroe Peak caldera in the central and northern Sevier Plateau and probably had an original volume of at least 250 km^3. Andesite to rhyolite lava flows overlie intracaldera deposits of the Osiris Tuff and are considered to represent the final stages of caldera volcanism; they have isotopic ages of 22–21 Ma and a total volume of about 100 km^3. The Osiris and its associated lava flows represent the last major calc-alkaline volcanic event in the Marysvale field. The magmatic cupola whose top collapsed to form the Monroe Peak caldera resurgently invaded the caldera fill, where it formed intracaldera plutons (23–22 Ma) and it hydrothermally altered the rocks and mineral deposits of the Marysvale district in the western part of the caldera (Steven and others, 1984; Rowley and others, 1986a, b, 1988a, b).

As reported by other participants of this workshop (Sullivan and others, 1991; Sullivan, this volume; Nelson and others, 1992; Nelson, this volume; Ross, this volume), Colorado Plateau laccoliths in southeastern Utah have similar ages not only to the plutons of the Marysvale volcanic field described here, but to volcanic and intrusive rocks in the San Juan volcanic field in southwestern Colorado. Similar ties appear to exist between these igneous centers and centers in the Great Basin extending west to the Reno area. The Colorado Plateau laccoliths also are closely similar in composition to the calc-alkaline rocks of the central Marysvale area, Markagunt Plateau, and Iron Axis (fig. 4). chemically, the Spry intrusion and the plutons of the Iron Axis are closely affiliated with each other, and they differ somewhat from Marysvale plutons. Plutons from the central Marysvale field and the Henry, Abajo, and La Sal Mountains are progressively more alkalic eastward (fig. 4). The only highly alkaline rocks known in the Marysvale field belong to a small alkali breccia pipe (24 Ma; Rowley, Mehnert, and others, 1994) containing nepheline and corundum 20 km southeast of Marysvale (Agrell and others, 1986).

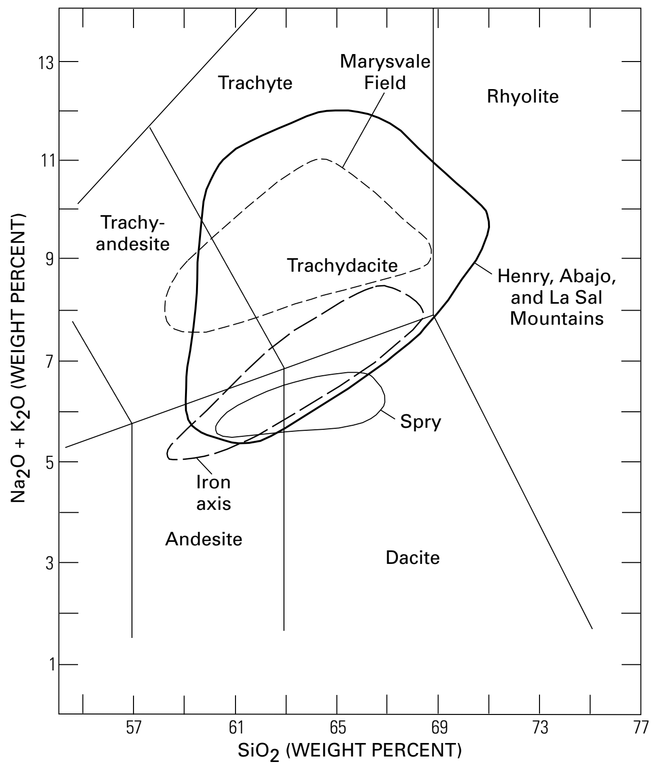

Figure 4. Total alkalies versus silica (Le Bas and others, 1986) of calc-alkaline rocks from the Marysvale volcanic field, Utah, compared with those from the Henry, Abajo, and La Sal Mountains, from the Iron Axis, and from the Spry intrusion. Data from Witkind (1964), C.B. Hunt (1958, 1980), G.L. Hunt (1988), D.S. Barker (University of Texas at Austin, unpub. data, 1985), and other sources.

UPPER CENOZOIC BASALT AND ALKALI RHYOLITE

As in other parts of the Western United States (see, for example, Christiansen and Lipman, 1972; Christiansen and McKee, 1978; McKee and Noble, 1986), the petrologic regime of the Marysvale volcanic field changed dramatically (Cunningham and others, this volume) from calc-alkaline to fundamentally bimodal (basalt and high-silica rhyolite) magmatism, starting no later than 23–22 Ma. The change occurred at different times in different parts of the field. Bimodal volcanic rocks make up about 5 percent of the total volume of the Marysvale field. The change to a bimodal petrologic regime is inferred to broadly coincide with the inception of crustal extension in a brittle crust that cooler and thicker overall than the crust formed during calc-alkaline magmatism (Lucchitta, 1990; Wernicke, 1992; Lucchitta and Suneson, 1993). As a result, little-contaminated basaltic magmas rose along more deeply penetrating fractures to higher in the crust, leading to a rise in isotherms and the development of alkali-rhyolite eutectic partial melts (C.G. Cunningham, unpub. data, 1988). An alternative model, which we do not favor, suggests that an increase in the emplacement of basalts formed some rhyolite calderas predominantly by fractional crystallization (Perry and others, 1993). Few faults that record extension beginning at 23–22 Ma at Marysvale have been identified, whereas those faults that record the main phase of deformation (basin-range extension), starting about 9 Ma in this area, are numerous and obvious.

Generally the first product of bimodal magmatism consists of potassium-rich mafic rocks from local centers, including the mafic lava flows of Birch Creek Mountain

(22.5 Ma; 40 km^3) and the mafic lava flows of Circleville Mountain (23 Ma; 25 km^3) in the southern Tushar Mountains. These rocks were called by several names, including "older basalt flows" (J.J. Anderson and Rowley, 1975), until we settled on "potassium-rich mafic volcanic rocks," following Best and others (1980); they are shown as such by Steven and others (1990). Best and others (1980), Mattox and Walker (1989, 1990), Walker and Mattox (1989), and Mattox (1991b, 1992) discussed their chemical composition. Potassium-rich mafic volcanic rocks with a volume of about 10 km^3 and similar ages (Best and others, 1980) also occur in the eastern Marysvale field (Williams and Hackman, 1965; Mattox, 1991a, 1992); some of these rocks may be as old as 26 Ma (Mattox, 1991a). Lava flows erupted from other centers (22–21 Ma; Best and others, 1980), on the southern and northern edges of the field, have a total volume of about 10 km^3. Mattox and Walker (1989) ascribed all this potassic mafic volcanism to the final demise of the subduction zone. Mattox (1992), however, noted a similarity in chemical and isotopic composition between the potassium-rich mafic volcanic rocks and the Mount Dutton Formation, which suggests that both are products of calc-alkaline magmatism. The upward movement of potassium-rich mafic magmas at depth may have supplied the heat to produce partial crustal melts erupted as early bimodal rhyolites.

The largest accumulation of volcanic rocks of the bimodal assemblage in the Marysvale field consists of rhyolite of the Mount Belknap Volcanics (Callaghan, 1939; Cunningham and Steven, 1979a), which was erupted from two concurrently active source areas in the central Tushar Mountains and Sevier Valley from 22 to 14 Ma. The westernmost of these two source areas subsided to form the large Mount Belknap caldera when its main ash-flow tuff, the Joe Lott Tuff Member (19 Ma), was erupted (Steven and others, 1984; Budding and others, 1987). Including associated intracaldera rhyolite flows and plugs, the original volume of rocks of the western source area was about 300 km^3. Heterogeneous silicic lava flows, volcanic domes, and tuff were erupted from the eastern source area. The small Red Hills caldera, in Sevier Valley, subsided with eruption of the Red Hills Tuff Member at 19 Ma. The volume of rhyolites of the eastern source area is about 25 km^3. Silicic intrusive rocks (rhyolite dikes and plugs and granitic plutons) were emplaced into the two source areas. The main mineral deposits from Mount Belknap magmatism are of uranium and molybdenum in the Central Mining Area of the Marysvale district (Cunningham and Steven, 1979b; Cunningham and others, 1982). In the eastern Tushar Mountains 10 km south-southwest of Marysvale, an intrusive cupola domed rocks of the Bullion Canyon Volcanics (including the Three Creeks Member) at Alunite Ridge (Cunningham and Steven, 1979c) and formed a halo of gold-bearing veins, a base- and precious-metal manto (Beaty and others, 1986), and, above that, coarse-grained vein alunite dated at 14 Ma (Cunningham and others, 1984).

The Mineral Mountains, at the western edge of the Marysvale field, are underlain by a 25- to 9-Ma composite batholith, the largest exposed in Utah (Nielson and others, 1986; Steven and others, 1990; Coleman and Walker, 1992). It is probably representative of the intrusive rocks that underlie the Pioche-Marysvale igneous belt and is exposed over a large area here only because the Mineral Mountains underwent huge amounts of uplift when the batholith was emplaced as part of a probable metamorphic core complex (Price and Bartley, 1992). Minor early phases (about 25 Ma) represent calc-alkaline intrusions, but later intrusions (22–11 Ma) are more silicic and alkaline and appear to be part of the bimodal association; still later granites (9 Ma) are clearly bimodal.

Beginning at 9 Ma, when the crust began to extend rapidly and a surge of basin-range faulting took place, small rhyolite and basalt eruptions occurred at many scattered locations around the edges of the Marysvale field. The largest of the rhyolite bodies is in the Kingston Canyon area of the southern Sevier Plateau, where rhyolite flows and a volcanic dome of 8–5 Ma make up a total volume of about 50 km^3 (Rowley, 1968; Rowley and others, 1981). Small rhyolite domes (9–8 Ma) that total about 25 km^3 are scattered in the Basin and Range province on the western side of the Marysvale field (J.J. Anderson and others, 1990b). The rhyolite of Gillies Hill (9–8 Ma; 20 km^3) of Evans and Steven (1982) was emplaced along north-striking basin-range faults on the northwestern side of the Beaver Basin, just east of the Mineral Mountains; subsurface plutons of the same age and same presumed lithology have associated mineralized rocks at the Sheeprock mine along the eastern side of the basin (Cunningham and others, 1984). The Sheeprock plutons, which are the intrusive equivalent of the rhyolite of Gillies Hill, may correlate with the youngest plutons of the Mineral Mountains batholith to the west (Coleman and Walker, 1992). Capping this batholith, on the crest and western flank of the Mineral Mountains, is the rhyolite of the Mineral Mountains, a series of Pleistocene (0.8–0.5 Ma) rhyolite domes, lava flows, and tuff (Lipman and others, 1978).

Basalt fields as young as Quaternary (see, for instance, Best and others, 1980, or Mattox, 1992) are scattered through and around the Marysvale field. Sites of eruption do not coincide with older vent areas of calc-alkaline rocks or of the oldest (23–16 Ma) high-silica rhyolite flows. Some basalt vents do coincide, however, with post-9-Ma high-silica rhyolite sources, suggesting that basalt emplacement supplied the heat for those rhyolite partial melts. We calculate the volume of basalt rocks to be at least 100 km^3 in the field and more than this beyond the field. Basalts in the field in general represent asthenospheric melts that worked their way to the surface with relatively little contamination (Fitton and others, 1991) along northerly striking fractures created during basin-range extension. Although erosion has removed or obscured many vents, the presence of basalt flows as louderbacks on the major tilt blocks suggests that many vents were on the edge of the upthrown (footwall) fault blocks in the manner observed in the field by R.E. Anderson (1988) and explained theoretically by Ellis and King (1991) as due to dilation in footwall blocks at depth during normal faulting.

REGIONAL VERSUS LOCAL EXTENSIONAL DEFORMATION

Extensional deformation in the Marysvale volcanic field took place during two different episodes that correspond roughly to the magmatic episodes. The concept of the two extensional episodes is hardly new and now is accepted by most workers (for example, Zoback and others, 1981), although many (for example, Wernicke, 1981, 1985; Von Tish and others, 1985) consider that extension in the Basin and Range province has taken place during a single long event. The evidence for two episodes in the Marysvale area is sketchy, so in the sections below, we bring in map evidence from other parts of Utah and Nevada where we have worked. From there, we expand our view of the Marysvale area and offer ideas on the Cenozoic geologic evolution of much of the Great Basin and Colorado Plateau.

MIDDLE CENOZOIC FAULTS

Local extensional deformation took place during middle Cenozoic calc-alkaline magmatism in the Marysvale volcanic field, as elsewhere in the transition zone and the Great Basin. The orientation of such pre-basin-range, syn-calc-alkaline faults indicates east-west extension at Marysvale, but the faults are sparse and widely scattered and are outside the main intrusive centers. More of these middle Cenozoic faults may exist, but either (1) they cannot be distinguished in age from the dominant late Cenozoic (basin-range) faults in the area or (2) field evidence for constraining the age of the faults of this episode was covered by volcanic rocks, obscured by the poorly known stratigraphy of these thick volcanic piles, or obscured by intrusive structures.

West of the town of Antimony in the southern Sevier Plateau, an angular unconformity due to a northerly striking fault zone separates units having 30° difference in dip. The age of movement is constrained by the Kingston Canyon Tuff Member (26 Ma) below and the Osiris Tuff (23 Ma) above (Rowley, 1968, p. 144, 167). Several west-northwest-striking horsts and grabens of 26–25 Ma were mapped in the northern Markagunt Plateau of the southern Marysvale field by J.J. Anderson (1971, 1985, 1988). Their unusual strike suggests that they represent deformation along the broad west-striking Blue Ribbon transverse structure (see “Transverse Structures” section, p. xx), which underlies the area of the horsts and grabens and was active during and after this time. Other pre-basin-range faults include local “mosaic faults” of numerous strikes that represent shattering of roof rocks above many calc-alkaline intrusions on the batholith complex of the Pioche-Marysvale igneous belt (Steven, 1988).

The northern Markagunt Plateau and adjacent areas also contain the following, perhaps genetically related, pre-basin-range structures, which probably resulted at least partly from the emplacement of a batholith complex under the eastern Delamar–Iron Springs igneous belt: (1) the Markagunt megabreccia, a major pre-24-Ma gravity-slide deposit (Sable and Anderson, 1985; J.J. Anderson, 1993; L.W. Snee, written commun., 1994); (2) the low-angle, 22.5–20 Ma, extensional Red Hills shear zone (Maldonado and others, 1989, 1990, 1992; Maldonado, 1995); and (3) 30- to 20-Ma thin-skinned thrust faults (Davis and Krantz, 1986; Lundin and Davis, 1987; Lundin, 1989; Bowers, 1990; Nickelsen and Merle, 1991; Merle and others, 1993). These diverse structures may represent gravity-driven volcanic spreading southward off the Marysvale volcanic field (Davis and Rowley, 1993).

The near absence of middle Cenozoic extensional faults in the Marysvale volcanic field, despite the abundance of faults of this age farther west in the Great Basin, was a significant conundrum for us for a while. Gans and others (1989, p. 48) best defined the problem when they asked why middle Cenozoic faults are least common in the largest volcanic fields in the West, such as the Marysvale and San Juan fields. We conclude that part of the answer to the puzzle derives from calculations and interpretations by Lachenbruch and Sass (1978) showing that emplacement of basaltic dikes can accommodate extension to the exclusion of normal faulting. Bacon (1982) applied these ideas to the Pleistocene Coso volcanic field of California, where he found a constant rate of emplacement of basalt and rhyolite feeder dikes, which he attributed to relief of crustal least principal stress normal to these dikes. Bursik and Sieh (1989) extended these concepts to the Long Valley–Mono Basin volcanic field in California. There, they concluded, Quaternary oblique faults of the same age as the volcanic field die out along strike into eruptive centers because dikes in the centers take up all extensional strain. (See also Moos and Zoback, 1993.) Thompson and others (1990), Parsons and Thompson (1991), and Parsons and others (1992) expanded the calculations into a more elaborate hypothesis that states that emplacement of basaltic dikes will increase horizontal stress (relieve least principal stress) so that normal faults will not form; this effect may explain the lack of seismicity in the eastern Snake River Plain of Idaho (Thompson and others, 1990). Pierce and others (1991, fig. C7), Pierce and Morgan (1992, plate 1, p. 41–42), and Pierce (oral commun., 1994) explained faults that die out into the Pleistocene Yellowstone caldera of Wyoming by the same mechanisms. In a summary of the calderas (15–9 Ma) of the Nevada Test Site, Sawyer and others (1994, p. 1316) noted that the areas of greatest volume of intracaldera intrusions contain the fewest faults;

they allowed that this could be explained by emplacement of intracaldera intrusions relieving the least principal stress but thought it more likely was due to the cooled and solidified intrusions and their underlying batholith creating a massive buttress that deflected faults around them. In contrast, we here expand on the ideas cited above to suggest that middle Cenozoic, syn-calc-alkaline faults are sparse in the Marysvale field and many other areas because most east-west least principal stress was taken up by passive intrusion of *shallow* batholiths, stocks, and dikes of all calc-alkaline compositions, rather than by faults. In other words, relief of least principal stress in the brittle upper crust can be, depending on the local stress field, by faults or by emplacement of hypabyssal intrusions, or by both. Most dikes and perhaps some shallow intrusions would be expected to strike perpendicular to the extension direction, assuming the general case that they filled joints that propagated during intrusion (Delaney and others, 1986; Grout, this volume). In the Marysvale field, however, dikes are uncommon, and we have noticed no preferred northerly trend for dikes or for the long directions of intrusions.

Faults of all types also are uncommon in the Iron Springs mining district and other parts of the Iron Axis, which are underlain by a large terrain of shallow, calc-alkaline intrusive rocks; we attribute the lack of syn-calc-alkaline (middle Cenozoic) faults also to stress relief by the shallow intrusions, although a general lack also of many younger basin-range faults is better explained by the buttress effect of Sawyer and others (1994; see also Blank, 1994). After our ideas were formulated, J.E. Faulds (University of Iowa oral commun., 1995) alerted us to similar ideas by him and his colleagues (Faulds, Olson, and Litterell, 1994) in which they refer to “magmatic extension” as an alternative to “mechanical extension” (that is, faulting).

Deeper intrusions are an entirely different matter, but still important during extension. Thompson and Burke (1974), Eaton (1980), Gans (1987), McCarthy and Parsons (1994), and others recognized that emplacement and underplating by deeper intrusions and lateral flow of lower crustal material are the main reasons why rapidly extending crust in areas such as the Great Basin is not also drastically thinned. Catchings and Mooney (1993) cited seismic data that suggest that deep and shallow intrusions occur where intensely faulted areas occur. When extending, brittle upper crust may be “decoupled from a more uniformly deforming (ductile) middle and lower crust” (Gans, 1987, p. 1), as also suggested by Eaton (1980, 1982) and Hamilton (1988a). The laccolith fields of the Colorado Plateau and, of course, the plateau itself, also are little broken by faults. One of us (T.A. Steven) suggests here that the Plateau “floated on” a hot, plastic, periodically and locally partially melted substratum that took up least principal stress in the area of the laccoliths, and Blank and others (this volume) propose that large masses of magma underlay the Colorado Plateau during the early and middle Miocene.

Although syn-calc-alkaline extensional faults in the Marysvale field are sparse, faults of this age are abundant in an area farther west where Rowley is mapping in the western (Nevada) part of the Delamar–Iron Springs igneous belt. This area, near Caliente, Nev. (fig. 1), was deformed by two episodes of east-west extension, preceded by a deformational episode whose age and explanation are not understood. This oldest deformation resulted in the mostly bedding-parallel Stampede fault (Axen and others, 1988), in which the transport direction of the upper plate may have been eastward. All that can be told about the age of deformation is that it preceded the oldest volcanic rocks (31 Ma; Rowley, Shroba, and others, 1994) in the area.

The Stampede was considered to be a detachment fault (Axen and others, 1988; Bartley and others, 1988; Taylor and others, 1989; Rowley, Shroba, and others, 1994), which Taylor and Bartley (1992) and Axen and others (1993) correlated with the Snake Range décollement (Miller and others, 1983; Gans and others, 1985) to the north (fig. 1), and thus they considered that it represents a significant early Tertiary episode of extension. However, these interpretations are questionable, for the Stampede also resembles bedding-parallel attenuation faults found in association with compressional structures that were mapped by Miller (1991), Nutt and Thorman (1992, 1993, 1994), and Nutt and others (1992, 1994) in northwestern Utah and northeastern Nevada. These authors showed the structures to be pre-lower(?) Eocene and suggested that they are of Sevier age. New data suggest that the structures predate a Cretaceous intrusion and perhaps a Jurassic intrusion (C.J. Nutt and C.H. Thorman, oral commun., 1995). The faults involved in these structures in the northeastern Great Basin are similar to mapped younger-on-older, bedding-parallel faults considered by Allmendinger and Jordan (1984) to be pre-Late Jurassic and by Miller (1991) to be Mesozoic. Similar faults associated with general compressional (Sevier) thrusts in the Raft River area of northwestern Utah and southern Idaho have been isotopically dated at Late Cretaceous (Wells and others, 1990). “Extensional faults” mapped in the Devonian Guilmette Formation south of the western Caliente caldera complex by Page and Scott (1991) were interpreted by Axen and others (1993) to be part of their early Tertiary extensional event, but these and other such faults in the area are now considered to be of Sevier age (Swadley and others, 1994; Page, 1995). This deformational episode will not be considered further because of uncertainties in its age and interpretation.

The faults of the two clear Cenozoic extensional episodes studied during mapping in progress by Rowley and R.E. Anderson are well exposed in the western Caliente caldera complex, Nevada-Utah (at, south, and east of the town of Caliente, fig. 1) and adjacent areas. The older episode, representing syn-calc-alkaline, pre-basin-range faults, is the main deformational event in the area and resulted in high-angle, mostly north-northwest- and north-northeast-striking oblique-slip, strike-slip, and normal-slip faults and in north-striking detachment faults (Axen and others, 1988; Burke, 1991; Rowley, Snee, and others, 1992). The ages of these older faults are well constrained by dated dikes (20–16 Ma; Rowley, Snee, and others, 1992; L.W. Snee, written commun., 1993) intruded into the fault zones, and field evidence suggests that the episode probably began close to 25 Ma; faulting continued until about 12 Ma. This episode coincided with magmatism in the Caliente caldera complex (23–13 Ma), a large (80 by 35 km) complex of inset calderas that erupted low-silica rhyolite (calc-alkaline) tuffs until about 18 Ma, then high-silica rhyolite (bimodal) tuffs after about 15.5 Ma (Rowley, Snee, and others, 1992; Rowley and others, 1995).

Middle Cenozoic strain in the Caliente area was complicated and heterogeneous, and it was similar to that described and interpreted by R.E. Anderson (1987), Siders and Shubat (1986), R.E. Anderson and Bohannon (1993), R.E. Anderson and Barnhard (1993a, b), R.E. Anderson and others (1994), and Scott, Grommé, and others (1995) east and south of the Caliente caldera complex (see next paragraph). On the basis of mapped structures and of fault kinematic data analyzed according to the methods of Angelier and others (1985) and Petit (1987), we follow the reasoning, though not necessarily the orientation of the extension direction, of Wright (1976; his "field II" faults), R.E. Anderson (1973, 1984, 1986, 1987, 1989, 1990), Angelier and others (1985), Zoback (1989), R.E. Anderson and Barnhard (1993a, b), and R.E. Anderson and others (1994). The middle Cenozoic extension direction (that is σ_3, the least-principal-stress axis) in the Caliente area is interpreted to be horizontal and east-west on the basis of a conjugate set of predominant north-northwest oblique (right-lateral, normal) and subordinate north-northeast oblique (left lateral, normal) faults (Rowley, unpub. data). This orientation differs from the east-northeast extension direction suggested by R.E. Anderson and Ekren (1977), Zoback and others (1981), and Michel-Noël and others (1990), all of whom interpreted the north-northwest-striking faults to be normal faults. The σ_1 (greatest-principal-stress axis) was locally or periodically vertical, in order to explain north-striking normal faults, but more generally it was oriented horizontally and trended generally northward (that is, it reflects north-south shortening); furthermore, σ_1 and σ_2 (intermediate-principal-stress axis) may have been similar to each other in magnitude and they may have interchanged at times (Wright, 1976; Zoback, 1989). When σ_1 and σ_2 were equal, north-striking dikes and intrusions would be most likely to form; repeated injections of magma as extension continued would have resulted in batholith complexes (igneous belts) elongated parallel to σ_3. We agree with Angelier and others (1985) that normal and strike-slip faults in the same area "represent stress oscillations in time and space rather than discrete stress reorganizations." The faults of this main episode occur both inside and outside the Caliente caldera complex and so are not a consequence of magmatism. The Caliente caldera complex differs from the northern and central Marysvale volcanic field in that intrusions are not commonly exposed at Caliente and thus are not shallow. The strike-slip faults in the Caliente area do not parallel the extensional kinematic axis (the direction of extension), so we do not interpret them as tear faults in the upper plates of detachment faults, as did Michel-Noël and others (1990).

There is a considerable literature documenting extension on pre-basin-range faults throughout other parts of the Great Basin. These are all called middle Cenozoic faults here, although in the northern Great Basin, they are as old as Eocene, as are the Tertiary igneous rocks there. One of the first to report these faults was Ekren and others (1968) from the Nevada Test Site (fig. 1); their conclusions were further documented by Ekren and others (1971), R.E. Anderson and Ekren (1977), and R.E. Anderson (1978). In fact, extension is a characteristic feature of subduction-related arc magmatism (Hamilton, 1988c, 1989, 1995). Like Elston (1984), we consider that in some parts of the Basin and Range province, including Caliente, middle Cenozoic extension was greater than late Cenozoic extension. Basins containing clastic sedimentary fill, smaller than those of the later basin-range episode, have been identified in some areas (for example, Bohannon, 1984; Seedorf, 1991; Christiansen and Yeats, 1992; Axen and others, 1993). Miller (1991) and Liberty and others (1994) interpreted seismic data in the central and northern Great Basin to indicate that there, as in the Marysvale area (see section on "Late Cenozoic Faults and Basin-Fill Deposits") and some other parts of the central Great Basin that they cited, broad sag basins preceded the narrow fault-bounded basins of the basin-range episode. Other basins of this age may be unrecognized or buried under younger basin-fill sediments deposited in the same place. Metamorphic core complexes and both related and unrelated low-angle detachment faults (for example, Proffett, 1977; Crittenden and others, 1980; Wernicke, 1981, 1985, 1992; Wernicke and others, 1985, 1988; Allmendinger and others, 1983; Hamilton, 1987, 1988a, b; Davis and Lister, 1988; Axen, 1991; Wernicke and Axen, 1988), developed in many places in the Basin and Range province during calc-alkaline magmatism (Lipman, 1992; Lister and Baldwin, 1993). The Basin and Range province now is characterized by high heat flow due to extension (Lachenbruch and Sass, 1978; Lachenbruch and others, 1994), and heat flow was probably greater during the middle Cenozoic (Lachenbruch and others, 1994), with a resulting high level for the brittle-ductile transition, thermal softening of brittle crust, and the development of voluminous plutons and metamorphic core complexes (Hamilton, 1988a; Armstrong and Ward, 1991). R.E. Anderson (1990) and R.E. Anderson and others (1994) theorized a method of middle to late Cenozoic "tectonic escape" and structural rafting of the brittle upper crust on a laterally (westward to southwestward) flowing mass of mid-crustal material in order to explain a complex array of faults characterized by north-south shortening and east-west extension and of synextensional intrusions in the Lake Mead–Las Vegas area (fig. 1). This idea is attractive for the extreme southern Great Basin and northern part of the southern Basin and Range province, but for other parts of the Great Basin during the middle Cenozoic, any such tectonic escape must be superimposed on a more regional pattern of subduction and extension to explain regional structural trends of which the extreme southern Great Basin is a part.

Zoback and others (1981) compiled data on faults of 20–10 Ma in the Basin and Range province and concluded that the overall orientation of σ_3 then was east-northeast, parallel to the subduction direction (Hamilton, 1989). Henry and Aranda-Gomez (1992) noted the same orientation for middle Cenozoic faults in the Basin and Range province of Mexico. This orientation appears to be valid as an average, but there are some exceptions. One is the east-west extension direction in the Caliente area. The Marysvale field also may be an exception, although the data are too sparse to constrain the extension direction to better than a general east-west trend.

In the Great Basin and other areas of major extension, a direct, one-to-one causal correlation between extensional faulting and older and synchronous calc-alkaline magmatism has been proposed (Gans and others, 1989), by which magmatism thermally weakens the crust and leads to faulting. Nonetheless, the hypothesis is difficult to prove (Best and Christiansen, 1991; Wernicke, 1992; Axen and others, 1993): 35–20 Ma ash-flow tuffs were spread over large parts of the central Great Basin but were accompanied by little evidence of regional extension or topographic relief, such as angular unconformities or clastic sedimentary rocks between them (McKee and others, 1970; Best and Christiansen, 1991; McKee and Noble, 1986; E.H. McKee, oral commun., 1992). However, local angular unconformities are commonly obscure and difficult to find, and local clastic sedimentary units are easily removed by later erosion, so this negative evidence should be used with caution. Shallow intrusions are closely associated in time and place with significant extension in some nearby areas, as in the Caliente caldera complex (Rowley and R.E. Anderson, mapping in progress), Kane Springs Wash caldera (Scott, Grommé, and others, 1995) just to the south, and Wilson Ridge pluton in the Lake Mead area (R.E. Anderson and Barnhard, 1991; Barnhard and R.E. Anderson, 1991; R.E. Anderson, 1993; R.E. Anderson and others, 1994). A close association has been documented between shallow intrusions and extensional faulting in other areas in the West, including those described by Tobisch and others (1986), Hutton (1988), Lipman (1988), Glazner and Ussler (1989), Armstrong (1990), Ferguson (1990), Haxel and others (1990), Armstrong and Ward (1991), Hardyman and Oldow (1991), Meyer and Foland (1991), Faulds (1993), Lister and Baldwin (1993), Sawyer and others (1994), and Minor (1995). Thus we conclude that extensional faulting in the southern Great Basin and Marysvale field took place throughout the episode of calc-alkaline magmatism, but either could dominate in any one place, depending on whether faults or passive intrusions took up the stress. Where they both occur in a local area, they will be the same age. Also, basement structures probably deflected stress trajectories and encouraged heterogeneous deformation. In other words, magmatism is not due to faulting nor vice versa; both result from extension in the brittle upper crust and they may or may not occur in the same place. As subduction proceeded, the extension of crust beneath the Basin and Range province resulted in different combinations of normal, oblique, strike-slip, and even reverse faults of different dip angles (all depending on local stress conditions) and in the generation and shallow emplacement of magma.

LATE CENOZOIC FAULTS AND BASIN-FILL DEPOSITS

In contrast to the poorly exposed middle Cenozoic faults in the Marysvale volcanic field, the evidence for late Cenozoic faults is strong. Faults of this episode largely postdate 10 Ma and produced the present topography of fault-block plateaus, ranges, and basins (Stewart, 1971). They are called basin-range faults following the definition of Gilbert (1928) and the spelling of Mackin (1960). The best evidence for the age of the main phase of basin-range faulting in the Marysvale area comes from the rhyolite in and north of Kingston Canyon, an antecedent canyon that cuts west through the southern Sevier Plateau (Rowley and others, 1981). Here rhyolite of about 8 Ma that caps the plateau predated the main part of basin-range extension. Canyon cutting took place during and after uplift of the plateau fault block at least 2,000 m along the Sevier fault zone. A rhyolite volcanic dome was emplaced in the bottom of the canyon at 5 Ma, after main-phase faulting. Abundant smaller basin-range faults in the area, however, are as young as Quaternary.

Grabens formed closed basins during basin-range extension in the eastern Great Basin and the transition zone, and they were filled with poorly to moderately consolidated, mostly clastic sedimentary rocks. Most basins in the Great Basin started to form at about 10 Ma (Zoback and others, 1981; R.E. Anderson and others, 1983; R.E. Anderson, 1989). The Beaver Basin near the western edge of the Marysvale field is typical. The upper part of its fill contains soil horizons, air-fall tuff, basalt flows, and fossils that constrain its age between Pliocene and Quaternary (R.E. Anderson and others, 1978; Machette and others, 1984; Machette, 1985; J.J. Anderson and others, 1990b). The beds in the lower part of the basin are not exposed, but the age of 9 Ma for the rhyolite of Gillies Hill, which is emplaced along boundary faults on both sides of the basin (Evans and Steven, 1982), and the age of 10 Ma for basalts adjacent to the basin in the Cedar City area (R.E. Anderson and

Mehnert, 1979) suggest that the basin began to form then. In most of the High Plateaus, where drainage is integrated, basin-fill deposits of the Sevier River Formation (Callaghan, 1938; J.J. Anderson and Rowley, 1975; J.J. Anderson, 1987; Rowley and others, 1988a, b) contain tuff beds and basalt flows ranging from 14.2 to 7.1 Ma (Steven and others, 1979; Best and others, 1980). The Sevier River Formation predates basin-range faults. Rowley and others (1979, p. 16; 1981, p. 600) suggested that lower parts of the formation were deposited in basins formed by broad warping because we were unable to find basin-bounding faults of that age. Overlying basin-fill deposits that formed synchronously with basin-range faults are probably Pliocene and Quaternary.

Most basin-range faults in the Marysvale volcanic field are high-angle normal faults that strike from north-northwest to north-northeast. Many of these faults are subvertical at the surface and may represent movement on older tensional fractures (E.M. Anderson, 1951) that formed perpendicular to a horizontal σ_3 when the amounts of stress in the other two axes (σ_1 and σ_2) were equal (J.J. Anderson and Rowley, unpub. data, 1992). Alternatively, the faults are subvertical where seen because they are close to their original surface of erosion, where σ_1 equals σ_2; only at greater depth did lithostatic pressure increase to produce a greater stress (vertical σ_1), and there the faults dip at closer to their ideal 60n., 1994). Tensional fractures related to the faults probably supplied subvertical feeders for the many basalt flows in the eastern Great Basin and the High Plateaus. Some faults may represent movement on older subvertical torsional fractures formed during twisting due to uneven amounts of vertical uplift on various parts of some fault blocks (Rowley, 1968). In most places in the area, the trend of σ_3 apparently was generally eastward, and σ_1 was vertical in late Miocene to Pliocene time. Strain, however, was heterogeneous and complex, and it has not been studied in detail. R.E. Anderson (1986) and R.E. Anderson and Barnhard (1992) ascribed a north-northeast-striking zone of left slip in the southern Pavant Range and Sevier Valley to east-west σ_3 and variation between north-trending σ_1 and σ_2, similar to relationships mentioned above for the Caliente area and the Nevada-Utah-Arizona border area. Similar complications are indicated by left slip along the north-northeast-striking Hurricane fault zone near Cedar City (R.E. Anderson and Mehnert, 1979) and the north-northeast-striking Paunsaugunt fault zone in Johns Valley (R.E. Anderson and Barnhard, 1993b).

On the basis of mapping by Rowley in the Caliente area, the third and youngest episode of extension, that of basin-range normal faulting, formed north-trending basins and ranges there (Rowley, Snee, and others, 1992). The explanation there seems relatively simple: namely, east-trending σ_3 and vertical σ_1. In the northeastern Great Basin, the extension direction is east-northeast and east (Pierce and Morgan, 1992, fig. 22). Overall in the Great Basin, Zoback and others (1981) interpreted a general west-northwest direction of σ_3, partly because many basin ranges trend north-northeast. More ranges trend north, however, and partly for this reason, an east-west extension direction is more common This direction represents a clockwise change with time of about 30°–45° from the σ_3 direction for pre-basin-range faults. The change took place when extension continued but resulted from an entirely different tectonic picture, that of oblique extension (Hamilton and Myers, 1966) of the Great Basin resulting from right slip along the San Andreas transform and along parallel fault zones of significant right-slip transform motion as far east as the Walker Lane belt. This idea is not much different from the concepts of Hamilton and Myers (1966), Atwater (1970, fig. 14), Christiansen and McKee (1978), and McKee and Noble (1986), which in turn relate to suggestions by Wise (1963) that the Western United States is a megashear. The Great Basin became a rift, with raised shoulders—the Sierra Nevada and Wasatch Front—on its western and eastern sides, respectively (Eaton, 1982). Such shoulders are typical of rift margins and may be related to isostatic responses in the lithosphere due to lithosphere changes (Schmidt and Rowley, 1986) or to unloading of footwalls by normal faults (May and others, 1994).

IGNEOUS BELTS AND MINERAL BELTS

Magmatism and mineralization in the Marysvale volcanic field were partly controlled by long-lived regional igneous and tectonic features. The main regional feature is the Pioche-Marysvale igneous belt, which was originally called the Pioche mineral belt by Shawe and Stewart (1976). They also noted a parallel subbelt to the south, which they called the Delamar–Iron Springs mineral belt. Most igneous rocks of both belts are middle Cenozoic calc-alkaline rocks, but some are upper Cenozoic high-silica rhyolite. We have modified the names to call them igneous belts because virtually all the mineral deposits that Shawe and Stewart plotted were localized by intrusive rocks, which we regard as cupolas on batholiths that largely underlie the belts (Steven and others, 1984). The intrusive rocks probably have a volume considerably greater than that of the overlying volcanic rocks. A north-south partial cross section across the Pioche-Marysvale batholith complex is exposed in the Mineral Mountains, where basin-range uplift as a core complex (Price and Bartley, 1992) led to exposures of 25- to 9-Ma plutons (Nielson and others, 1986; Coleman and Walker, 1992). Magmatism in the Pioche-Marysvale belt generally migrated eastward over time (Steven and others, 1984). The plutons of the Pioche-Marysvale and Delamar–Iron Springs igneous belts are clearly delineated by short-wavelength aeromagnetic anomalies, and thus we have modified Shawe and Stewart's shape of the two belts (fig. 1), but not their overall trend, to reflect these aeromagnetic data (Zietz and others, 1976, 1978; Mabey and others, 1978; Hildenbrand and Kucks, 1988a, b; Blank and Kucks, 1989).

The Pioche-Marysvale and slightly younger Delamar–Iron Springs igneous belts (fig. 1) are responsible for only some of the short-wavelength aeromagnetic anomalies in the Great Basin that are interpreted to indicate the presence of igneous rocks. Maps of mineral belts by Shawe and Stewart (1976) and Bagby (1989) document other east- and east-northeast-trending belts in the eastern half of the Great Basin, as well as complex, dismembered, northwest- to west-northwest-trending belts in the Walker Lane belt. These alignments of mining districts, which reflect igneous centers (most of them middle Cenozoic), are also shown by aeromagnetic anomalies (Stewart and others, 1977; Zietz and others, 1976, 1978; Hildenbrand and Kucks, 1988a, b; Blank and Kucks, 1989; Viki Bankey, unpub. data, 1992). The two halves of the Great Basin are separated by a north-trending belt called the "quiet zone" by Stewart and others (1977), which is characterized aeromagnetically by rocks of low magnetic intensity and relief (fig. 1). The shape of the aeromagnetic anomalies indicates a bilateral symmetry to the Great Basin about the quiet zone, and this symmetry is even more obvious in the regional (long wavelength) gravity field, the topography, and other parameters, but not in the exposed geology (Eaton and others, 1978). The mostly east-northeast-trending belts in the eastern Great Basin parallel the middle Cenozoic extension direction of Zoback and others (1981), as do eastern parts of some belts just west of the quiet zone (fig. 5 of Stewart and others, 1977). Thus all belts probably are remnants of middle Cenozoic magmatism.

The Pioche-Marysvale and Delamar–Iron Springs igneous belts are integral parts of a great east-trending arcuate swath of middle Cenozoic igneous rocks, which has been described by Stewart and others (1977) and Best, Christiansen, and others (1989) to extend from west of Reno to Marysvale. It contains the most significant Oligocene and early Miocene igneous centers in the Great Basin. Best (1988) and Sullivan and others (1991) noted that the Colorado Plateau laccoliths and the San Juan volcanic field seem to be an eastward extension of the swath, and they referred to the overall belt as the Reno–San Juan magmatic zone. This zone reflects not only intensity of volcanic activity, but also time; it is only the 31- to 20-Ma part of the overall Great Basin calc-alkaline volcanic province, and represents the "ignimbrite flareup" of peak volcanism in the Great Basin (Best and Christiansen, 1991). Parallel igneous belts in the Great Basin are progressively older (as old as Eocene) to the north and northeast and younger to the south and southwest (Stewart and others, 1977; Cross and Pilger, 1978; Christiansen and Yeats, 1992). This pattern of migration of magmatism is but one of several in the west (Cross and Pilger, 1978; fig. 4; R.E. Anderson, 1989, fig. 6; Best, Christiansen, and others, 1989, fig. 2), some of which diverge from others at high angles. When taken together, these complicated patterns are confusing. The southward and southeastward migration of igneous belts or zones in the Great Basin, which is at an angle to an expected eastward to northeastward subduction trend, has been explained as a progressive steepening or lateral foundering (first in the north, then progressing southward) of the northern part of a subducted slab of oceanic lithosphere (Best and Christiansen, 1991, fig. 2); such foundering allowed a southward-moving wedge of asthenosphere (Hamilton, 1995, figs. 5–6) to move upward and to generate mafic magmas, apparently by supplying heat to melt the dewatering slab of crustal material (Lipman, 1980, fig. 14b, c; Best and Christiansen, 1991, fig. 2). This explanation is one of several (Cross and Pilger, 1978; Stewart, 1983; Henderson and others, 1984; Gans and others, 1989) that have been proposed.

The igneous belts in the Great Basin, which are younger southward, are accompanied by mostly northerly striking normal faults of the same age as the magmatism. This idea, stated by Gans and others (1989), Dilles and Gans (1995), and Scott, Unruh, and others (1995), can be further demonstrated by mapping in progress by R.B. Scott, R.E. Anderson, and Rowley in southeastern Nevada. The middle Cenozoic calc-alkaline igneous belts in the Great Basin have generally been explained by a petrogenetic model involving a subduction zone dipping east-northeast beneath the Western United States (Stewart, 1983). Although this subduction model is not universally accepted, it finds support in the igneous geochemistry, which indicates subduction signatures (Gill, 1981; Fitton and others, 1991). Furthermore, when we look at the larger picture, we note that the volcanic rocks of the Mogollon-Datil field in Arizona and New Mexico (McIntosh and others, 1992) and the Sierra Madre Occidental field in Mexico (Ward, 1991) have ages and igneous compositions similar to those of the Reno-San Juan magmatic zone (Armstrong and Ward, 1991). Thus the various trends of southward younging calc-alkaline activity in the Basin and Range province are smaller patterns within the more regional pattern of volcanism in North America and elsewhere along the Pacific rim. These smaller patterns seem to be most abundant where this global belt is anomalously wide, and their explanation is tied probably to the combination of processes responsible for the large width of the calc-alkaline province in the West. Combining the geochemical signatures linking these rocks with subduction and the overall tectonic evidence, we favor an origin for all these igneous rocks related ultimately to a subduction zone dipping eastward to northeastward beneath the continent. The flat subducted slab that existed during Sevier and Laramide deformation and during the early Tertiary apparently slowed, shortened, and perhaps steepened in the middle Cenozoic as the East Pacific Rise approached the trench, and thus the oceanic lithosphere became younger and less likely to form coherent, long-lasting slabs when subducted (Severinghaus and Atwater, 1990).

TRANSVERSE STRUCTURES

In addition to igneous belts, important tectonic features in the Great Basin that appear to be locally associated with volcanism are long-lived, east-northeast- to east-southeast-striking transverse structures (fig. 1). These have been called “lineaments” by many of us for years but, because many of these structures have now been mapped and therefore evidence for some is finally solid (as opposed to those types of “lineaments” observed only on aerial photos or by remote sensing), the nongenetic term “transverse structure,” from Duebendorfer and Black (1992) and some others, is here adopted. Many transverse structures have been recognized for years by mining and petroleum geologists (such as Fuller, 1964; Hilpert and Roberts, 1964; Roberts, 1964, 1966), and citations to many others were given by Eaton and others (1978, fig. 3–11B), Mabey and others (1978), Rowley and others (1978), Stewart (1983), and Duebendorfer and Black (1992). Use of a nongenetic term is required because we now know that the structures have different types, scales, and origins, but all are similar in that they strike transverse to most Great Basin topographic and geologic features and, importantly, in that they formed parallel to the extension direction of the time. They can be thought of as “continental transform” structures because, on the one hand, they are like transform faults in the ocean basins in that they may form as the result of different amounts and directions of magmatic spreading north and south of them. In the Great Basin, on the other hand, only part of this spreading is magmatic, and most is due to faulting. (R.N. Anderson and Noltimier, 1973, showed that some spreading along ocean-basin spreading centers also is by faults.) Transverse structures terminate abruptly along strike as their displacement is taken up by other structures or when spreading or extension north and south of them is the same. In both transform faults (Wilson, 1965) and transverse structures, actual slip may be opposite to the apparent displacement.

The transverse structures were best documented by Ekren and others (1976), who showed that they include some broad and some concentrated zones of recurrent fault offset and local folding. Most of the transverse structures began in the Oligocene, though some probably were late Mesozoic, and most are so young that basin ranges are offset or terminated along them. In other places along strike, the structures mark terminations or interruptions of topographic features and of magnetic or other geophysical anomalies. Tertiary volcanic units terminate or thin across the transverse structures in some places, generally reflecting predepositional topographic highs formed by them. The structures have localized plutons and volcanic centers over a significant period, which implies that they have deep crustal control (Ekren and others, 1976). We would not, however, call them rifts, as Bartley and others (1992) did, nor think of them as spreading centers, as Bartley (1989) did, because they formed parallel to the extension direction.

Offset along the transverse structures is difficult to determine, probably because some are large, long-lived shear zones that are so badly deformed and hydrothermally altered that kinematic indicators were destroyed. Some transverse structures, however, seem to have undergone oblique slip or strike slip, although the sense of strike slip may reverse itself along strike. Strike-slip motion, where present, does not seem to be great in many transverse structures; those structures that cross the northern Nevada rift (see section on “Spreading”) show little strike-slip offset of the rift. Some transverse structures are well-known zones that transfer different amounts and types of extension on either side of them, such as the east-striking, left-lateral Garlock fault of California (Davis and Burchfiel, 1973), the west-northwest-striking right-lateral Las Vegas valley shear zone (fig. 1; Fleck, 1970; R.E. Anderson and others, 1972; Liggett and Childs, 1977), or the related northeast-striking, left-lateral Lake Mead fault system (Duebendorfer and Simpson, 1994; R.E. Anderson and others, 1994) just to the south. Other transverse structures are broad zones of distributed shear, and these may include individual faults that are locally concealed beneath the volcanic rocks that erupted along them (Rowley and others, 1978; Hudson and others, 1993, 1995; Hudson and Rosenbaum, 1994). The term *accommodation zone*, which has been around a long time (for example, Davis and Burchfiel, 1973) and is now applied by Faulds and others (1990) to a tilt-block domain boundary, is an appropriate genetic name for the type of deformation seen in some transverse structures. Perhaps the best example of an accommodation zone is the 75-km-long Black Mountains–Highland Spring Range accommodation zone of Arizona and Nevada, which separates rocks of opposite structural polarity (domino-like west-dipping faults with east-dipping beds north of the zone, and east-dipping faults with west-dipping beds south of the zone). Faulds and others (1990) originally explained this zone as an expression of torsion between two oppositely dipping detachment faults. In a drastic modification of his model, Faulds (1992, 1994) showed that detachment faults and strike-slip faults are not required under the accommodation zone. His zone thus may actually represent *less* deformed rock than the terrains north and south of it, inasmuch as north-striking faults to the north and south die out, and beds flatten, into the zone. In another modification of his model, Faulds, Olson, and Littrell (1994) used “rupture barrier” as a synonym for accommodation zone and suggested that the control on a barrier’s location is a belt of igneous rocks. Not all accommodation zones are transverse structures because some are not transverse to regional structures or parallel to the regional extension direction (Faulds, Olson, and Littrell, 1994), and others may be relatively minor structures. Most transverse structures represent different amounts of spreading (which we define to consist of both faulting and magmatism) north and south of them either by shearing or by scissor-like torsion.

Davis and Burchfiel (1973) and Duebendorfer and Black (1992) explained how transverse structures can terminate along strike or exhibit opposite directions of strike slip along strike. Mapping by Bartley (1989, 1990), Bartley and others (1992, 1994), Overtoom and others (1993), Overtoom and Bartley (1994), and Brown and Bartley (1994) found little evidence of significant strike slip along easterly striking faults within the Blue Ribbon transverse structure, and thus they interpreted them to be normal faults oriented perpendicular to the extension direction. We instead interpret them to be examples of transverse structures along which accommodation was done with little if any strike-slip motion and perhaps even with little shearing, as explained in the models of Davis and Burchfiel (1973), Duebendorfer and Black (1992), and Faulds (1992, 1994). Stewart (1980) and Stewart and Roldan-Quintana (1994) plotted tilt-block domains in the Basin and Range province and showed that transverse structures bound many regions of opposite structural polarity north and south of them. Hurtubise (1994) documented opposite polarity north and south of the Blue Ribbon transverse structure. Detailed mapping by Stoeser (1993) and C.J. Nutt (oral commun., 1994) found that some transverse structures accommodate different amounts and types of low-angle attenuation and detachment faulting north and south of them.

Eaton and others (1978) and Eaton (1979a) suggested that transverse structures are transform faults oriented perpendicular to a possible north-northwest-striking spreading ridge that was coaxial with or near the quiet zone and around which the bilateral symmetry developed in the Great Basin (see next section); they concluded that the transverse structures are analogous to transforms perpendicular to spreading ridges in ocean basins. We find that, during the middle Tertiary, transverse structures formed parallel to the igneous belts and probably localized the igneous belts, but the transverse structures now strike east and cut the igneous belts at angles of as much as 30° because of the clockwise change in σ_3 noted by Zoback and others (1981). Because they formed parallel to the subduction direction, perhaps transverse structures formed above, and were localized by, oceanic fracture zones (that is, fossil transform faults; Menard and Chase, 1970) that formed in the Farallon/Vancouver plates and then were subducted under the Western United States. Severinghaus and Atwater (1990, figs. 8–13) suggested that such fracture zones controlled the geometry of subducting slabs and that by about 10 Ma, the eastward projection of these fracture zones under the Western United States had changed from east-northeast to east, like the overlying transverse structures.

Whatever the origin of the transverse structures, by the late Cenozoic they were oriented east, generally parallel to the late Cenozoic extension direction, and most functioned to accommodate adjacent areas that had different amounts of extension. In other words, they separated domains of different geologic tilting, extension, magmatism, and history. In the same way that the bases of strike-slip faults require some kind of subhorizontal plane of uncoupling (Stone, 1986) to allow their movement, so also do many low-angle faults in the crust require some kind of bounding subvertical fault zone to allow their movement. Transverse structures are to a degree analogous to tear faults (parallel to the transport direction) in hanging walls of low-angle faults (see examples in R.E. Anderson, 1971), but most are major features, so the underlying subhorizontal zone of detachment into which they feed is probably the brittle-ductile transition zone in the crust. Detachment, attenuation, and high-angle faults are lesser features bounded by transverse structures.

One transverse structure in the Marysvale volcanic field is the Blue Ribbon transverse structure (formerly called the Blue Ribbon lineament). Rowley and others (1978) concluded that this structure is about 25 km wide (north-south) and 280 km long (east-west) in Utah-Nevada and connects across the quiet zone with the Warm Springs “lineament” in Nevada (Ekren and others, 1976), for a combined length of about 600 km. Hurtubise (1989, 1994) provided valuable data about the feature by mapping it across the quiet zone east of where Ekren and others (1976) showed it, and he called it the Silver King lineament. We retain the name Blue Ribbon transverse structure for the part of the overall zone that is east of Ekren’s Warm Springs transverse structure. The feature extends eastward through the southern side of the Marysvale volcanic field (figs. 1, 3), where Rowley, Mehnert, and others (1994) interpreted it to control features as different in age as the string of stratovolcano vents (32(?)–21 Ma) for the Mount Dutton Formation and several rhyolite centers (9–5 Ma). Another transverse structure in the Marysvale field is the east-striking Cove Fort transverse structure, which is at least 150 km long. It contains the east-striking Clear Creek downwarp between the Tushar Mountains and Pavant Range. It also includes the east-striking Cove Creek fault of Steven and Morris (1983), which is marked by hot springs (Ross and Moore, 1994); an extensive zone of faults, fractures, and hydrothermally altered rocks; and down-to-the-south, pre-Osiris Tuff offset. The transverse structure defines the northern side of the igneous centers of the Pioche-Marysvale igneous belt, and, like most transverse structures, it has strong geophysical expression on both regional (Zietz and others, 1976; Blank and Kucks, 1989; Saltus and Jachens, 1995; Viki Bankey, USGS, unpub. data, 1992) and local (Campbell and others, 1984; Cook and others, 1984) scales. Additional evidence that the Cove Fort transverse structure has first-order structural significance was the recognition by T.A. Steven (written commun., 1985) of what he called the “Sevier River oval,” just north of the transverse structure. He defined this as a north-northeast-striking oval-shaped domal uplift about 115 km long, whose eastern side is the central and southern Canyon Range, Valley Mountains, and Pavant Range; whose center is the southern Sevier Desert and Black Rock Desert; and whose western side includes the Cricket Mountains (fig. 3). The Sevier River oval has strong geophysical expression (Thompson and Zoback, 1979; Saltus and Jachens, 1995). Steven interpreted it to be a late Miocene and younger, incipient (little eroded) metamorphic core complex. He interpreted the west-dipping Sevier Desert detachment of McDonald (1976), which has been imaged by seismic reflection data west of there (Allmendinger and others, 1983; Von Tish and others, 1985), to be genetically related to the core complex. Otton (1995) may have found the “breakaway” scarp of this detachment on the western side of the Canyon Range, and he has dated large rock avalanches shed from the scarp at 13–12 Ma.

In previous discussions of Marysvale geology, we have not stressed the significance of transverse structures. However, recent mapping of the Caliente caldera complex near the western edge of the Delamar–Iron Springs igneous belt by Rowley has emphasized their significance (Best and others, 1993). Thus, for example, the east-striking Timpahute “lineament” of Ekren and others (1976), which they recognized to be at least 140 km long, extends another 30 km east to the Utah border and defines the northern side of the Caliente caldera complex, the Chief mining district, plutons, and other features (Ekren and others, 1977; Rowley, mapping in progress).

Similarly, the newly recognized east-striking Helene transverse structure (Rowley, unpub. mapping, 1992) defines the southern side of the Caliente caldera complex, the major Delamar gold mining district, two smaller gold districts, many east-striking faults, east-striking rhyolite dikes, rhyolite domes, plutons that fed the dikes and domes and localized the gold, a diatreme, and other features for at least 40 km (fig. 1). In fact, the Caliente caldera complex can be thought of as an east-trending volcano-tectonic trough, in the sense of those described by Burke and McKee (1979), bounded by the transverse structures. The Caliente caldera complex thus can be considered to be a smaller version of an igneous belt: although it has been recognized for years that most simple calderas in the Great Basin are elongate east-west (Best, Christiansen, and others, 1989) due to later extension, the Caliente caldera complex is much more asymmetrical than most (80 km east-west versus 35 km north-south). This is because recurring, simultaneous faulting and magmatism extended the complex parallel to the direction of middle Cenozoic extension. In partly the same way, igneous belts may grow or “spread” parallel to the extension direction by combined north-striking dikes, shallow intrusions, and northwest- to northeast-striking faults in zones bounded by transverse structures.

PLATE-TECTONIC HISTORY

Locally significant Late Cretaceous and early Cenozoic calc-alkaline magmatism (Lipman, 1992) and Sevier and Laramide deformation are interpreted to be the results of rapid subduction along a shallow east- to northeast-dipping slab of oceanic lithosphere that reached as far east as Colorado (see, for example, Atwater, 1989; Severinghaus and Atwater, 1990; Helmstaedt and Schulze, 1991). The stress regime at this time consisted of generally east-west compression, yet transverse structures probably started forming by Laramide and Sevier time.

Starting in the middle Cenozoic (Eocene to Miocene), voluminous calc-alkaline magmatism became widespread in the Western United States as far east as Colorado (Lipman and others, 1972), but the origin of the magmas and their plate-tectonic setting are much less clear than is the plate-tectonic setting of the West during the Late Cretaceous and early Cenozoic. Sevier- and Laramide-age east-west compression in the Great Basin had changed to east-northeast extension during middle Cenozoic calc-alkaline magmatism. The subduction model requires that the shallow-dipping, obliquely subducting slab continued to reach as far east as Colorado (Atwater, 1989; Severinghaus and Atwater, 1990), but some workers rejected the model and, following Coney and Reynolds (1977), concluded that the slab steepened through time and did not continue far inland. We endorse a subduction model (Atwater, 1989) in which the slab was far inland in the early and middle Cenozoic, then shortened and perhaps steepened throughout the middle Cenozoic. Regardless of the model, thermal softening of the lithosphere may have allowed diapir-like masses of asthenospheric mantle (Eaton, 1979a) to follow various upward paths, influenced perhaps in part by structures in the lithosphere (Lipman, 1980, 1992), by warps in the slab (Best and Christiansen, 1991), or by foundered pieces of the subducted slab. Warps and foundered pieces of the slab, in turn, may have been localized by the dominant structures of the slab, namely oceanic fracture zones (Menard and Chase, 1970) and transform faults. Severinghaus and Atwater (1990) projected these structures east-northeast from the Pacific plate under the North American plate, and some evidence for this interpretation comes from teleseismic *P*-wave images that reveal a northeast grain to the upper mantle under the Great Basin (Humphreys and Duecker, 1994, fig. 7). The transverse structures and the bilateral symmetry of the Great Basin both developed largely during the middle Cenozoic, although they probably began to form earlier. Nd isotope data indicate that voluminous middle Cenozoic calc-alkaline magmas throughout the Western United States and Mexico contain a major, commonly dominant, mantle component; this component most likely was derived by fractional crystallization of mantle-derived basaltic magmas and assimilated crustal rocks melted by the heat of the basaltic magmas (Smith, 1979; Gans and others, 1989; Fitton and others, 1991; Johnson, 1991; Coleman and Walker, 1992; Lipman, 1992; Wiebe, 1993). After about 28 Ma, the subduction system began to be replaced by the San Andreas transform fault system, and subduction rates slowed as the Farallon/Vancouver plates were progressively consumed (Atwater, 1970; Christiansen and Lipman, 1972; Best and Christiansen, 1991).

Beginning no later than 23 Ma, compositionally bimodal magmas (Christiansen and Lipman, 1972) that lack geochemical subduction signatures were emplaced in the Marysvale area. This change in the Basin and Range province broadly coincided with progressing diminution of the Farallon/Vancouver plates west of the province as the San Andreas transform fault expanded northward and southward along the western edge of the continent; transform motion was accommodated by other faults parallel to the San Andreas fault, extending east to include the Walker Lane belt (Atwater, 1970; Eaton, 1980), which defines the western side of the Great Basin. Subduction progressively ceased northward in the Western United States except under the Cascade Range as the transform lengthened (Atwater, 1970; Christiansen and Lipman, 1972; Christiansen and McKee, 1978; Dickinson and Snyder, 1979). The Great Basin in the late Cenozoic underwent oblique extension as a broad megashear (Atwater, 1970, fig. 14), resulting in bimodal magmatism that began locally before 20 Ma and in basin-range deformation that culminated sometime after 9 Ma in the Marysvale area and most other areas; the Great Basin became a very broad rift. Extension was oriented east to east-southeast on average, though in some places, it was in a somewhat different direction. Some faulting continued along the previously formed transverse structures (Ekren and others, 1976), which now took on a generally eastward strike, synchronous with north- to north-northeast-striking normal faults that were concentrated along and near the older northern Nevada rift (Wallace, 1984; Blakely, 1988; Blakely and Jachens, 1991; Catchings, 1992) and were broadly distributed elsewhere through the Basin and Range province.

SPREADING

We propose a variant of the subduction model for the middle Cenozoic rocks of the Great Basin that draws on the ideas of Scholz and others (1971), Proffett (1977), Eaton and others (1978), Eaton (1979a, 1982, 1984), and Mabey and others (1978). These geologists and geophysicists proposed that the calc-alkaline igneous rocks in the Great Basin represent magmatic spreading about a single axis, during subduction of the Farallon/Vancouver plates, leading to the bilateral symmetry in the Great Basin. Upwelling asthenosphere underlying the symmetry axis was implied (Scholz and others, 1971). In this model, the Great Basin was underlain by a north-northwest-trending spreading axis, oriented perpendicular to the subduction direction, and hydrothermal activity along the axis altered the ferromagnesian minerals in the rocks and reduced their magnetic susceptibility, thereby forming the magnetically quiet zone. Supposedly, as the elongate asthenosphere diapir hit the base of the crust, it spread out horizontally in both directions, and outward-directed tractional stresses on the crust led to extensional faulting and bilateral symmetry. The east-striking transverse structures in this model represent transform faults that strike perpendicular to the spreading axis (Eaton and others, 1978); the intrusive bodies of the igneous belts were emplaced along these faults.

A north-northwest-trending aeromagnetic high just west of the magnetic quiet zone was noted by Mabey and others (1978), who suggested that it represents a ridgelike product of arc spreading. Zoback and Thompson (1978) called the aeromagnetic high the northern Nevada rift (fig. 1) and noted that (1) it is filled with 17- to 14-Ma basaltic and rhyolite dikes, and (2) its trend, like that of normal faults and dikes in tension fractures, is perpendicular to the middle Cenozoic direction of extension. McKee and Noble (1986), Blakely (1988), Blakely and Jachens (1991), and Zoback and others (1994) described the rift in greater detail and extended it to southern Nevada, near the geographic axis of the middle Cenozoic Great Basin. R.E. Anderson and others (1994) extended it into the Lake Mead area of southern Nevada and adjacent Arizona and concluded that here the rift had at least 15 km of extension. It may pass northward into Oregon; Stewart and others (1975) connected it with the west-northwest-striking Brothers fault zone of Oregon and called it the Oregon-Nevada lineament, whereas Pierce and Morgan (1992) called it the Nevada-Oregon rift zone. But because it is best exposed in northern Nevada and its continuation into Oregon is in doubt (Mabey and others, 1978), we will resist the temptation to rename it. The northern Nevada rift is cogenetic with the McDermitt caldera, at its northern end, and the Columbia River Basalt Group, north of it (Zoback and others, 1994). Christiansen and Yeats (1992) suggested that the north-northwest-striking western Snake River Plain graben was formerly the northern part of the northern Nevada rift but that the graben was offset 100 km by an east-northeast right-lateral fault or transform fault that underlies the eastern Snake River Plain. The northern Nevada rift and its parallel features are similar in size to the Rio Grande Rift, but are less eroded. On the basis of paleomagnetic data, Li and others (1990) suggested counterclockwise rotation of the rift, but this interpretation was rebutted by Zoback and others (1994).

A setback to the spreading model of Scholz and others (1971) came from Blakely (1988), who concluded from analysis of geophysical data that the low magnetic susceptibility of the rocks in the quiet zone is not due to former high heat flow but may be due instead to a low proportion of magnetite relative to ilmenite in igneous rocks of the quiet zone. Certainly an interpretation for spreading of a type similar to sea-floor spreading, involving convective upwelling of magma along a single mid-oceanic type ridge spreading center, remains speculative and unlikely (R.J. Blakely and E.H. McKee, oral commun., 1992).

In our variant of the model of Scholz and others, we follow Hamilton (1988a, 1989) and perhaps others and use "spreading" in the sense of a widening of the Great Basin by a combination of pure-shear extension and voluminous magmatism about many axes perpendicular to the extension direction during both the middle and late Cenozoic. The axes include, in addition to the northern Nevada rift, other parallel magnetic anomalies and zones of seismic activity and heat flow that have been described by Wallace (1984), Blakely (1988), McKee and Blakely (1990), Blakely and Jachens (1991), and Catchings (1992) in northern and central Nevada and interpreted by them to indicate zones of concentrated extension. We suggest that, whether by faults or by intrusions, heat flow increased (Lachenbruch and Sass, 1978) under these axes and the brittle-ductile transition zone (below the seismogenic crust; Sibson, 1989) rose and arched under them to form blisterlike forms (metamorphic core complexes) or ridgelike forms. Perhaps with continued extension under these axes, the brittle-ductile transition zone rose higher and broadened perpendicularly outward, providing a driving force for extension and magmatism outward in both directions. Either the crust at and above the raised brittle-ductile transition zone softened, leading to a reduction in shear strength, or stresses dropped after extension along the axis (see Mandl, 1987) and thus extension ceased. East-west least principal stress continued in the Great Basin, however, and thus extension migrated generally toward the cooler, more brittle crust at the margins of the Great Basin. Hamilton (1985, 1988c) pointed out that spreading in North Island, New Zealand (Stern, 1985), may be similar to that in the Great Basin, and that, according to Hamilton, spreading in both areas may actually be due to extension outpacing magmatism. The result in New Zealand is crustal thinning and subsidence, the opposite to the development of most continental magmatic arcs.

Evidence exists for Cenozoic magmatic spreading that generally migrated progressively from northerly striking centers near the central Great Basin outward to its western and eastern boundaries (the motion is relative to the centers, if we hold the North American plate motionless with respect to the Pacific plate) (Eaton, 1980). Early compilation of isotopic dates (Armstrong and others, 1969; McKee, 1971) noted such an outward migration of magmatism and faulting. Later compilations (Armstrong and Ward, 1991) showed that this picture is not that simple, because spreading as it is known in the ocean basins probably does not occur on continents, because a single center at any one time is unlikely, and because younger centers may have been superimposed on or near older ones. The best evidence of progressive migrations of eruptive centers is from the northwestern Great Basin, where MacLeod and others (1976) noted rhyolite domes and centers just southwest of the Brothers fault zone that become progressively younger west-northwest, from 10 Ma in southeastern Oregon to Quaternary in central Oregon. Good data from the eastern Snake River Plain just north of the northeastern Great Basin, summarized by Pierce and Morgan (1992), shows that rhyolite calderas and other eruptive centers there, in contrast, migrated east-northeast from 16 Ma at the Nevada-Oregon-Idaho border to Quaternary at Yellowstone National Park. Other migrations have been documented in the Great Basin, although some of these are subtle, as in northeastern Utah (Miller, 1991) and in the Tintic–Deep Creek igneous belt of west-central Utah (Stoeser, 1993). Farther south, Steven and others (1984) noted that at least the eastern two-thirds of the middle Cenozoic part of the Pioche-Marysvale igneous belt is generally younger (35 to 23 Ma) eastward. On the southwestern side of the Great Basin, Silberman and others (1975) observed andesites that are older (27 Ma) in the east and younger (25 to 20 and 8 to 7 Ma) in the west. Dickinson and Snyder (1979, fig. 8) also recognized a progressive north-to-northwest migration of igneous centers with time (16 to 5 Ma). More details on these northwest migrations of rhyolite centers were noted by Luedke and Smith (1981), Sawyer and others (1994), and A.M. Sarna-Wojcicki (written commun., 1994), starting with eruptions from the 15- to 11-Ma calderas of the Nevada Test Site (Sawyer and others, 1994), the 9-Ma Black Mountain caldera (Sawyer and others, 1994) 5 km northwest of the Test Site, the 7.5-Ma Stonewall Mountain volcanic center (Weiss and others, 1993; Sawyer and others, 1994) 40 km farther northwest, the 7- to 3-Ma Mount Jackson dome field (Weiss and others, 1993) 20 km farther west, the 6-Ma Silver Peak volcanic center (Robinson and others, 1968; Sawyer and others, 1994) 40 km farther northwest (fig. 1), and ending with the Pliocene and Quaternary Long Valley–Mono Basin volcanic center (Bailey, 1989; Bursik and Sieh, 1989) 80 km farther west-northwest, in California. Thompson and others (in press) documented a westward and northwestward migration of 8- to 3-Ma volcanic rocks and post-6.5-Ma faults in the Death Valley area. Except for young basalts and faults that form an axial fracture in the central Great Basin (Smith and Luedke, 1984; Wallace, 1984; Blakely, 1988; Fitton and others, 1991), it has long been known that the youngest basalts in the Great Basin occur on its eastern and western boundaries (Christiansen and McKee, 1978; Eaton and others, 1978; Eaton, 1979b; Smith and Luedke, 1984; Fitton and others, 1991). Within these Quaternary basalt fields at the eastern boundary, Luedke and Smith (1978) and Nealey and others (1994) noted an eastward migration with time, as has been observed at the eastern edge of the southern Basin and Range province in northern Arizona (Best and Brimhall, 1974; Tanaka and others, 1986; Condit and others, 1989).

Spreading due to faulting probably also migrated outward from centers in the central Great Basin, and this scenario is certainly consistent with the current high seismicity along the basin's western and eastern boundaries (Christiansen and McKee, 1978; Eaton, 1979b). Evidence within the basin, however, is poor, with one main exception: Hamilton (1988b) showed a westward younging of termination of major extension, from middle Miocene in the Nevada Test Site, to late Miocene in the Bullfrog Hills (fig. 1) and in the Funeral Mountains of California, to still active in the Death Valley of California. Also, on the eastern side of the southern Basin and Range province in northern Arizona, Faulds, Gans, and Smith (1994) noted progressively younger faulting eastward (16 to 10 Ma).

Formerly, various names were applied to the type of middle Cenozoic calc-alkaline magmatic arc in the Western United States that contains components of spreading or extension. The terms "ensialic back arc" or "marginal basin" were most commonly used, but these seem inappropriate because most back arcs (as described by Saunders and Tarney, 1984) are underlain by oceanic lithosphere and contain thick marine sedimentary rocks and local basalt rather than subaerial calc-alkaline volcanic rocks (see, for example, Rowley and others, 1991; Rowley, Kellogg, and others, 1992). Furthermore, when applied to the Western United States, such terms leave unconsidered the location of the concurrent magmatic arc west of any back arc. Thus Elston (1984) instead proposed "extensional orogen," but we prefer the name "magmatic arc." Neglecting such semantic matters, however, the point to be made here is that middle Miocene calc-alkaline magmatism and late Cenozoic bimodal magmatism in the Great Basin are associated with significant extension, possibly along many north-northwest- to north-northeast-trending, perhaps diffuse, central axes, and thus a spreading analogy can be made. Thus the Great Basin extended much more than adjacent areas on the Pacific rim.

Spreading was partly accommodated by transverse structures that formed parallel to the spreading direction, that occur throughout the Great Basin, and that partly bound the basin on the north (Christiansen and McKee, 1978; McKee and Noble, 1986, fig. 6) and the south (Davis and Burchfiel, 1973). Accommodation by transverse structures allowed igneous belts to grow parallel to the spreading direction along centers oriented perpendicular to the spreading direction, whereas areas north and south of the igneous belts spread by faulting.

A model for the development of igneous belts is provided by the eastern Snake River Plain. This broad lava plain is generally shown on physiographic maps to bound the Great Basin on the northeast, but geologically it is an uneroded version of the igneous belts within the Great Basin, and to its north, basin-range topography continues (Eaton and others, 1978; Kuntz and others, 1992). The plain is oriented east-northeast to northeast, parallel to the extension direction in the area (Kuntz, 1992; Kuntz and others, 1992). Southward from the eastern Snake River Plain, this direction has gradually swings about 40-west extension direction of the northeastern Great Basin (Zoback and others, 1981; Zoback, 1989; Pierce and Morgan, 1992, fig. 22). Most rocks at the surface of the plain are Quaternary basalt lava flows that erupted along north-northwest-striking linear vents; basin-range faults north of the plain strike in the same direction, whereas those south of the plain strike about north, all being perpendicular to the extension direction (Kuntz and others, 1992, fig. 11). Calderas or other vents generally below the surface of the plain erupted mostly rhyolite ash-flow tuffs and record the progressive east-northeast to northeast migration of eruptions, from 16 Ma to Quaternary (Pierce and Morgan, 1992). In the same way that we would not give the name "rift" to transverse structures, we agree with Kuntz and others (1992) and Pierce and Morgan (1992) that Hamilton's (1989) use of "rift" for the eastern Snake River Plain is inappropriate; the plain was oriented parallel, not perpendicular, to the extension direction when it formed. We speculate instead that the eastern Snake River Plain is bound on its north and south sides by buried east-northeast- to northeast-striking transverse structures. No faults of the same strike are exposed, although they have been modeled by Sparlin and others (1982). The transverse structures that we propose may not be expressed primarily by faults. These zones allowed east-northeast spreading to be taken up primarily by magmatism within the plain, and primarily by normal faulting to the north and south (Kuntz and others, 1992). The plain is thus an igneous belt formed by basalt feeder dikes oriented perpendicular to the spreading direction and by shallow rhyolitic intrusions whose roofs collapsed to form calderas. The eastward progression in rhyolite eruptions here in the (geologically defined) eastern part of the Great Basin is contrasted with the westward progression in rhyolite eruptions in the western part of the Great Basin south of the Brothers fault zone. The Brothers fault zone is a transverse structure that includes right slip and defines part of the northwestern edge of the Great Basin (Lawrence, 1976; McKee and Noble, 1986).

The two transverse structures that we suggest underlie the eastern Snake River Plain continue northeast of Yellowstone into Montana (Eaton and others, 1975; Mabey and others, 1978), parallel to Precambrian structures (Erslev and Sutter, 1990; Pierce and Morgan, 1992, p. 32). One of these structures, the northeast-striking Proterozoic Madison mylonite zone of Erslev and Sutter (1990) in the southern Madison Range, may have been followed by our suggested transverse structure that bounds the northwest side of the plain; farther northeast, the transverse structure may even control the northeast-striking Holocene Emigrant fault of Pierce and Morgan (1992, plate 1) south of Bozeman. Our speculations thus are similar to those of Eaton and others (1975), Mabey and others (1978), and Christiansen and McKee (1978), who suggested that Precambrian crustal flaws controlled the location of the eastern Snake River Plain. We would emphasize, however, that parallelism to the extension direction is the dominant control of the magmatism in the plain. The two transverse structures also continue southwest into Nevada, and the overall 1,000-km-long belt in Montana, Idaho, and Nevada was called the Humboldt zone by Mabey and others (1978) and Rowan and Wetlaufer (1981). The zone is characterized by high heat flow (Lachenbruch and Sass, 1978; Blakely, 1988) of the Battle Mountain high. Pierce and

Morgan (1992, p. 32), however, doubted that the two "lineaments" of Eaton and others (1975) and Mabey and others (1978) were the dominant control of the eastern Snake River Plain. "Why," they asked, "would two such lineaments be exploited simultaneously rather than the failing of one ***?" This remains a good question.

Pierce and Morgan (1992), Zoback and others (1994), and others ascribed the northeast migration of igneous activity in the eastern Snake River Plain and Yellowstone National Park area to the passage of the continental plate over a "hot spot" from an underlying mantle plume. Their evidence is compelling (see also Pierce and others, 1992), but the theory does not explain the synchronous but symmetrically opposite migration along and near the Brothers fault zone (Lipman, 1992, p. 491), as well as other igneous belts in the Great Basin.

Was subduction a prerequisite for volcanism in the Great Basin and nearby areas? Mutschler and others (1987, 1991, this volume) answered "no" and proposed instead that volcanism is due to subcrustal lithospheric thinning and mantle upwelling above passive hot spots. Lipman (1992) and we disagree with their model (see section on Igneous Belts and Mineral Belts, p. xx). Mutschler and his colleagues identified several passive hot spots in the Western United States, including the possible Great Basin spreading center discussed above, which they called the Great Basin regional gravity low. Blank and others (this volume) propose another to underlie the entire Colorado Plateau. Asthenospheric upwarps have been suggested by others, including Eaton (1987), who proposed a large bulge that uplifted the southern Rocky Mountains and formed a crestal graben, the Rio Grande Rift. White and others (1987) traced the development of ocean basins that started to form by upwelling of asthenosphere along continental rifts; decompression of the upwelled material theoretically produced magmas by partial melting. They suggested that the Great Basin is an example of a continental rift that did not stretch sufficiently to form a new ocean basin. According to White and others (1987), if stretching stops before the lithosphere has been reduced to less than half its original thickness, the thinned region subsides to form a sedimentary basin like the North Sea, whereas if stretching continues and the continental lithosphere breaks, new oceanic lithosphere is formed by sea-floor spreading, bordered on either side by rifted continental margins, as in the Red Sea or the Atlantic Ocean. Rosendahl (1987, p. 461) coined the word "pretransform" to describe transverse structures in the East African rift that parallel the extension direction and in the future will become true transforms in oceanic crust when rifting advances to the stage that the area becomes an ocean basin.

CONCLUSIONS

The large Marysvale and San Juan volcanic fields occur on the edges of the Colorado Plateau, where voluminous hot mantle-derived magma formed shallow crustal calc-alkaline magma chambers that congealed to form composite batholiths. In contrast, the laccolith clusters in the rigid Colorado Plateau between these volcanic fields are minor, widely scattered igneous features that had few effusive products. According to Gilbert (1877) and Corry (1988), laccoliths form, whether in the Colorado Plateau or elsewhere, when magma rises to a level in the crust where the rocks have about the same specific gravity as the melt. At about that level, relatively close to the surface (about 2–4 km or less in southeastern Utah, the northern Markagunt Plateau, and the Iron Axis), the magma may spread out along low-dipping incompetent sedimentary beds. In this near-surface environment alone, σ_3 becomes vertical and σ_1 equals σ_2, so magma spreads out laterally and lifts thin roof rocks. Parsons and others (1992), however, argued that rheological boundaries between horizontal packages of host rocks are sites for emplacement of horizontal sills or laccoliths only after the emplacement of vertical dikes (perpendicular to horizontal extension) has relieved tension and caused the least principal stress above this boundary to become vertical, whereas the vertical stress axis below the boundary remains as σ_2. Regardless of the dominating mechanisms, most of the laccoliths of the Iron Axis spread out in gypsiferous shale in the lower part of the Carmel Formation and above the massive Navajo Sandstone, the laccoliths on the southern flank of the Marysvale field spread out in shale in the lower part of the Claron Formation, and laccoliths of the Colorado Plateau and locally in the San Juan area spread out in the Mancos Shale and underlying units. Laccoliths differ from unfloored stocks in having much smaller volumes of magma available for eruption, so attendant volcanic units tend to be smaller, although the products erupted from both sources are similar. Such volcanic rocks of limited extent were erupted from the Spry and Iron Peak laccoliths along the southern side of the Marysvale field (J.J. Anderson and Rowley, 1975; J.J. Anderson and others, 1990a), at some laccoliths of the Iron Axis farther south (Blank and others, 1992; D.B. Hacker, Kent State University, written commun., 1992; Rowley, unpub. mapping, 1992), at laccoliths in West Texas (Henry and Price, 1988; Henry and others, 1991), and in some other areas (Corry, 1988). Laccoliths form some significant ore bodies, as at Iron Springs (Mackin, 1947), but most of their hydrothermal-type deposits are small.

Generally, middle Cenozoic calc-alkaline volcanic rocks similar to those of the Marysvale field are voluminous elsewhere in the Western United States (Johnson, 1991), and their eruptive centers are underlain by even larger batholith complexes. The great breadth of these similar types of igneous rocks is best explained as a result of magmatism above a subduction zone, as elsewhere around the Pacific margin and other parts of the world. The generation of calc-alkaline rocks appears to be driven by "large fluxes of mantle-derived basaltic magmas" (Smith, 1979; Johnson, 1991). The intrusive rocks and related hydrothermal mineral deposits constitute a treasure house of valuable resources. During the middle Cenozoic, extension oriented generally east-northeast was expressed by local high-angle normal faults, many of which flatten at depth, by metamorphic core complexes, and by shallow intrusions. In many places, depending on the stress field, oblique- and strike-slip faults were also produced. Some of the most productive mineral belts in the Great Basin, mostly of gold, silver, and chalcophile elements, resulted where permeability due to faulting and brecciation coicided in space and time with magmatism and resulting hydrothermal convection cells (John and others, 1989, 1991; Hardyman and Oldow, 1991; Henley and Adams, 1992; Rowley, Snee, and others, 1992; Willis and Tosdal, 1992).

Upper Cenozoic bimodal igneous rocks are less voluminous than calc-alkaline rocks, but they also generated many deposits of base and precious metals, as well as lithophile elements. Bimodal volcanism began in the Marysvale area by about 23 Ma and in the Caliente area no later than 15.5 Ma. Bimodal magmatism in the Basin and Range province correlates with transform movement on the San Andreas fault zone and the end of subduction. Most rhyolites probably resulted from partial melting of local areas of crust. Extension in the Great Basin continued, although now oriented east-west or perhaps east-southeast, yet the main late Cenozoic deformational episode that established the present topography, that of basin-range faulting, did not begin until later (after 10 Ma in the Marysvale area and the southeastern Great Basin). Most basin-range faults are relatively high-angle normal faults where exposed, probably decreasing to lower angles with depth, but strike-slip faults occur locally.

The result of volcanism and major extension in the Great Basin during the late Cenozoic and probably the middle Cenozoic can be considered a type of east-west "spreading" or widening of continental crust, creating bilateral symmetry. Numerous northerly trending axes involving different combinations of spreading by faulting and hypabyssal intrusion appear to occur in the central Great Basin but may also be present throughout the subprovince. Long-lasting, recurring combinations of faults and shallow intrusions, as in the northern Nevada rift, Wilson Ridge pluton, and Caliente caldera complex represent examples of local extensional spreading by this twofold combination. Transverse structures oriented east-northeast in the middle Cenozoic, parallel to the subduction direction, represent zones separating different amounts of extension and calc-alkaline magmatism north and south of them. They controlled the shape of the east-northeast-trending igneous belts within the Great Basin. The igneous belts became younger to the south-southeast, but each belt grew eastward or westward (generally westward in the western Great Basin and eastward in the eastern Great Basin) away from a central axis. In the late Cenozoic, during oblique extension that culminated in basin-range deformation, accommodation offset continued along some of the transverse structures, bringing them to their present east-west trend, and some bimodal volcanism also took place along them. The Great Basin continues to spread about north-northeast- or north-trending axes, and most recent basalts and faults occur at the basin's axis and eastern and western margins.

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