Space-time patterns and tectonic controls of Tertiary extension and magmatism in the Great Basin of the western United States

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ABSTRACT

Structural and stratigraphic relations in the Great Basin indicate widespread pre–middle Miocene crustal extension that appears to define two north-trending belts. Most extension in these belts was Oligocene age, but locally it began earlier or lasted into early Miocene time. The eastern belt straddles the Nevada-Utah border and includes the Snake Range, Nevada, area, with its southern end near 37.5°N and its western edge at the Seaman-Butte Mountains breakaway. The southern boundary of the eastern belt is occupied by the 26–15 Ma Caliente caldera complex and approximately coincides with the present east-west–trending margin of the Great Basin north of Saint George, Utah. Crust north of this boundary extended approximately east-west before volcanism began at 30–32 Ma, but to the south, extension began after about 15 Ma. This boundary may have been a rooted zone of left-slip faults that allowed the footwall of the Stampede detachment to move west relative to unextended terrain to the south. The eastern margin of the eastern belt is probably located near the present eastern edge of the Great Basin, but its northern end is poorly defined. The western belt runs from the Funeral and Grapevine Mountains, California, to the Ruby Mountains, Nevada, and north-northeast to the Albion Range, Idaho. Tens of kilometers of crustal extension occurred at least locally in both belts, but magnitude of extension is poorly known for large areas of each.

Tertiary volcanism in the Great Basin began in the north in Eocene time with predominantly effusive volcanism and swept southward, ending with voluminous Oligocene-Miocene ignimbrite eruptions from calderas in an irregular, discontinuous belt between Marysvale, Utah, and Reno, Nevada. A result of the southward migration of volcanism is that the onset of extension in both belts was syn- or post-volcanic in the north but was pre-volcanic in the south. Late Paleogene extension and crustal magmatism coincided in both time and space only locally, where south-migrating magmatism overlapped active north-south–trending extensional belts.

Most calderas in the southern Great Basin formed in previously extended belts or on their margins. Southward migration of ignimbrite sources was apparently blocked by unextended crust to the south. In contrast, volcanism north of the ignimbrite province was dominated by nonexplosive effusion of lava prior to, or during, crustal extension. This is consistent with observations in the southern Basin and Range, where volcanism and crustal extension were generally synchronous, and volcanism was dominantly effusive. Thus, caldera formation may be controlled by the distribution of upper-crustal extension, although the physical mechanism for this control remains speculative.

The space-time patterns of late Paleogene extension in the Great Basin are consistent with extension being triggered by thermal weakening of subducted oceanic lithosphere rather than by effects transmitted from the plate margin, but being driven by gravitational collapse of thick crust. Space-time patterns of Tertiary volcanism in the Basin and Range also appear to conform to patterns of thermal weakening or destruction of the subducted slab. Both active and passive rifting mechanisms are inapplicable on the scale of the extensional belts, because both predict close spatial and temporal association of extension and magmatism, which is not generally observed.

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INTRODUCTION

Time-space patterns of faulting and volcanism provide key constraints on tectonomagmatic models of crustal extension. The southern Great Basin (Fig. 1) is ideally suited for such analysis because of good exposure, well-understood Paleozoic miogeoclinal stratigraphy, and abundant distinctive, regionally extensive Tertiary ash-flow tuffs that serve as time-stratigraphic markers.

The distribution in space and time of volcanic rocks and their sources in the Great Basin is well known. Stewart and Carlson (1976) documented a southward migration of volcanism in the Great Basin, ending with the "ignimbrite flare-up": voluminous caldera-forming eruptions that dominated volcanic activity from 34–14 Ma in the southern Great Basin (Fig. 2). Space-time patterns of caldera formation probably reflect the first-order space-time patterns of crustal magmatism during that interval. Luedke and Smith (1978, 1981) presented maps of post–middle Miocene volcanic sources in the region, and Best and others (1989a) identified sources of major ash-flow tuffs in the Great Basin.

Space-time patterns of extension in the Great Basin are less well known than those of

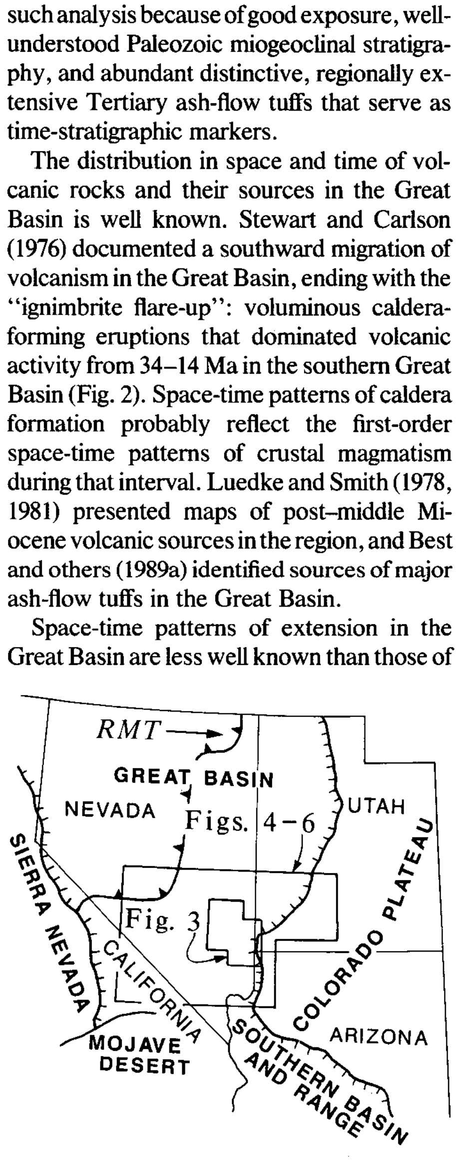

Figure 1. Map of the Great Basin with locations of Figures 3 to 6. The Roberts Mountains thrust (RMT) forms the west limit of widely exposed miogeoclinal strata.

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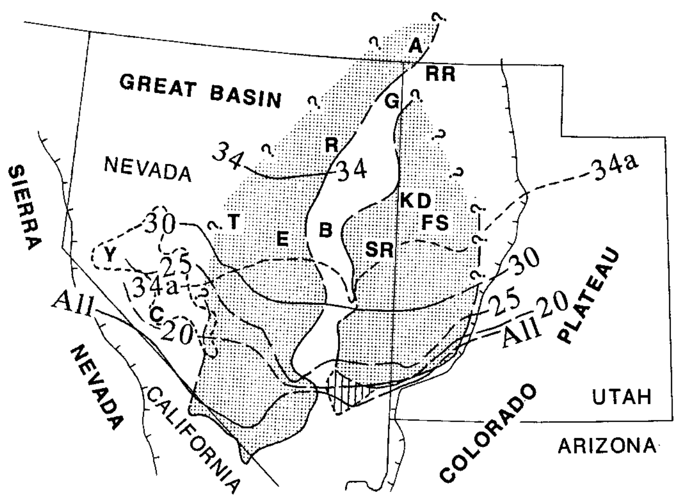

Figure 2. Map of the present distribution of late Paleogene–early Miocene extensional belts defined in this paper (shaded; vertically lined area represents possible continuation of eastern belt). Solid lines depict the southern limit of calderas and are labeled in Ma, except the southern limit of all calderas older than 5 Ma (All); dashed line is the south limit of >34 Ma andesite (34a) (from Best and others, 1989a; Stewart and Carlson, 1976). Localities discussed in text: A, Albion Mountains; B, Butte Mountains; C, Candelaria Hills; E, Eureka district; FS, Fish Springs Range; G, Goshute Range; KD, Kern Mountains–Deep Creek Range; R, Ruby Mountains; RR, Raft River Range; SR, Snake Range; T, northern Toiyabe Range; Y, Yerington District. Area outlined by short dashed line in western Nevada contains localities that probably experienced early to mid-Miocene extension (John and others, 1989).

volcanism, especially for pre-Miocene extensional fault systems that did not form metamorphic core complexes. Such systems are incompletely known due to overprinting by younger faults, scarcity of Mesozoic and early Tertiary deposits, concealment of rocks and structures beneath volcanic cover, ambiguity about the origin of some faults (for example, caldera-related, extensional, or contractional), poor access to military reservations, and the common *a priori* assumption that extensional tectonism did not predate volcanism.

We review available geologic data that bear on the age of onset of extension in the Great Basin and infer the existence of two late Paleogene extensional belts that affected previously thickened crust (Fig. 2). We then focus on the relation(s) between the initiation of extension and major crustal magmatism and suggest that calderas preferentially formed in previously extended crust, whereas effusive volcanism is favored in unextended or actively extending areas. Comparison of this crustal evolution with a thermal model of the subducted oceanic lithosphere under the Great Basin (Severinghaus and Atwater, 1990) suggests that extension and volcanism were triggered by changes in the thermal state of the subducted plate. These interacting factors allowed various disparate timing relations between extension and volcanism to evolve in different areas of the Basin and Range.

EASTERN EXTENSIONAL BELT

Structural Evidence

Extension in the Dry Lake Valley area (Fig. 3) began before volcanism at about 32–30 Ma (Bartley and others, 1988; Taylor and others, 1989). Extension there was probably controlled by the Stampede detachment (Axen and others, 1988), a system of faults that omit stratigraphic section and that are subparallel to bedding in their mildly deformed footwall. Generally east-dipping domino-style normal faults extended upper-plate bedding and tilted it west (for example, Axen, 1986; Lewis, 1987; Taylor, 1990). Displacement on the detachment is poorly known due to a lack of offset markers common to both the upper and lower plates. The Stampede detachment or related structures are exposed in all ranges surrounding Dry Lake Valley (Fig. 3; Bartley and others, 1988; also, Langenheim and others, 1969; Axen, 1986; Lewis, 1987; Sleeper, 1989; Rowley and others, 1990, in press; Burke and Axen, 1990; Taylor, 1990; Burke, 1991; Taylor and Bartley, 1992).

Western Boundary. The Seaman fault forms the western boundary of the Dry Lake Valley extensional belt and is interpreted as the breakaway zone for the Stampede detachment (Figs. 3–5; Taylor and Bartley, 1992). Although concealed by Pliocene to Quaternary fill of southern White River Valley and Miocene volcanic rocks in the southern North Pahroc Range, its existence and Oligocene age are inferred from structural, stratigraphic, and paleogeographic relations (Taylor and Bartley, 1992; Taylor and others, 1989; Taylor, 1989, 1990). The breakaway can be confidently traced north to Fox Mountain and the southern Egan Range (Fig. 4; Taylor and Bartley, 1992).

The western boundary may bend eastward immediately south of the North Pahroc Range (heavy dashed line in Fig. 5), or it may project south (dotted line in Fig. 5), based on isopach maps of ignimbrites (discussed below) and prevolcanic structures. East- to northeast-striking faults with left separation cut pre-Tertiary rocks of the southwest Delamar Mountains and are overlapped by 26–17 Ma tuffs (Figs. 4 and 5; Tschanz and Pampeyan, 1970; Ekren and others, 1977). Their age and sense of separation permit an origin in a late Paleogene sinistral southern boundary proposed below for the eastern belt. West-directed faults thin Devonian strata in this area (Page and Scott, 1991) and are probably older than 27–24 Ma (R. B. Scott, 1991, oral commun.).

Southern Boundary. Pre–32 Ma extension in the Dry Lake Valley region ended 12–16 m.y. before extension began farther south. The Stampede detachment and related structures persist to the north edge of the Caliente caldera complex, where they are intruded by the 25 Ma Cobalt Canyon stock (Figs. 3, 5, and 6; Axen and others, 1988; Rowley and others, 1990, in press). Extension in the Mormon Mountains–Muddy Mountains area south of the Caliente caldera complex began at about 20–15 Ma (Table 1; for example, Bohannon, 1984; Axen and others, 1990).

The Caliente caldera complex (Ekren and others, 1977) apparently formed along the boundary between then-unextended crust to the south and extended crust to the north (Figs. 3 and 6). Voluminous ash flows issued from this complex from about 23 Ma (Condor Canyon Formation) to 18.6 Ma (Hiko Tuff), with smaller eruptions until about 15 Ma (for example, tuff of Rainbow Canyon; Williams, 1967; Ekren and others, 1977; Bowman, 1985; Rowley and Siders, 1988; Mehnert and others, 1990). The widespread Leach Canyon

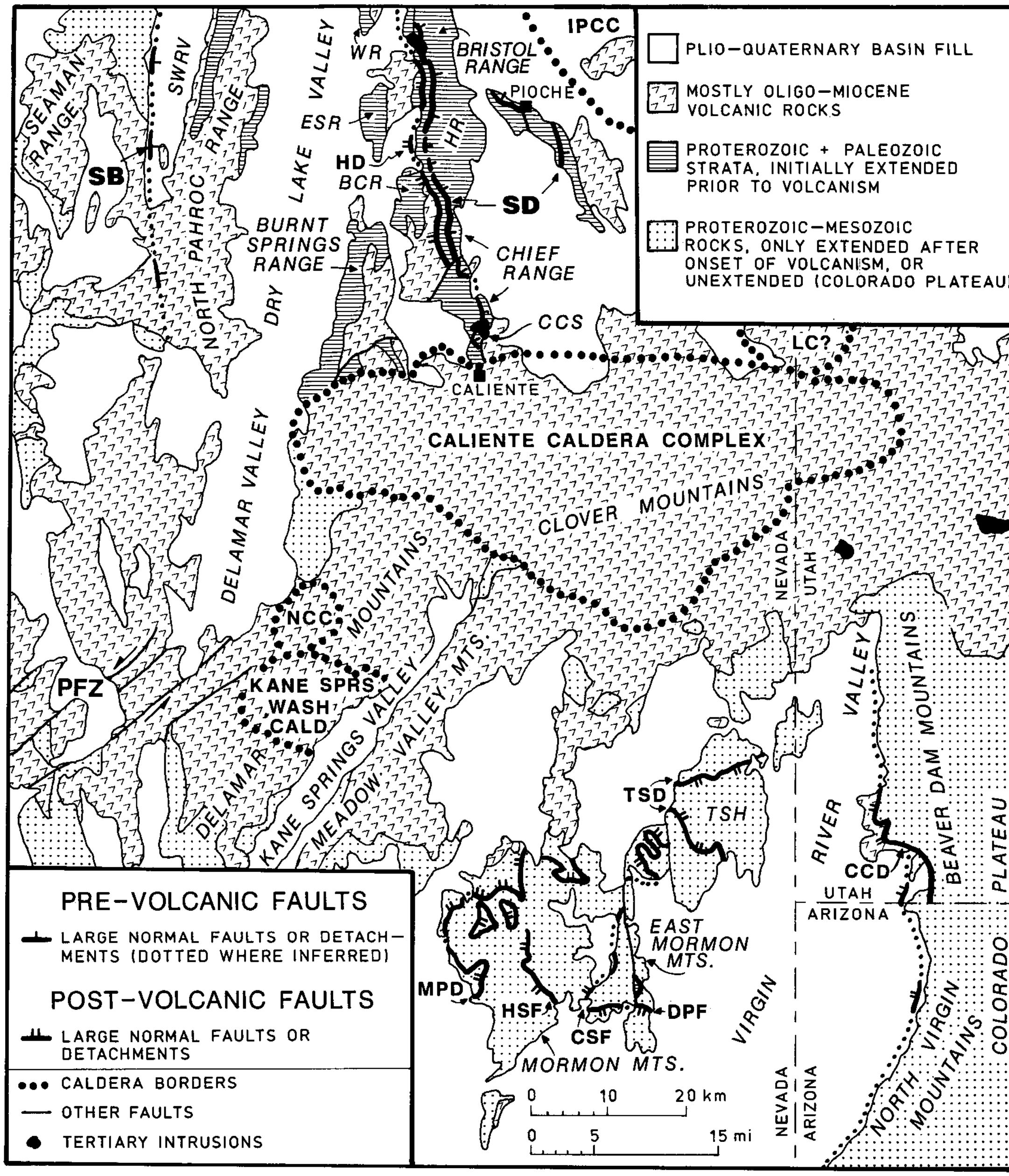

Figure 3. Map of the Dry Lake Valley–Mormon Mountains region, with the distribution of rocks extended in late Paleogene time in the upper plate of the Stampede detachment (SD) and the associated Seaman breakaway (SB). Other faults are mid-Miocene or younger: CCD, Castle Cliff detachment; CSF, Carp Summit fault; DPF, Davidson Peak fault; HD, Highland detachment; HSF, Horse Spring fault; MPD, Mormon Peak detachment; PFZ, Pahranagat fault zone; TSD, Tule Springs detachment. Other abbreviations are: BCR, Black Canyon Range; CCS, Cobalt Canyon stock; ESR, Ely Springs Range; HR, Highland Range; IPCC, Indian Peak caldera complex; LC, questionable source of the Leach Canyon Tuff; NCC, Narrow Canyon caldera; SWRV, southern White River Valley; TSH, Tule Springs Hills; WR, West Range. From sources cited in text.

Formation of the Quichapa Group (~24 Ma) has a similar distribution to later tuffs of that group (Williams, 1967; Fig. 7) and probably also erupted from this complex. (Best and others [1989a] show the Leach Canyon source farther north [Fig. 6], but more recent work [M. G. Best, 1990, written commun.] has revealed no evidence for it.)

Eastern Boundary. Pre- and synvolcanic extensional structures can be traced to within 50 km of the eastern edge of the Basin and Range in southern Utah (Fig. 4). Normal faults in the northern Beaver Lake Range are intruded by Oligocene(?) granodiorite (Lemmon and Morris, 1984; Walker and Bartley, 1991). Correlative granodiorite in the northern Mineral Mountains, dated as ~25 Ma by Aleinikoff and others (1986), cuts normal faults in Cambrian wall rocks (for example, Coleman and Walker, 1990). These faults in turn cut a Cretaceous thrust fault (Walker and Bartley, 1990, unpub. mapping), so that a Mesozoic age for the normal faults is unlikely. The eastern boundary is concealed under the Escalante Desert farther south.

The Indian Peak caldera complex, source of the 33–27 Ma Needles Range Group (Fig. 6; Best and Grant, 1987; Best and others, 1989b), obscures prevolcanic geology east of Pioche, Nevada, but several ranges farther east contain prevolcanic extensional structures (Fig. 5). Paleozoic strata in The Needles area (Fig. 4) are cut by small but numerous normal faults that locally are overlapped by the 34–32 Ma Escalante Desert Formation (Bartley, 1989, unpub. mapping). High- and low-angle normal faults in the Wah Wah Mountains cut Cretaceous thrust faults and predate basal Oligocene volcanic rocks (Ab-

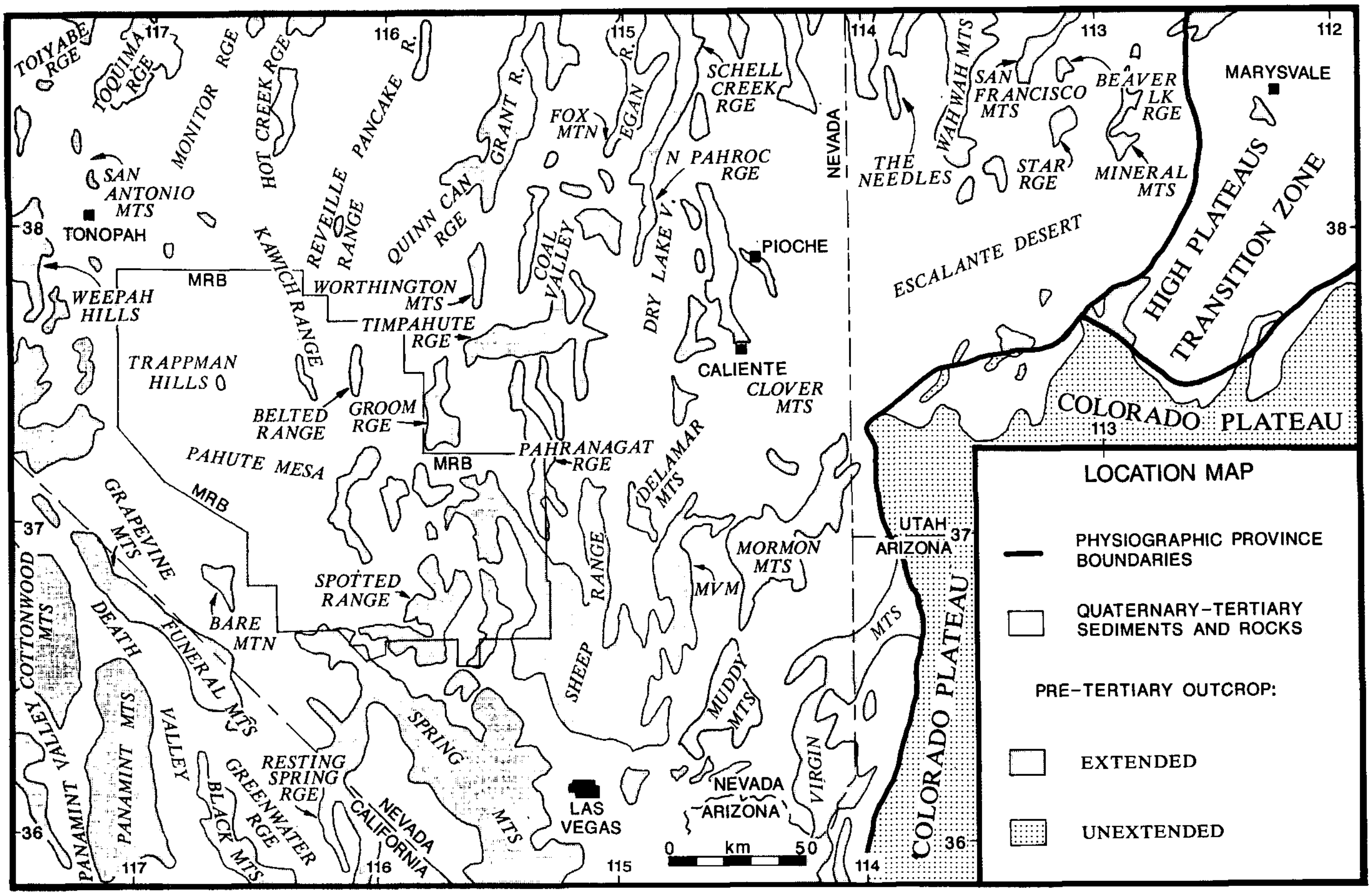

Figure 4. Location map of the principal area considered, with ranges discussed on a base of pre-Tertiary outcrop. The military reservation boundary (MRB) includes the Nevada Test Site, Nellis Air Force Base Bombing and Gunnery Range, and the Indian Springs Air Force Base. MVM, Meadow Valley Mountains.

bott and others, 1983; Best and others, 1987; Friedrich and Bartley, 1992a, 1992b). We infer late Eocene to Oligocene ages for these faults, but we cannot exclude older ages.

Stratigraphic Evidence

Sparse Tertiary sedimentary rocks in the southern part of the eastern belt provide important evidence that extension there was not Mesozoic in age (compare, for example, Allmendinger and Jordan, 1984) and help to locate the boundaries of the belt.

Western Boundary. The early Oligocene formation of Rattlesnake Spring in the North Pahroc Range provides good evidence for mid-Tertiary rather than Mesozoic extension in the Dry Lake Valley area (Taylor, 1989; Taylor and Bartley, 1992). This formation comprises conglomerate, lacustrine limestone, and mafic flow rocks. The mafic rocks are stratigraphically lower than beds that contain Oligocene fossils, and the formation is overlain by approximately 31 Ma tuff. New $^{40}Ar/^{39}Ar$ data from the mafic rocks support this age: concordant whole-rock total gas and plateau (>50% of released Ar) ages of 31.98 and 32.38 ± 1.21 Ma, respectively (W. J. Taylor and D. R. Lux, 1992, unpub. data). The geometry, paleogeography, and provenance of the formation of Rattlesnake Spring all clearly indicate deposition in a closed(?), north-south elongate basin adjacent to, and synchronous with or shortly after, activity on the Seaman breakaway fault (Taylor and Bartley, 1992).

Southeastern Boundary. The Claron Formation (Fig. 5) straddles the southeast margin of the Great Basin and apparently records the transition from erosion of Sevier thrust belt highlands to extensional tectonism (Taylor, 1991, in press). The distribution and internal stratigraphy of Claron exposures in the Great Basin pertain to the location of the southeastern boundary of the eastern belt.

The Claron Formation is divided into three parts. A lower Paleocene–lower Eocene(?) part contains detritus derived from Mesozoic and Paleozoic formations (for example, Hintze, 1988; Goldstrand, 1990; Taylor, 1991). The Eocene middle part is dominated by fine clastic sediments and laterally continuous fresh-water limestone beds (Gregory, 1950; Bowers, 1972; Taylor, in press). The lower and middle parts are widespread on the Colorado Plateau and in the southeastern Great Basin and predate the onset of extension.

Age and stratigraphy of the upper Claron Formation suggest that the southeastern boundary of the eastern extensional belt is located north of Newcastle, Utah (Fig. 5). The upper Claron Formation is commonly overlain by ~27–24 Ma tuffs (the Isom and Leach Canyon Formations), but locally contains volcanogenic sandstone derived from the Oligocene Needles Range Group, indicating an Oligocene age. Coarse conglomerate dominates the upper part of the Claron For-

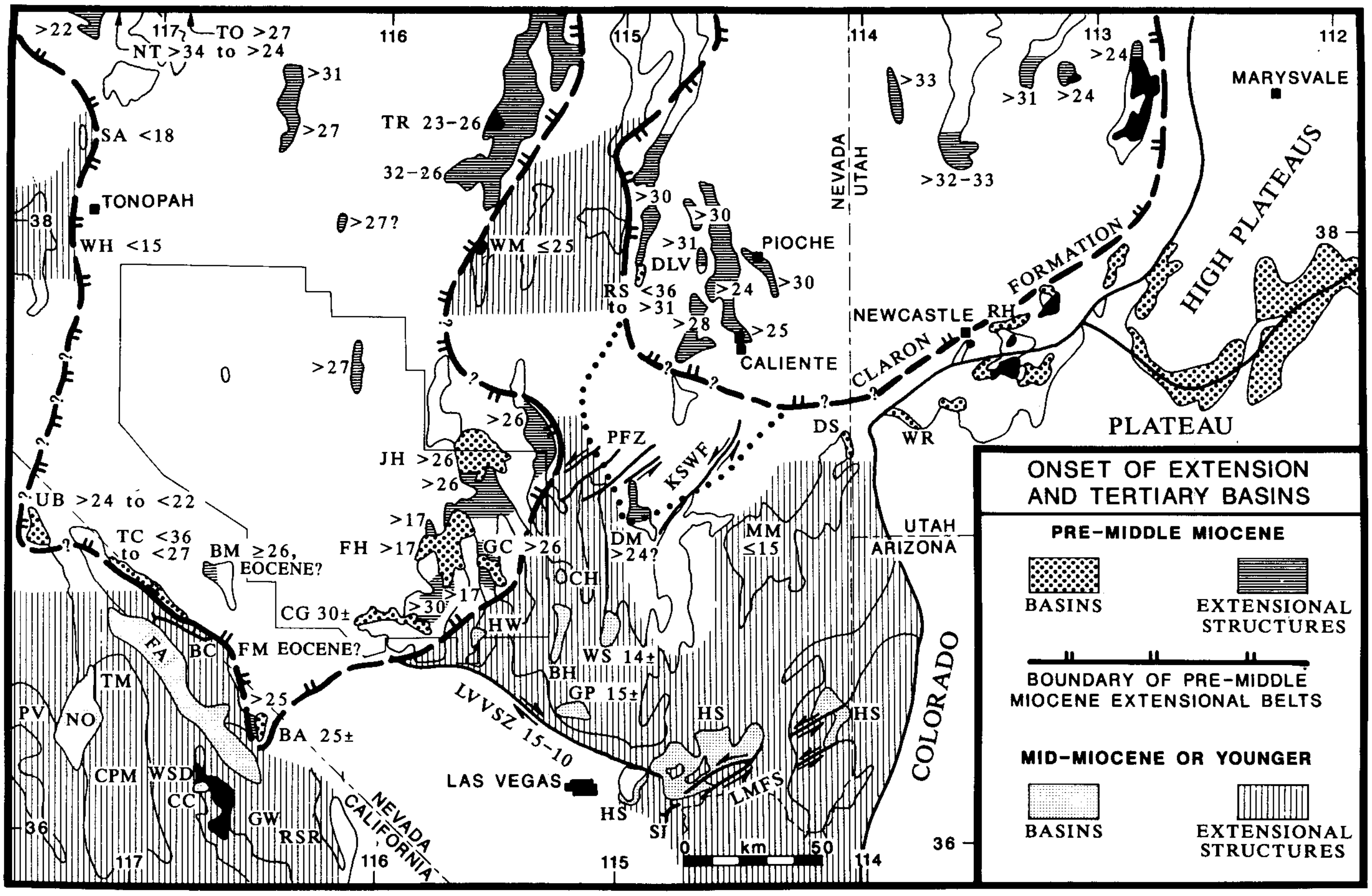

Figure 5. Map of same area as Figure 4, with locations of basins of probable extensional origin and areas where the age of onset of Tertiary extension is known or can be confidently inferred. Numbers refer to the age (Ma) of extensional structures or basins. Heavy dotted line encloses possible extension of eastern belt. See text for discussion of areas inside and near boundaries of extensional belts, and Table 1 for summary of areas where extension began in mid-Miocene or later time. Other localities: BA, Bat Mountain; BM, Bare Mountain; CG, Crossgrain basin; DLV, Dry Lake Valley (see Fig. 3); DM, Delamar Mountains; DS, Dodge Spring; FH, Fallout Hills; FM, Funeral Mountains; GC, Gravel Canyon basin; JH, Jumbled Hills; NT, northern Toiyabe Range (off figure); RH, Red Hills; RS, Rattlesnake Spring basin; TC, Titus Canyon basin; TO, Toquima Range; TR, Troy granite; UB, Ubehebe basin; WR, White Rock Spring.

mation in the Red Hills and also is present in sections southeast of Newcastle and at White Rocks (Fig. 5), being interbedded with volcanogenic sandstone at the latter locality (Taylor, in press). The influx of coarse detritus into the upper part of sections near Newcastle and White Rocks suggests proximity to a structural boundary, which we infer to be the southeastern boundary of the eastern extensional belt.

Ignimbrite Accumulations. Generalized isopach maps of ignimbrites of the Needles Range and Quichapa groups (Fig. 7) suggest that differentiation between the Great Basin and High Plateaus started before volcanism began, in accord with the structural data from the Dry Lake Valley area. Rowley and others (1978) noted that the 29.5 Ma Wah Wah Springs Formation is present on the High Plateaus of southwest Utah, but younger Needles Range Group ignimbrites are absent or very thin there. They inferred an Oligocene highland east of the "ancestral Wasatch front." Most Oligocene strata in the eastern belt (but outside of calderas) are ignimbrite sheets, and of these, most of the thicker deposits (>100 m) older than 24 Ma are restricted to within the eastern belt. We interpret this pattern to indicate a broad paleobasin that trapped the thicker parts of outflow sheets. The distribution and thickness of ignimbrite, however, may also be significantly affected by eruptive volume, wind, directed blasts, paleotopography, or subsequent erosion. These uncertainties and the thickness data presently available do not allow precise location of the inferred basin margins. For example, abrupt thickness changes across the Seaman breakaway (Taylor and Bartley, 1992) are poorly expressed by the isopachs. Similar abrupt thickness changes elsewhere are probably masked by the generalized contours.

The three oldest ignimbrites of the Needles Range Group are only sparsely preserved outside of the eastern extensional belt (Figs. 7a and 7b). The Wah Wah Springs Formation, the most voluminous, is only locally present northeast of the Beaver Dam Mountains. There it ranges from 0–15 m in thickness and overlies, or interfingers with, the Claron Formation (Hintze, 1986; Taylor, 1991). All four major ignimbrites of the Needles Range Group locally crossed one or both of the eastern and western boundaries, but only the youngest ash flow, the Isom Formation, reaches a significant thickness south of the

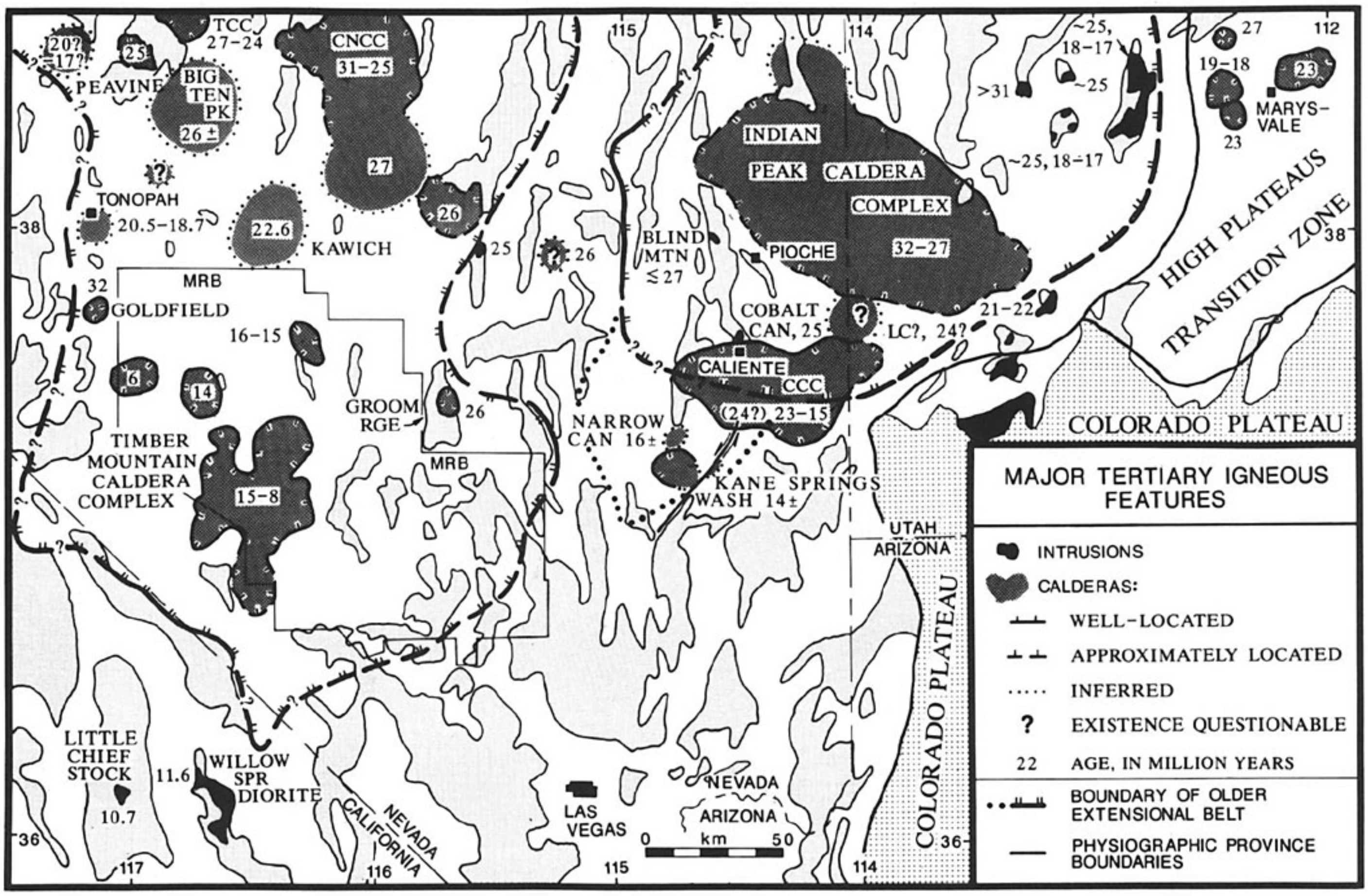

Figure 6. Locations and ages of calderas and Tertiary intrusions on a pre-Tertiary base from Figure 4. Abbreviations: CCC, Caliente caldera complex; CNCC, central Nevada caldera complex; LC, questionable source for the Leach Canyon tuff; TCC, Toquima caldera complex. From Best and others (1989a) and other sources cited in text.

extensional belt. By Isom time, the southern boundary may have been subdued enough to allow thick accumulations there.

Outflow facies of the Leach Canyon and Condor Canyon Formations (Fig. 7c) are distributed asymmetrically with respect to their source, being thicker to the north. Asymmetry of their zero isopachs has been interpreted to reflect erosion (Best and others, 1989a), further suggesting a highland to the south. Strong asymmetry is also apparent in the 100-m isopachs, suggesting that the Needles Range Group had not completely filled the inferred depression. By about 21 Ma, the southern boundary of the eastern belt apparently exerted little control on the distribution of outflow sheets (Harmony Hills tuff, Fig. 7c).

Outflow sheets of the Quichapa Group are thickest west-southwest of the Caliente caldera complex (Fig. 7c). Closely spaced isopachs there led Williams (1967) to infer mountainous paleotopography to the west. The isopachs trend parallel to the strike of the Seaman fault, supporting the possibility that the Delamar Mountains lay in the eastern belt. Tens of kilometers of mid-Miocene or younger left-lateral offset on the northeast-striking Pahranagat fault zone and Kane Springs Wash fault (Figs. 3, 5, and 6; Tschanz and Pampeyan, 1970; Liggett and Ehrenspeck, 1974; Harding and others, 1991) have distorted the initial distribution of isopachs, but probably have not strongly affected their spacing.

Kinematic-Geometric Model of the Eastern Extensional Belt

We interpret the Stampede detachment as an east-rooted master normal fault underlying upper-plate normal fault arrays, with the Seaman breakaway fault as its original up-dip surface trace (Fig. 8). Southward, the detachment may gradually lose displacement, or abut a strike-slip boundary that transferred extension elsewhere. Such a fault has not been mapped east of the Caliente caldera complex. In order to accommodate westward motion of the footwall of the Stampede detachment relative to the unextended terrain to the south, such a fault would necessarily have been left-lateral and rooted beneath the Colorado Plateau, forming a lateral boundary to the lower plate of the detachment (Fig. 8). A left-lateral fault (zone?) may have existed west of the Caliente caldera complex and may be represented by left-lateral prevolcanic faults in the Delamar Mountains (see above). A much more complicated zone than shown in Figure 8 is likely.

WESTERN EXTENSIONAL BELT

A similar Eocene to early Miocene extensional belt is inferred to have developed northwest of Las Vegas (Fig. 5), but its geometry and kinematics are less well known than those of the eastern belt. The south end

TABLE 1. EXTENSIONAL LOCALITIES OF MID-MIOCENE OR YOUNGER AGE

Abbreviation	Locality	Comments and references
BC	Boundary Canyon detachment	Active at ~10–5 Ma (Reynolds and others, 1986; Holm and Dokka, 1991); possible Eocene or Cretaceous activity (Hoisch and Simpson, 1989; Hodges and Walker, 1990)
BH	Black Hills basin	Correlated with nearby mid-Miocene basins (Guth and others, 1988)
CC	Copper Canyon basin	Deposited from ~7.5 to <5 Ma (Holm and Lux, 1991), synchronous with uplift of the Black Mountains, bracketed between 11 and 5 Ma (Asmerom and others, 1990)
CH	Chowderhead basin	Correlated with nearby mid-Miocene basins (Guth and others, 1988)
CPM	Central Panamint Mountains	Extension in mid-Miocene(?) to Pliocene time (McKenna, 1986; McKenna and Hodges, 1990; Hodges and others, 1990)
FA	Furnace Creek/Artist Drive basin	Filled from ~14–4 Ma, concurrent with slip on the right-lateral Furnace Creek fault zone and large-scale crustal extension to the south (Cemen and others, 1985)
GP	Gass Peak basin	Deposits include ~15 Ma tuff (Guth and others, 1988)
GW	Greenwater Range	Tilting and normal faulting at 9–4 Ma (Wright and others, 1984)
HS	Horse Spring basin	Records widespread crustal sag at ~19–18 Ma, and major extension in mid-Miocene (Bohannon, 1984)
HW	Heaven's Well basin	Correlated with nearby mid-Miocene basins (Guth and others, 1988)
KSWF	Kane Springs Wash fault	Offset of 4–10 km after 14 Ma (Tschanz and Pampeyan, 1970; Liggett and Ehrenspeck, 1974; Harding and others, 1991)
LMFS	Lake Mead Fault system	~65-km mid-Miocene left-lateral slip linked to extensional faulting and deposition of the Horse Spring Formation (Anderson, 1973; Bohannon, 1984)
LVVSZ	Las Vegas Valley shear zone	~40-km mid-Miocene right-lateral slip associated with extension to the north and westward translation of the Spring Mountains (Fleck, 1970; Guth, 1981; Bohannon, 1984; Wernicke and others, 1982)
MM	Mormon Mountains region	~55-km extension, beginning at or after ~15 Ma and continuing to Pliocene(?) (Wernicke and others, 1984, 1985; Axen and others, 1990)
NO	Nova Basin	Synorogenic deposition from ~7 to <3 Ma (Hodges and others, 1989, 1990; Snow, 1990)
PFZ	Pahranagat fault zone	10- to 16-km mid-Miocene to Quaternary left-lateral slip, possibly transferring extension from Dry Lake Valley area to west of the Sheep Range (Tschanz and Pampeyan, 1970; Liggett and Ehrenspeck, 1974; Wernicke and others, 1984)
PV	Panamint Valley	Pliocene or younger opening (Burchfiel and others, 1987; MIT Field Geophysics Course and Biehler, 1987)
RSR	Resting Spring Range	Tilting between 12–9 Ma (Troxel and Heydari, 1982; Wright and others, 1984)
SA	San Antonio Mountains	40°–80° of tilting beginning between 25 and 16 Ma and continuing past 16 Ma (Shaver and McWilliams, 1987)
SI	Saddle Island detachment	~20-km mid-Miocene or younger normal slip coordinated with the Lake Mead Fault system (Weber and Smith, 1987; Duebendorfer and others, 1990; Duebendorfer and Wallin, 1991)
TM	Tucki Mountain	Detachment faulting probably beginning in mid-Miocene time and continuing into late Miocene or Pliocene (Wernicke and others, 1988b; Hodges and others, 1989)
WH	Weepah Hills region	Deposition of the widespread mid-Miocene and younger Esmeralda Formation synchronous with detachment faulting (Stewart and Diamond, 1990)
WM	Worthington Mountains	North-directed ductile shear of a 25 Ma granitic body, subsequent west-directed brittle normal faulting (Martin, 1987; Martin and Bartley, 1989)
WS	Wamp Spring basin	Synorogenic deposits include an ~14 Ma tuff (Guth and others, 1988)
WSD	Willow Spring Diorite	Emplaced at mid-crustal levels ~11 Ma uplifted <6 Ma (Asmerom and others, 1990; Holm and Lux, 1991)

of the western belt was in the northern Death Valley area (Figs. 4 and 5); extension in the Death Valley–Spring Mountains transect was mid-Miocene to Pliocene in age, and extension in the Sheep Range area began in mid-Miocene time, synchronous with motion on the Las Vegas Valley shear zone (Fig. 5 and Table 1; Wernicke and others, 1988a, 1989).

Stratigraphic Evidence

Strata in extensional(?) basins of definite or probable Oligocene age mark the southeastern and southwestern edges of the western belt (Fig. 5). The Crossgrain basin, the Titus Canyon Formation, the Ubehebe basin, and stratigraphically lowest Tertiary rocks at Bat Mountain are most reliably dated. Other pre-middle Miocene basins of the western belt (Fig. 5) are only bracketed as younger than their Paleozoic substrate, but several observations suggest that all of these basins formed due to late Paleogene extension.

Southeastern Boundary. Paleogene and Miocene strata of the Crossgrain basin (Fig. 5; Barnes and others, 1982; Guth and others, 1988) help define the southeastern boundary of the western belt. The lower sequence of the Crossgrain basin was assigned to the Miocene Horse Spring Formation (defined near Lake Mead; see Bohannon, 1984), but contains a thin crystal-rich ash-flow tuff giving a K/Ar-biotite age of 30 ± 0.9 Ma (Barnes and others, 1982; Hinrichs, 1968; recalculated after Steiger and Jäger, 1977; Dalrymple, 1979). The lower sequence consists mainly of argillaceous limestone, with intercalated conglomerate layers as much as 30 m thick, suggesting subsidence of a closed basin with intermittent incursions of high-energy deposits. The conglomerate contains clasts to 30-cm diameter, chiefly derived from Precambrian quartzite with minor Paleozoic carbonate clasts (Barnes and others, 1982). As much as 500 m of sediment underlies the dated tuff (Barnes and others, 1982). The older sequence is overlain in angular unconformity by the rocks of Pavits Spring, which contain clasts of the 18.5 Ma Hiko Tuff near their base.

Strata in nearby basins are similar to those of the Oligocene Crossgrain basin, containing well-rounded cobbles of locally derived Precambrian quartzite, Paleozoic carbonate, and second-cycle chert pebbles (Tschanz and Pampeyan, 1970; Guth and others, 1988). We view these deposits as broadly correlative with those of the Crossgrain basin and deposits in the northwestern Pahranagat Range (Jayko, 1990; not in Fig. 5). Conglomeratic strata in the Jumbled Hills and Gravel Canyon basin are overlain by 33–26 Ma volcanic rocks, and beds in the Fallout Hills are overlain by 26–17 Ma volcanic rocks (Fig. 5; Ekren and others, 1977); all overlie Paleozoic rocks.

Basal Tertiary(?) conglomerate of undescribed clast content underlies the Monotony Tuff (~27 Ma) in the hanging wall of the Badger Mountain fault (Jayko, 1990). The conglomerate is absent on the footwall, where Monotony Tuff lies directly on Paleozoic rocks (Jayko, 1990; Best and Christiansen, 1991). Jayko (1990) infers Oligocene activity on that fault, which we suggest was synchronous with conglomerate deposition.

Prevolcanic strata of the Gravel Canyon basin, Jumbled and Fallout hills, and northwestern Pahranagat Range are similar to the section of the Crossgrain basin. This, along with proximity to that basin, permissive age constraints, and overlap relations with normal faults (see below), suggests deposition in late Paleogene extensional basins.

Extension did not begin until mid-Miocene time in the area southeast of these basins, synchronous with deposition of mid-Miocene sediments there and with motion on the Las Vegas Valley shear zone (for example, Guth, 1981; Fig. 5, Table 1).

Southwestern Boundary. Oligocene and lower Miocene deposits north and east of Death Valley mark the southwestern margin of the western extensional belt (Fig. 5). The Oligocene Titus Canyon Formation in the northeastern Grapevine Mountains contains synorogenic conglomerate and megabreccia intercalated with lacustrine limestone and mudstone (Figs. 4 and 5; Stock and Bode,

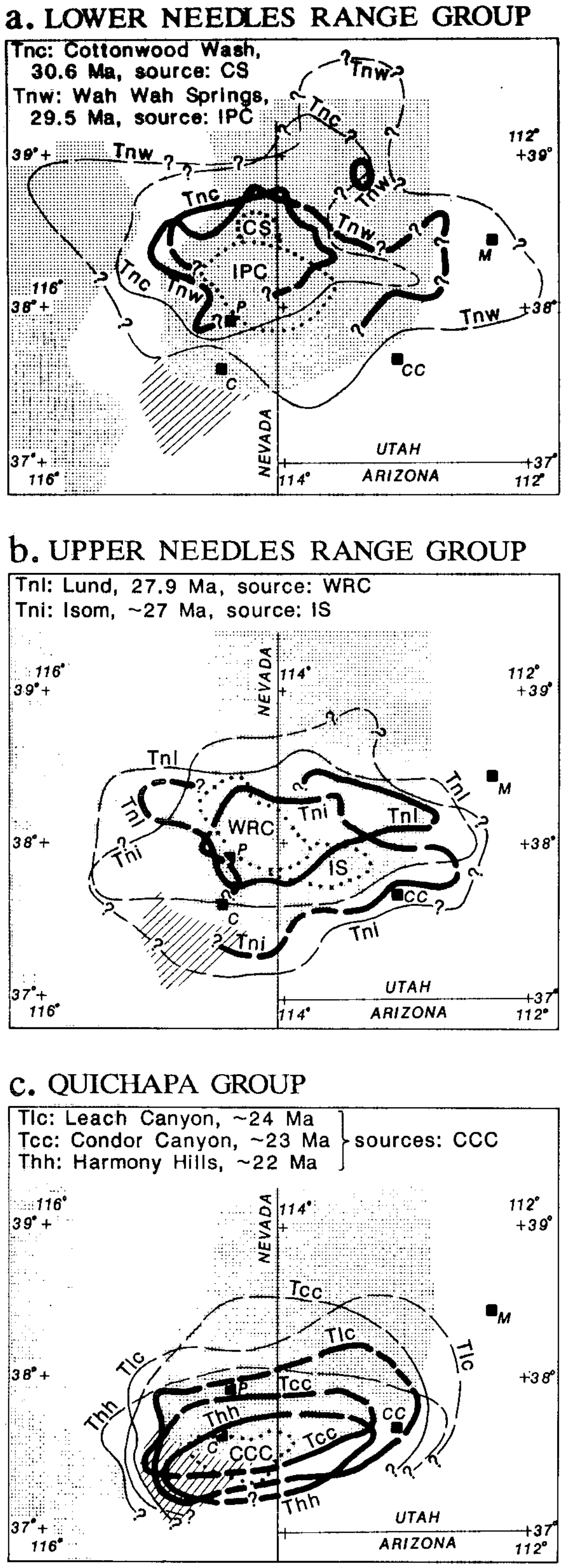

Figure 7. Generalized isopach maps of major ignimbrite deposits from the Indian Peak and Caliente caldera complexes, compared with the pre-Miocene extensional belts (shaded; ruled area shows possible extension of eastern belt). Thin lines are 0-m, and thick lines are 100-m isopachs. Dots outline known or inferred source calderas. Isopachs from Best and others (1989b) and Williams (1967). Abbreviations: C, Caliente; CC, Cedar City; M, Marysvale; P, Pioche.

1935; usage here follows Reynolds, 1969, 1974a; also Saylor and Hodges, 1991). These rocks accumulated near an active fault scarp and contain an intercalated 28 Ma tuff (Reynolds, 1969, 1974b; Saylor and Hodges, 1991; age recalculated after Dalrymple, 1979). Early Oligocene (about 36–30 Ma) vertebrate fossils were found in the lower Titus Canyon Formation (Reynolds, 1969, 1974a; Stock and Bode, 1935). The Titus Canyon Formation is unconformably overlain by strata that are about 22.6–20.5 m.y. old (Reynolds, 1969, 1974a, 1974b; recalculated after Dalrymple, 1979).

As much as 200 m of angular fanglomerate at Bat Mountain (Fig. 5) was derived from a local highland to the south (Cemen and others, 1982) that we interpret to be the margin of the western belt. The fanglomerate is conformably overlain by playa deposits that contain a 24.7 ± 0.3 Ma tuff near their base (Cemen and others, 1985) suggesting that the fanglomerates are also Oligocene. Overlying fine-grained red sandstone interpreted to represent fluviolacustrine deposition in a period of tectonic quiescence accumulated until at least 19.8 ± 0.2 Ma (Cemen and others, 1985; Cemen and Wright, 1988, 1990).

Synorogenic Oligocene–early Miocene strata are present in the Ubehebe Formation of Snow (1990), which consists of basal conglomerate, a middle clastic sequence, and an upper clastic and tuffaceous section with 23.9–22.6 Ma $^{40}Ar/^{39}Ar$ ages (Snow, 1990; the lower conglomerate unit and subunits Tu1–Tu3 of the Ubehebe sandstone of Snow and White, 1990). The lower two map units thicken northward across the Ubehebe Crater fault and are synkinematic with it (Snow and White, 1990; Snow, 1990).

All of the western-belt deposits discussed above probably were deposited either in a single regionally continuous extensional basin, or in adjacent basins related to a late Paleogene extensional event. Snow (1990) correlated the Ubehebe Formation with the sequence from Bat Mountain and with the Titus Canyon Formation on the basis of age, lithology, and proximity when Neogene structures are palinspastically restored (Fig. 9; see also Snow and Wernicke, 1989). The reconstruction of Wernicke and others (1988a; Fig. 9) places the Ubehebe, Titus Canyon, and Bat Mountain deposits south of the Crossgrain basin. If a single basin existed, then it had dimensions of about 25–30 km in an east-west direction, and 140 km north-south (Fig. 9), similar to larger modern basins in Nevada. Its paleogeography also was analogous to modern basins: deposition along fault scarps on its south and west sides, with lacustrine and fluvial deposition common.

Structural Evidence

Definite prevolcanic normal faults are sparse in southwestern Nevada, probably due to the small area of exposed prevolcanic rock,

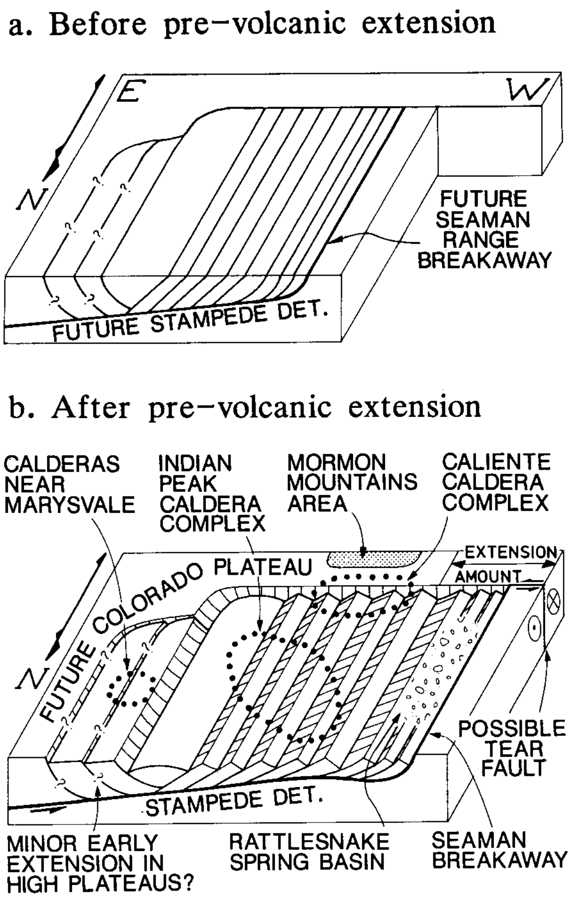

Figure 8. Schematic block diagram (view to the southeast), showing a possible geometry of the southern boundary of the Dry Lake Valley extended terrain and its relation to early Oligocene sediments (Rattlesnake Spring basin) east of the Seaman breakaway and the subsequent locations of Oligo-Miocene caldera complexes. Note that the postulated tear fault at the southern margin must be rooted in order to allow westward translation of the footwall of the Stampede detachment relative to the unextended terrain to the south. Such a fault (zone) may have provided a magmatic conduit to the upper crust when the migrating volcanic arc reached this latitude.

and no master detachment has been identified there. The pre-Tertiary stratigraphy and structural history are complex, causing ambiguity in interpretation of pre-Tertiary structures. For example, sedimentologically complex deposits of the eugeosyncline north of Tonopah and west of the central Nevada caldera complex (Figs. 4, 5, and 6) were deformed in the Antler and Sonoman orogenies (for example, Stewart, 1980) and may have experienced Permian to Jurassic contraction and extension also (for example, Oldow and Bartel, 1987; Smith and Miller, 1990; Bartley and Taylor, 1991).

Southeastern Boundary. Several normal faults adjacent to the southeastern boundary of the western belt are either overlapped by the oldest local volcanic or Tertiary(?) sedimentary rocks or offset prevolcanic strata more than volcanic strata. These faults have a similar distribution to sedimentary rocks of known or inferred Oligocene age (discussed above; Fig. 5) and support an extensional origin for those deposits.

Normal faults in the Spotted Range are overlain by strata of the Fallout Hills and Crossgrain basin (Figs. 4 and 5; Tschanz and Pampeyan, 1970), showing that extension there began before late Oligocene time. Oligocene(?) sedimentary strata of the Gravel Canyon basin also overlap east- and west-directed normal faults and tear faults in the Pintwater Range (Fig. 4 of Guth, 1990).

Two large faults in the Pahranagat Range area are probably early Tertiary normal faults (Figs. 4 and 5). The Badger Mountain and East Pahranagat faults offset pre-Tertiary deposits much more than they displace Oligo-Miocene ash flows, most likely with normal sense (Jayko, 1990). The Badger Mountain fault probably controlled distribution of Oligocene(?) prevolcanic sediments (Jayko, 1990; see above), consistent with Oligocene activity on that fault. Some normal faults in the Jumbled and Fallout Hills are either overlapped by prevolcanic strata or Oligocene–early Miocene tuffs, or they can be shown to offset prevolcanic rocks more than the Oligo-Miocene section (Jayko, 1990; Tschanz and Pampeyan, 1970; Ekren and others, 1977). We infer a late Paleogene age for these structures based on their spatial association with basins of Oligocene(?) age. Extension east of the Pahranagat Range began after about 10 Ma (Jayko, 1990).

Southwestern Boundary. Stratigraphic evidence for the location of the southwestern boundary (Fig. 5) is supported by mapped faults in the Funeral and Cottonwood Mountains, as well as by thermochronometry. Oligocene(?) strata at Bat Mountain overlap west-dipping normal faults (McAllister, 1971; Cemen and Wright, 1990). The oblique left-lateral normal Ubehebe Crater fault was active during deposition of the Ubehebe Formation (Snow and White, 1990; Snow, 1990; see above).

$^{40}Ar/^{39}Ar$ and K/Ar mineral cooling ages from the Funeral Mountains and western Bare Mountain (Fig. 5) range from Cretaceous to Miocene and have been variously interpreted (for example, Carr and Monsen, 1988; DeWitt and others, 1988; Hoisch and Simpson, 1989; Monsen and others, 1990; Dumitru and others, 1991). An important Miocene extensional event clearly is indicated (Reynolds and others, 1986; Holm and Dokka, 1991), and Cretaceous extension has also been invoked (for example, Hodges and Walker, 1990; Applegate and others, 1991). Paleogene cooling ages may reflect an extensional event of that age, as suggested by the nearby Oligocene synorogenic strata.

Central Part. Sparse prevolcanic normal faults help define the center of the western extensional belt. Low-angle normal faults that dismember a large Mesozoic(?) fold in the Belted Range are overlain by the ~27 Ma Monotony Tuff (Figs. 4 and 5; Ekren and others, 1967). Normal faults in the Hot Creek Range are overlapped by pre–31 Ma rhyodacite flows and pre–27 Ma sedimentary rocks (Gilmore Gulch Formation; Figs. 4 and 5; Quinlivan and Rogers, 1974). Mesozoic ages cannot be ruled out for these structures, however.

Generally sparse depositional contacts between Tertiary and older strata complicate interpretation of faults in other pre-Tertiary rocks of the central western belt. For example, mid-Tertiary strata in fault contact with pre-Tertiary rocks in the Reveille and Park Ranges (Fig. 4; the latter is north of the Hot Creek Range) are mapped as being much less normal faulted than the older rocks (Ekren and others, 1973; Dixon and others, 1972). Many faults in pre-Tertiary rocks may be late Paleogene normal faults, but their origin is obscured by differential rotations between Tertiary and pre-Tertiary blocks.

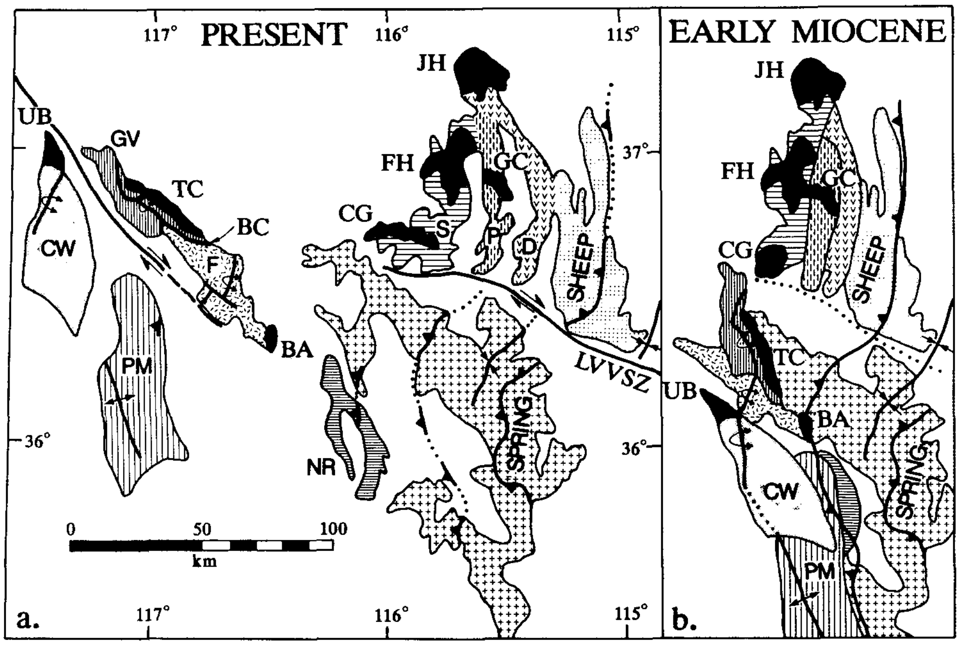

Figure 9. Pre–middle Miocene palinspastic reconstruction of basin deposits included in the western belt, with individual range blocks patterned; mainly based on Wernicke and others (1988a) and Snow and Wernicke (1989). The Grapevine Mountains (GV) are restored relative to the Funeral Mountains (F) across the Boundary Canyon detachment (BC) by realigning west-vergent fold couples (see also Snow, 1990). Restoration of ranges north of the Las Vegas Valley shear zone (LVVSZ) is only qualitatively correct and was achieved by subjective retrodeformation of left-lateral faults in the vicinity of the Cross-grain basin (CG), 12 km of closure between the Pintwater (P) and Sheep ranges, and 8 km of closure between the Spotted (S) and Pintwater Ranges, with the south ends of those ranges kept at a constant distance from the LVVSZ. Total east-west closure between the Cross-grain basin area and the Sheep Range is about equal to the ~40 km of slip on the LVVSZ. Other abbreviations are the same as in Figure 5, except the following: CW, Cottonwood Mountains; D, Desert Range; NR, Nopah and Resting Springs ranges; and PM, Panamint Mountains.

Northeastern Boundary. Latest Oligocene–earliest Miocene extensional structures are present in the Grant and Quinn Canyon Ranges (Figs. 4 and 5). Low-angle normal faults in the Grant Range are bracketed between 27 and 14 Ma (Fryxell, 1988; Taylor and others, 1989) but are inferred to be coeval with 27–23 Ma normal faults in the Quinn Canyon Range, where extensional tectonism and magmatism overlapped in time (Bartley and others, 1988; Taylor and others, 1989; Bartley and Gleason, 1990; Bartley and Taylor, 1992, unpub. $^{40}Ar/^{39}Ar$ data). K/Ar and $^{40}Ar/^{39}Ar$ mica ages of the Cretaceous Troy granite in the Grant Range are between 26 and 23 Ma (Armstrong, 1970; Taylor and others, 1989). They are interpreted as uplift ages (Fryxell, 1988) and may record an extensional episode.

Western Boundary. The western margin is best located southwest of the Toiyabe and Toquima Ranges and northeast and east of the San Antonio Mountains and Weepah Hills (Figs. 4 and 5), although it is poorly located overall due to volcanic and Quaternary cover and complex pre-Tertiary geology. Boden (1986) and Brem and others (1985) argued for structural control of the Toquima and Peavine caldera complexes in the southern Toquima and Toiyabe Ranges, respectively (Figs. 4 and 6). Brem and others (1985) noted that early faults of this set in the Toiyabe Range have normal separation. Ferguson and Cathcart (1954) mapped high-angle normal faults and both older-on-younger and younger-on-older low-angle faults in the southeastern Toiyabe Range. These faults predate dikes of the Jett Canyon system that appear to be related to caldera collapse during eruption of the 23 Ma Darrough Felsite (Speed and McKee, 1976).

A large west-dipping normal fault in the Toiyabe Range north of Figures 4 and 5 was correlated by Means (1962) and Kleinhampl and Ziony (1985) with a steep east-side-up reverse fault shown by Ferguson and Cathcart (1954). Both are prevolcanic (>22 Ma; Speed and McKee, 1976; Kleinhampl and Ziony, 1985) and cut the Cretaceous(?) Ophir pluton (Kleinhampl and Ziony, 1985), which has a K/Ar-biotite (uplift?) age of 53.9 ± 1.5 Ma (Speed and McKee, 1976). The east-dipping reverse fault probably originated as a west-dipping normal fault that was subsequently tilted along with presently west-dipping volcanic rocks. Large-magnitude, west-directed extension in the northern Toiyabe Range (Fig. 2) began before deposition of the pre–34 Ma tuff of Hall Creek and probably continued during deposition of tuffaceous sandstone and conglomerate older than the ~24 Ma Bates Mountain Tuff (Smith, 1989, 1992).

Kleinhampl and Ziony (1985) described prevolcanic normal faults in the Monitor Range (north of Figs. 4 and 5) that cut north-trending folds and are older than 34–27 Ma (their map unit "Twa"). Two low-angle younger-on-older prevolcanic faults are also present in the range, but they may be related to Permo-Triassic emplacement of the Golconda allochthon, because they juxtapose Paleozoic rocks of very different facies.

The San Antonio Mountains (Figs. 4 and 5) apparently belong outside of the western belt. Shaver and McWilliams (1987) found 40°–80° east dips in a conformable section of 29–25 Ma ignimbrites there. The steepest dips are similar to the tilt inferred paleomagnetically for the subjacent Cretaceous Hall stock, suggesting that little, if any, of the tilting there is pre-Miocene (Shaver and McWilliams, 1987). Although as much as 50° of tilting may be prevolcanic, Shaver and McWilliams (1987) concluded that all of the tilting is post-Oligocene. Onset of postvolcanic tilting at Hall began between 25 and 16 Ma and continued episodically to the Holocene epoch (Shaver and McWilliams, 1987).

Extension in the Weepah Hills and adjacent areas apparently did not begin until mid-Miocene time, so that area also lies outside the western belt (Figs. 4 and 5). Deposition of the widespread Esmeralda Formation in the Weepah Hills, Mineral Ridge, and adjacent areas began at about 13 Ma, probably synorogenic with underlying detachment faults (Stewart and Diamond, 1990).

DISCUSSION

Northern Continuation of the Extensional Belts

These late Paleogene extensional belts can be traced into northern Nevada (Fig. 2), but their boundaries generally are located better in the area of Figures 4 to 6 than farther north. The northeastern boundary of the eastern belt and the northwestern boundary of the western belt are particularly poorly defined (Fig. 2).

Eastern Belt. Prevolcanic extension in the Dry Lake Valley area is probably related to late Paleogene extension in the Egan, Schell Creek, and Snake Ranges (Fig. 2; Bartley and others, 1988; Taylor and others, 1989). East-dipping normal faults formed above east-directed detachments at nearly identical stratigraphic levels in both areas, and their breakaway zones can be reasonably connected (Axen and others, 1988; Taylor and Bartley, 1992). The major difference between the areas is the metamorphic grade of the detachment footwalls: high-grade in the Northern Snake Range, low-grade in the Southern Snake Range, and unmetamorphosed in the Dry Lake Valley area (Miller and others, 1983; Bartley and others, 1988).

Extension in the Egan and Schell Creek ranges began at, or before, about 36 Ma and was ongoing by 35 Ma during emplacement of the Kalamazoo Tuff and slightly younger dacite lavas and intrusions (Gans and others, 1989). These rocks lie above angular unconformities and locally truncate faults with kilometers of normal slip. Synorogenic clastic deposition in the Schell Creek Range and northernmost Snake Range started in Oligocene time. These rocks lie in angular unconformity on previously tilted and normal-faulted strata, contain fanned dips, and are commonly tilted as little as 10°–15°, suggesting that extension was waning. Synorogenic sedimentary strata (about 32 to 29 Ma) at Sacramento Pass in the Snake Range are deformed by the Northern Snake Range décollement. The footwall of that fault cooled below the argon blocking temperature for biotite by 24 Ma (Lee and Sutter, 1991), but the age of last slip is unknown.

Mylonitization and detachment faulting in the Kern Mountains–Deep Creek Range are locally dated at 39–35 Ma and continued into early Miocene time, with cooling below 100 °C by 17 Ma (Gans and others, 1990; Fig. 2). Low-angle attenuation faults in the Fish Springs and House Ranges (Fig. 2) predate dikes correlated with nearby 30.8 Ma dikes (Hintze, 1978; Lindsey and others, 1975). Although the time of onset of this extensional faulting is unknown, we include the Fish Springs and House Ranges in the eastern belt.

The western boundary of the eastern belt is projected north from the Seaman fault to the east side of the stable Butte Mountains, which lie west of the Snake Range extended terrain (Fig. 2; Miller and others, 1983; Bartley and Wernicke, 1984). It probably continues east of the stable Goshute Mountains (Fig. 2; for example, Gans and Miller, 1983, Fig. 20; Wernicke, 1990).

Western Belt. The western belt can be traced north from the area of Figure 5 to the northern Toiyabe Range (see above) and the Eureka District (Fig. 2). Bartley (unpub. mapping) has confirmed the existence of extensional structures younger than the Cretaceous Newark Canyon Formation and older than 34 Ma volcanic rocks near Eureka (see Nolan, 1962; Nolan and others, 1971, 1974). These faults define two mainly west-dipping sets: one corresponds largely to the Dugout Tunnel "thrust zone" (Nolan, 1962); the other is near Pinto Summit on U.S. Highway 50 (Nolan and others, 1974). Faults of the former set cut strata at high angles, consistently place younger rocks on older (Nolan and others [1974] show several older-on-younger relations due to misidentification of altered Pogonip Group), have consistent top-to-the-west separation, and are overlapped by late Eocene–early Oligocene volcanic rocks in Ratto Canyon (Nolan and others, 1974). Faults near Pinto Summit cut the Lower Cretaceous Newark Canyon Formation and are overlapped by ~35 Ma volcanic strata (Bartley, unpub. mapping; Nolan and others, 1974). Extension in the Eureka district occurred between Late Cretaceous and earliest Oligocene time and may represent early deformation in the western belt or an older event (see below; Fig. 10).

Late Eocene–early Miocene extension in the Ruby Mountains area was probably part of western-belt deformation (Fig. 2; Snoke and Howard, 1984; Hurlow and others, 1991). West-directed mylonitic rocks overprint plutonic rocks as young as 29 Ma (Wright and Snoke, 1986) and are cut by 17 Ma basalt dikes (Snoke and Miller, 1988). $^{40}Ar/^{39}Ar$ and fission-track thermochronometry indicates cooling of the footwall through ~500 °C between 45 and 30 Ma, through ~300 °C at 22–21 Ma (Dallmeyer and others, 1986), and through 70 °C by 22 Ma (Dokka and others, 1986). Thermobarometry of extensional mylonite indicates 13–16 km of Oligocene uplift (Hurlow and others, 1991; Hodges and others, 1992).

The Grouse Creek–Albion metamorphic core complex (Fig. 2; Snoke and Miller, 1988) probably belongs in the western belt. Late Eocene to early Oligocene top-to-the-west ductile shear in the western part of the complex (Saltzer and Hodges, 1988) is kinematically and temporally compatible with deformation in the Ruby Mountains area. Top-to-the-east fabrics in the eastern part of the complex (Raft River Range) are enigmatic: they may be thrust or extension related and Cretaceous or Tertiary in age (Snoke and Miller, 1988; Malavieille, 1987) but may belong in the eastern belt.

It has been suggested that Paleogene extension and magmatism in the Great Basin migrated southward together (for example, Coney, 1980; Gans and others, 1989; Wernicke, in press; many others), implying that the extensional belts outlined here become gradually older northward. Armstrong and Ward (1991) reviewed the ages and distribution of Tertiary magmatism and metamorphic core complexes in the United States and Canada, allowing a first-order evaluation of this implication (Fig. 10).

A gross trend toward a southward decrease in age of extensional tectonism is clear from Canada to the southern Great Basin (for example, Gans and others, 1989; Fig. 10). That pattern, however, may reflect discrete south-stepping episodes of extension rather than a gradual southward progression of extension. Extension in Canada, Washington, and Idaho was largely synchronous in early to middle Eocene time. In contrast, distinct extensional events of early to middle Eocene age and mainly Oligocene age are documented locally from the Lost River–Lemhi Ranges area to

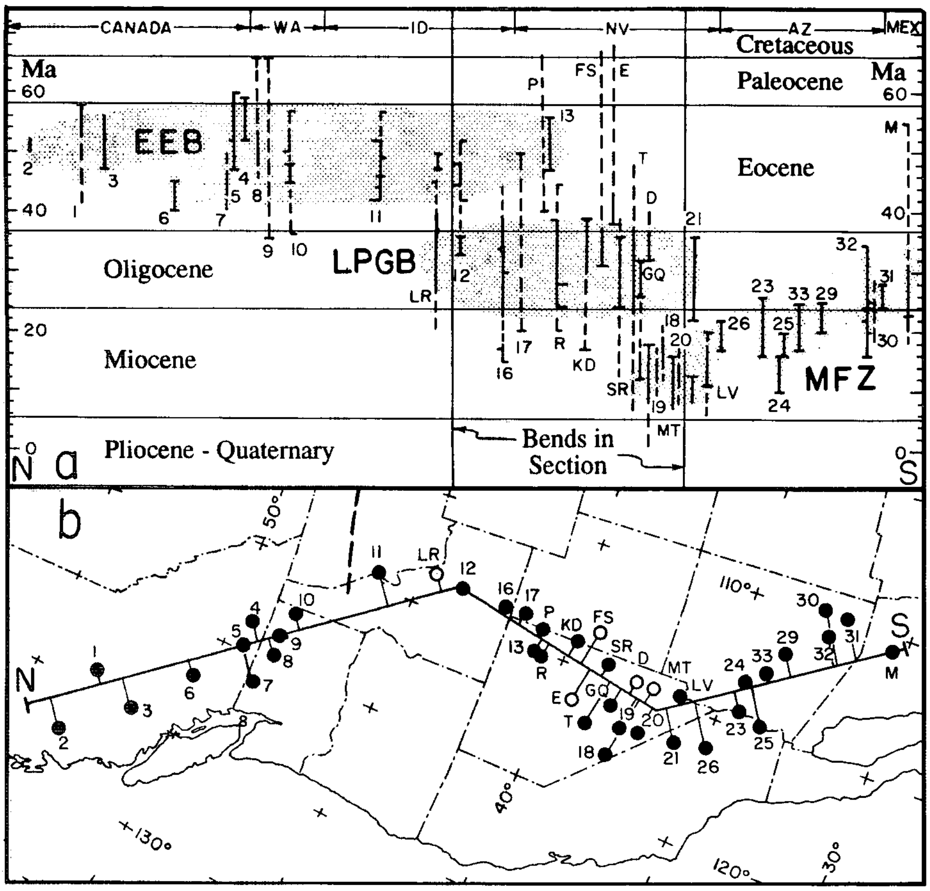

Figure 10. (a) Plot of age of extension versus location of metamorphic core complexes and some other highly extended terrains in western Canada, the United States, and Mexico. Solid lines indicate well-established age of extension, and dashed lines indicate permissible age; short horizontal marks show age brackets of different authors. Shaded areas indicate discrete extension episodes inferred to have occurred in different parts of the Cordillera: EEB, Eocene extensional belt of Canada, Washington, Idaho, and northern Nevada; MFZ, extended terrain in the southern Basin and Range inferred to have been related to the northward passage of the subducted Mendocino fracture zone (for example, Glazner and Bartley, 1984); LPGB, late Paleogene–early Miocene extensional belts of the Great Basin described here. Location line shown in b. (b) Map of the western United States and adjacent Canada and Mexico, showing the locations of metamorphic core complexes (solid circles) and other highly extended areas (open circles) projected onto the line of section in a. Heavy dashed line shows the north edge of the flat subducted lithosphere slab inferred from Laramide structural and magmatic patterns. Number and letter designations are given in Table 2. Data are from Armstrong and Ward (1991; numbers correspond to their Table 1), Anderson and others (1980), Miller and others (1987), Nourse (1990), and references in text. Note that Miocene-Holocene extension occurred in many areas south of the Lemhi Range but is not shown.

the Wood Hills–Ruby Mountains area (LR to 13 and R, Fig. 10; Silverberg, 1988, 1990; Thorman and Snee, 1988; Snoke and Miller, 1988; Janecke and Snee, 1990; Janecke, 1991; Hodges and others, 1992). Similarly, there is good evidence farther south in Nevada and Utah for the mainly Oligocene extensional event discussed here. The mainly Paleocene-Eocene Sheep Pass Formation, however, is locally preserved but widely distributed over much of the same area in central Nevada. The Sheep Pass is generally interpreted as synextensional and may record Eocene extension (Kellogg, 1964; Moores and others, 1968), although deposition may have begun as early as Late Cretaceous time, possibly synchronous with extension of that age (Fouch, 1979; Vandervoort and Schmitt, 1990). More data are needed to distinguish between these possibilities.

TABLE 2. CORE COMPLEXES AND OTHER HIGHLY EXTENDED AREAS*

No.	Locality
1	Nulki
2	Coast plutonic complex
3	Tatla Lake
4	Valhalla
5	Okanagan
6	Bridge River
7	North Cascades
8	Okanogan Dome
9	Kettle Dome
10	Newport Fault–Spokane Dome
11	Bitterroot Range
12	Pioneer Mountains
13	Wood Hills
16	Albion Range
17	Raft River Mountains–Grouse Creek Range
18	Mineral Ridge
19	Trappman Hills
20	Bullfrog Hills
21	Grapevine-Panamint Mountains
23	Whipple Mountains
24	Rawhide Mountains
25	Big Maria Mountains
26	Central Mojave Metamorphic Core Complex
29	South Mountain
30	Pinaleno Mountains
31	Rincon Mountains
32	Santa Catalina Mountains
33	Harcuvar Mountains
D	Dry Lake Valley
E	Eureka District
FS	Fish Spring–House Ranges
GQ	Grant–Quinn Canyon Ranges
KD	Kern Mountains–Deep Creek Range
LR	Lemhi Mountains–Lost River Range
LV	Las Vegas–Lake Mead Area
M	Sierra Magdalena
MT	Mormon Mountains–Tule Springs Hills
P	Pilot Range
R	Ruby Range
SR	Snake Range
T	Toiyabe Range

*See Figure 10.

Relations between the Eastern and Western Belts

The well-studied detachment systems in the two extensional belts were rooted in opposite directions, suggesting that they bounded a downward-widening crustal wedge in between. The belts are separated at the surface by a relatively narrow (~50 km) terrain that remained unextended until Miocene time, and the east side of this stable terrain is formed by east-directed breakaways for detachments in the Snake Range and Dry Lake Valley areas. If the Raft River Mountains belong to the eastern belt, then the east-directed detachment there also fits this pattern. Similarly, detachments in the Ruby Mountains, Toiyabe Range, and Grouse Creek–Albion Ranges are west directed and west rooted. Thus, the crust and mantle lithosphere column between the two belts and in their footwalls may have remained almost completely intact until mid-Miocene time or later.

Magnitude of Extension

The magnitude of late Paleogene extension in the two belts is poorly known overall but is locally large. We emphasize the importance of widespread evidence of late Paleogene extension but await more quantitative data concerning its magnitude and lateral variability. It is likely that the net magnitude of extension within each belt varied gradually along strike, although the extension accommodated by individual structures may have varied abruptly. Similarly, it is presently impossible to prove that extension was active simultaneously throughout each belt, although we favor that interpretation. The space-time resolution of extensional strain is generally too poor to address even the relative magnitudes of Paleogene versus Neogene extension, and it is premature to speculate on the meaning of apparent lateral variations in Paleogene strain.

In the eastern belt, large-magnitude late Paleogene extension (tens of kilometers) occurred in the Snake Range area, and ~28 km of extension occurred on the detachment fault in the Kern Mountains–Deep Creek Range (Fig. 2; Gans and Miller, 1983; Miller and others, 1983; Bartley and Wernicke, 1984; Gans and others, 1985, 1986, 1990). Prevolcanic extension of 60%–160% was accommodated by normal fault sets in the Ely Springs and northern Burnt Springs Ranges (Fig. 3; Axen, 1986; Lewis, 1987). Hintze (1978) described nearly bedding-parallel faults in the Fish Springs and House Ranges (Fig. 2) that attenuate the Paleozoic section to only 60% of its original thickness, implying large offsets. The magnitude of late Paleogene extension is unknown elsewhere in the eastern belt, but large-magnitude extension may have been common.

The magnitude of late Paleogene extension in the western belt is poorly determined also, especially in the southern parts. Large-magnitude extension has been documented in the Albion Mountains area (for example, Snoke and Miller, 1988), Ruby Mountains (Snoke and Howard, 1984; Snoke and Miller, 1988; Hacker and others, 1990), northern Toiyabe Range (Smith, 1989, 1992), and Grant Range (for example, Fryxell, 1988).

The magnitude of upper-crustal extension in the two belts may be laterally variable, but the significance of this variation is unclear. For example, the fact that metamorphic core complexes are exposed only in the northern parts of the belts may reflect a southward decrease of extensional strain to low values at the southern ends of the belts. Exposure of Paleogene tectonites in the core complexes, however, is partly related to Miocene to Holocene extensional fault systems that have further uplifted metamorphic cores. Less fortuitous overprinting may explain the lack of metamorphic tectonites in the southern parts of the belts or between core complexes. Also, large-magnitude extension need not expose middle crustal rocks. For example, about 125 km of extension between Tucki Mountain and the Spring Mountains (about 500% extension) left mid-crustal rocks exposed only in the Black Mountains and on Tucki Mountain (Wernicke and others, 1988a).

Relations between Late Paleogene Extension and Volcanism

The regional distributions of late Paleogene extensional tectonism and volcanism were quite different from one another (Fig. 2), requiring somewhat independent developmental controls at crustal levels. The distribution of extended upper crust, however, apparently affected eruptive style, because most calderas formed in previously extended crust (compare Figs. 5 and 6).

Spatial and Temporal Relations. The southward sweep of volcanism depicted by Stewart and Carlson (1976) is reflected in a general way by the southward migration of ignimbrite sources over a period of about 30 m.y. (Fig. 2). In contrast, late Paleogene upper-crustal extension occurred in north-trending belts that subsequently extended more in Neogene time; removal of the younger strain restores the older belts to narrower aspects (for example, Fig. 9). Extension was synchronous with, or post-dated, magmatism in the northern parts of both belts but predated magmatism in their southern parts.

Thus, the east-west belt of active crustal magmatism coincided only locally and temporarily with the active north-south–trending extensional belts, rather than regionally and typically. Best and Christiansen (1991) studied 30 extracaldera stratigraphic sections in the southern Great Basin and came to a similar conclusion. Each section contains several ignimbrites and records about 10 m.y. of history during the most vigorous volcanism in the Great Basin. They generally lack angular unconformities, fanned dips, significant clastic sediment, or other indicators of synvolcanic extension. In fact, a growing body of data suggests that synvolcanic extension in the Great Basin was directed north-south, nearly perpendicular to the extension direction inferred for the belts outlined here (Best, 1988; Bartley, 1989, 1990; Gans, 1990).

Relations between outflow sheets and extensional structures can be misleading indicators of space-time relations between extension and magmatism. For example, the Yerington District (Proffett, 1977; Proffett and Dilles, 1984) is cited as an example of space-time relations between extension and magmatism by Gans and others (1989; see also Wernicke and others, 1987). Voluminous outflow sheets of late Oligocene–early Miocene ignimbrites there are steeply tilted by domino-style faults, with younger andesite intruded along these and younger faults. Relation(s) between extension at Yerington and ignimbrite-forming magmatism are obscure because the ignimbrite sources lie well to the east (Ekren and others, 1980). Similarly, John and others (1989) described areas in southwestern Nevada (Fig. 2) where late Oligocene–early Miocene ignimbrite sequences are overlain with angular or erosional unconformity by mid-Miocene flows. They inferred that initial extension corresponded to a change from pre-extensional explosive eruptions to effusive eruptions. Most of the tuffs studied, however, are outflow facies, and systematic attempts to locate pre-ignimbrite extensional structures were not reported. Therefore, the relations between explosive magma systems and extension are uncertain there.

The ignimbrite sequence of the Candelaria Hills (Fig. 2) filled a fault-bounded trough that subsided in response to its eruption (Robinson and Stewart, 1984). Early tuffs filled paleovalleys of modest relief. Younger voluminous and widespread ash flows led Robinson and Stewart (1984) to conclude that no pre-eruption extension had occurred. The distribution of the post-extensional Needles and Quichapa groups is similar, however, discrediting such criteria for identifying prevolcanic extension. Although Robinson and Stewart (1984) noted that prevolcanic normal faults are not known in surrounding areas, such faults may be missed or misinterpreted in the complex pre-Tertiary rock assemblage there. We tentatively exclude the Candelaria Hills from the western belt.

Structural Control of Caldera Formation. Most calderas of the southern Great Basin are restricted to within, or near, the boundaries of the late Paleogene belts, where extension had already occurred but had ceased or slowed(?) (Fig. 6). This is also reflected by the coincidence of the stable terrain between the belts with the central Nevada magnetic quiet zone (Stewart and others, 1977; Mabey and others, 1978). Magnetic anomalies near the quiet zone are mainly due to Tertiary calc-alkaline igneous rocks, and a lack of intrusive rocks (and calderas) in the quiet zone is probably responsible for the lack of magnetic anomalies there, in spite of the widespread volcanic cover (Stewart and others, 1977; Fig. 6). In contrast, pre- or early synextensional volcanism was dominated by effusion of intermediate lava (for example, Gans and others, 1989). The location of most Great Basin calderas within previously extended crust is inconsistent with the suggestion that caldera formation is favored in unextended crust (Gans and others, 1989).

Few Great Basin calderas do not fit the generalizations made above. The slightly extended Marysvale field includes four small calderas (Fig. 6) but is dominated by flow rocks (for example, Cunningham and others, 1983; Steven and others, 1984). The oldest caldera (27 Ma) post-dates extension farther west and is inferred to lie above the root zone of the pre–32 Ma Stampede detachment (Figs. 5 and 8). Anderson (1971) described 26–24 Ma normal faults in the nearby plateaus, older than the other three calderas (23–18 Ma, Fig. 6). A small, 26 Ma caldera is inferred to lie between the two belts (Best and others, 1989a), but that location is uncertain (M. G. Best, 1990, written commun.). The Kane Springs Wash and Narrow Canyon calderas (Novak, 1984; Scott and others, 1991) are shown outside the eastern belt, but evidence discussed above suggests that the southwestern boundary of the eastern belt may include these calderas. The Miocene Timber Mountain caldera complex lies in the previously extended western belt, but was syntectonic with Miocene extension farther east, south (Fig. 5), and within the complex itself (for example, Scott, 1990; Carr, 1990). Extension was apparently active during eruption of some tuffs and less-voluminous lavas near the Candelaria Hills trough, but prevolcanic extension cannot be ruled out (Fig. 2; see above). The source of the Kalamazoo Tuff is inferred to lie in northern Spring Valley. Adjacent ranges contain Kalamazoo outflow sheets that overlie normal faults and Oligocene angular unconformities (Gans and others, 1989). Although broadly synextensional, the Kalamazoo eruption clearly postdated nearby extension also.

Structural control of eruptive style is also supported by relations near the southern margins of the belts where the southward migration of caldera formation stalled (Fig. 2). This is particularly clear in the eastern belt, where isochrons of the southern limit of calderas are closely spaced after 25 Ma, the youngest calderas there being about 15 Ma. Isochrons in the western belt are much more evenly spaced, and the southward progression continued to about 8 Ma but was synextensional. Mid-Miocene and younger effusive volcanism in the Death Valley region was broadly contemporaneous with early extension there (for example, timing relations summarized by Wernicke and others, 1988a, 1989; and in Fig. 5 and Table 1) but was also contemporaneous with formation of the nearby Timber Mountain complex in previously extended crust. Preconditioning of the crust in the Timber Mountain area by late Paleogene extension apparently favored caldera formation; such preconditioning did not occur farther south, impeding caldera formation in the Death Valley area where effusive volcanic rocks are common.

Similar controls may have operated at the Jemez volcanic field on the Colorado Plateau margin in New Mexico (for example, Self and others, 1986). The Jemez field lies on the west edge of the Rio Grande rift at its intersection with the older Jemez lineament and is larger than, and compositionally distinct from, other volcanic fields along the lineament, probably due to its location at the intersection (Smith and Luedke, 1984). Intermediate lava and subsidiary bimodal flows erupted from 13–7 Ma, synchronous with early extension. A tectonic and magmatic lull followed from 7–4 Ma, after which time renewed extension accompanied repeated ignimbrite eruption from the Valles caldera complex.

The space-time patterns of extension and magmatism in the southern Basin and Range are different from those in the Great Basin (for example, Glazner and Bartley, 1990). The pattern of migration of volcanism in the southern Basin and Range is the topic of some debate. Some authors favor a westward migration through time, especially in southern Arizona from 30–20 Ma, and relate that trend to a westward-steepening subducted slab (Coney and Reynolds, 1977; Spencer and Reynolds, 1989; discussed below). Others favor a more northerly migration, due to the east-west trend and northerly progression of a best-fit plane regressed through age-distribution maps of volcanic rocks and the correspondence of those trends to the projected location of the subducted Mendocino fracture zone (Glazner and Supplee, 1982). Consideration of voluminous Eocene-Oligocene volcanism in Mexico and late Miocene volcanism in the Las Vegas–Death Valley area leads us to favor a more northerly pattern of volcanic migration overall. Interestingly, the evolution of volcanic style in the southern Basin and Range and northern Mexico is opposite to that of the Great Basin: voluminous Oligocene ignimbrite eruption in the Sierra Madre Occidental of Mexico was followed by dominantly effusive volcanism in Arizona–Death Valley.

Nevertheless, comparison of space-time patterns of extensional tectonism (Fig. 10) and volcanism in the southern Basin and Range allows an interpretation consistent with ours for the Great Basin. Extension in the southern Basin and Range swept north (-west?) approximately simultaneously with volcanism, with the Mendocino triple junction, and with the subducted Mendocino fracture zone (Glazner and Supplee, 1982; Glazner and Bartley, 1984; Severinghaus and Atwater, 1990). A significant volume of Tertiary lava was erupted, but few calderas formed in the southern Basin and Range when compared to the ignimbrite provinces of Mexico or the southern Great Basin, consistent with our generalization that effusive eruption is favored in actively extending crust.

Huge volumes of mainly Oligocene ignimbrite dominate the Sierra Madre Occidental of Mexico and were preceded by intermediate lava flows and followed by eruption of alkaline basalt (for example, Aguirre-Diaz and McDowell, 1991; Wark, 1991). Extension in Mexico is typically assigned a syn- or post-ignimbrite age (for example, Wark and others, 1990; Henry and others, 1991), similar to previous suggestions for the Basin and Range of the United States. The controls on caldera formation proposed here, however, suggest that the possibility of precaldera extension in Mexico should be investigated.

The cause(s) of caldera formation being favored in extended crust must be related to the dynamics of magma generation and movement. Seismically imaged magma bodies typically are located near a mid-crustal velocity jump that probably reflects a density increase, suggesting that the rise of mafic magmas is impeded by low-density crust (Glazner and Ussler, 1988). Extended areas should have thinner upper crust than adjacent unextended regions. Mafic magmas therefore may rise to shallow depths in previously extended areas, forming large, near-surface magma chambers that precede caldera collapse. Active extensional structures may hinder formation of magma bodies large enough for caldera collapse or allow semicontinual release of volatiles and therefore favor effusive eruption. Rising mafic magmas in unextended crust may stop at depths where lithostatic pressure precludes the massive vesiculation needed for ignimbrite eruption.

Volcanotectonic Rifting Models

Active and Passive Rifting Models. The data summarized here (Fig. 2) are incompatible with passive and active rifting models (Sengor

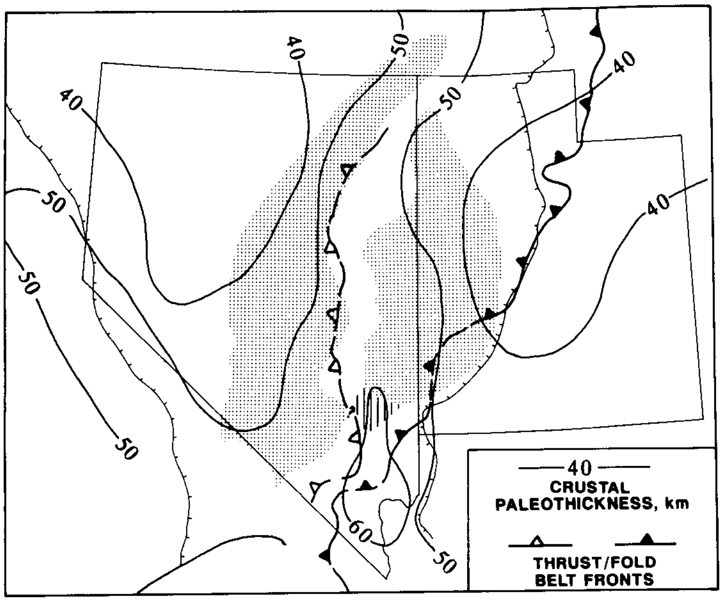

Figure 11. Comparison of the Oligocene extended belts (shaded) with the Mesozoic–early Cenozoic Sevier thrust front (closed teeth; Armstrong, 1968), the east limit of the Jurassic(?) central Nevada thrust belt (open teeth; for example, Bartley and Taylor, 1991; see also Nolan, 1962; Martin, 1987; Fryxell, 1988; Bartley and Gleason, 1990; many others), and paleo-isopachs of crustal thickness (Coney and Harms, 1984). Thin lines with ticks show east edge of the Sierra Nevada and west edge of the Colorado Plateaus and Rocky Mountain physiographic provinces.

and Burke, 1978) for the Great Basin as a whole. In passive rifting models, tectonically driven extension thins the lithosphere and generates magma by decompression melting. In active rifting models, lithospheric tumescence caused by advective heating and associated magmatism is followed by gravitational collapse and attendant extension. Both models predict specific relations of extension and magmatism in time *and* space and thus may be applicable on the scale of a few basin-range pairs (Gans and others, 1989; Gans, 1990). They are not applicable to the whole Great Basin because a variety of space-time relations exist in that area, including large areas in which magmatism and extension are temporally and spatially distinct (for example, Bartley and others, 1988; Taylor and others, 1989; Best and Christiansen, 1991) or in which major extension occurred with no magmatism (Guth, 1981, 1990; Bohannon, 1984; Wernicke and others, 1985; Guth and others, 1988; Axen and others, 1990).

Alkalic, Bimodal, and "Fundamentally Basaltic" Volcanism. The space-time patterns of late Paleogene extension and magmatism (Fig. 2) also suggest that the onset of extension is not necessarily closely related to the onset of alkalic, bimodal, or "fundamentally basaltic" volcanism, which occurred at about 20–16 Ma in the southern Great Basin (Christiansen and Lipman, 1972). Approximately the first 10 m.y. of volcanism after initial extension is represented by calc-alkaline volcanic rocks. The transition to alkaline or peralkaline ignimbrite eruption in the eastern belt occurred after eruption of the Hiko Tuff (18.5 Ma), but before eruption of the 16–14 Ma Ox Valley and Kane Springs units (for example, Novak, 1984), and may have been synchronous with initial extension in the Mormon Mountains–Tule Springs Hills area (Axen and others, 1990). Most of the ignimbrites that erupted from the Timber Mountain caldera complex (16–8 Ma) are also alkaline or peralkaline and were erupted during extension (for example, Scott, 1990; Carr, 1990; discussion above). Miocene extension in these areas occurred in a strike-slip plate margin regime, whereas late Paleogene–early Miocene extension and volcanism occurred in a subduction margin setting (Atwater, 1970; Glazner and Bartley, 1984; Severinghaus and Atwater, 1990). The conclusion of Christiansen and Lipman (1972) that the transition from calc-alkaline to "fundamentally basaltic" volcanism is related to inception of a strike-slip plate margin is permitted in the Great Basin.

Gravitational Spreading of a Thick Crustal Welt. Many authors infer that gravitational collapse of a thick crustal welt relic from thrusting caused extension in the Basin and Range province (for example, Coney and Harms, 1984; Glazner and Bartley, 1985; Wernicke and others, 1987). Our compilations are consistent with this concept because the late Paleogene extensional belts are within, and subparallel to, the Sevier and central Nevada thrust belts (Fig. 11; Armstrong, 1968; Bartley and Taylor, 1991).

Our compilation does not clearly support the four-stage history forwarded for Tertiary extension in western North America by Wernicke and others (1987): (i) formation of early intermontane basins; (ii) eruption of predominantly intermediate to silicic volcanic rocks; (iii) areally restricted large-magnitude extension (that is, the formation of metamorphic core complexes), occurring during or immediately after stage ii; and (iv) basaltic or bimodal volcanism accompanied regionally by varied amounts of extension. Late Paleogene extension in the Dry Lake Valley area was probably of large magnitude (see above), so that stages ii and iii were reversed in time. The southern part of the western belt may have had a structural history similar to the Dry Lake Valley area, but the magnitude of extension is completely unknown for that part of the western belt.

The Dry Lake Valley area lacks Mesozoic intrusive rocks, but it lay in the hinterland of the Sevier thrust belt and east of the central Nevada thrust belt prior to late Paleogene extension (Fig. 11; Bartley and others, 1988). At the onset of extension, this area more closely resembled the amagmatic area to the south than areas to the north where Mesozoic plutons are common. Wernicke and others (1987; after Sonder and others, 1987) argued that gravitational spreading of thickened crust caused extension in western North America (for example, Coney and Harms, 1984), with extension triggered by thermal weakening of the crust due to heat from older magmas. If this was the case, then the Dry Lake Valley area should have extended at the same time as the amagmatic area to the south.

Relations to the Subducted Slab. Tertiary changes of plate velocity patterns in the Pacific basin are commonly cited as influencing the onset of extension and volcanism in western North America. The location and orientation of the Tertiary extensional and magmatic belts discussed here, however, are not easily related to events at the western margin of North America in Eocene-Oligocene time. For example, the effects of decreased convergence rates and rapid Pacific plate reorganization at about 55, 42, and <30 Ma (Stock

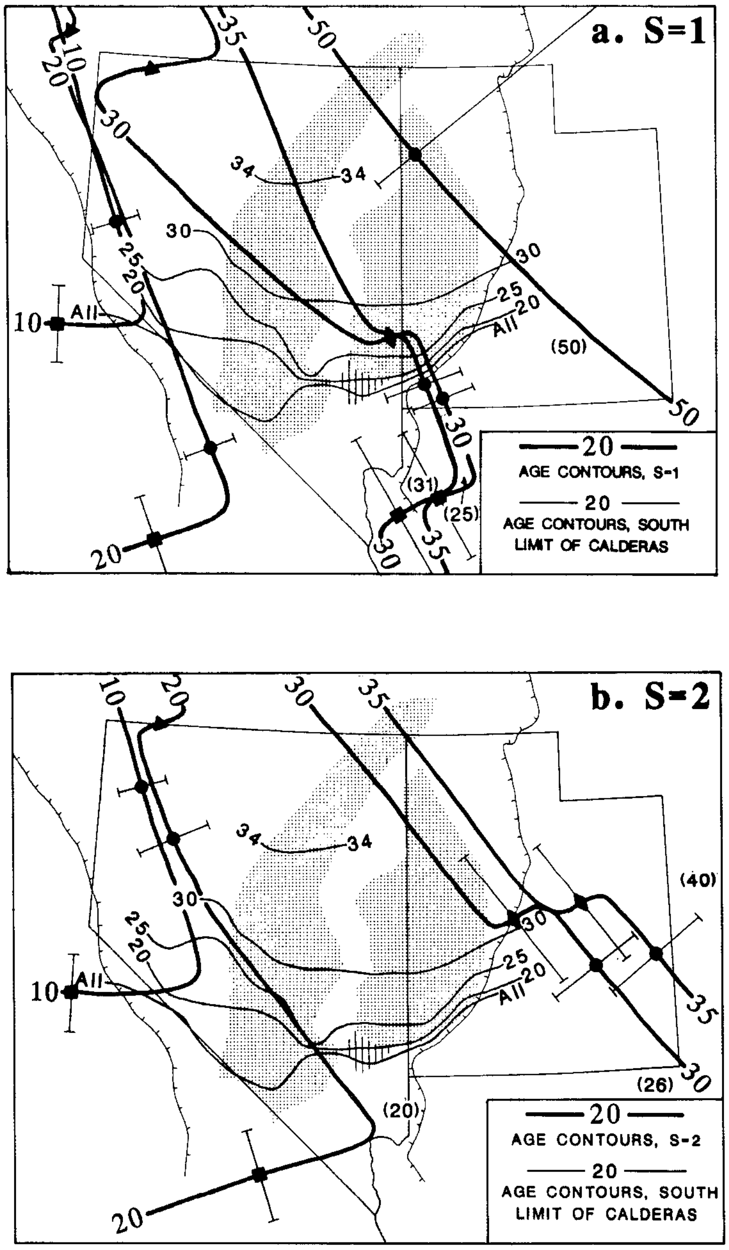

Figure 12. Comparison of the pre-Miocene extended belts (shaded, lined area is possible extension of eastern belt) with age contours (Ma) of the calculated position of the subducted slab parameter S of Severinghaus and Atwater (1990) and age contours of the southern limit of calderas (from Fig. 2). (a) The value S = 1 corresponds roughly to the position of the east end of paleo-Benioff zones, taken to indicate the maximum extent of mechanically strong subducted slab. (b) S = 2 contours represent aseismic slab that may still affect tectonics or magmatism of the overlying lithosphere (high slab temperatures at S = 3–4 allow rapid olivine flow). East- to northeast-trending S contours correspond to inferred positions of subducted fracture zones: Mendocino, squares; Surveyor, diamonds; and Sedna, triangles. Northwest-trending S contours follow the inferred positions of magnetic anomalies; they are not parallel because they are plotted on modern geography rather than on the retrodeformed base of Severinghaus and Atwater (1990), allowing several ages to be shown on the same map. Numbers in parentheses indicate the age (Ma) when subducted of oceanic lithosphere at the position plotted (Severinghaus and Atwater, 1990). Error bars on fracture zones are based on plate reconstruction models for the location of the Mendocino triple junction; error bars anchored with dots are based on subjective evaluation (Severinghaus and Atwater, 1990). See text for discussion.

and Molnar, 1988; Atwater, 1989) on margin-normal stress or on subduction angle are commonly stressed (Atwater, 1970; Best and Christiansen, 1991; Janecke, 1991; many others). These reorganizations can be correlated with tectonic events in some areas (for example, extension in the Pacific northwest and Canada at about 55–45 Ma, Fig. 10) but do not correspond to Cordillera-wide tectonic events.

Instead, changes in slab geometry and strength probably controlled the onset of extension in the Basin and Range (for example, Atwater, 1989). Severinghaus and Atwater (1990) modeled the condition of the subducted slab under southwestern North America in terms of a "slab parameter S" and presented contour maps of S values at various times (Fig. 12). S reflects the thermal state of the slab and increases with its temperature, which is a function of age of subducted lithosphere and elapsed time since subduction (Severinghaus and Atwater, 1990). For modern slabs, S = 1 corresponds to the limit of Benioff zones, beyond S = 3–4 rapid olivine flow is expected at plate rates, and S = 5–7 corresponds to slab temperature approaching asthenospheric temperatures and to assimilation of the slab into the asthenosphere (Severinghaus and Atwater, 1990). Thus, S = 1 marks the edge of slab material able to support significant elastic stresses. The absolute positions of S contours at a given time are not well known relative to the overlying lithosphere, but the spacing of S contours is relatively robust (Fig. 12).

We assume that the subducted lithospheric slab beneath the Basin and Range was gently inclined in Laramide time (Coney and Reynolds, 1977; Cross and Pilger, 1978, 1982; Helmstaedt and Schulze, 1991) and that subsequent modifications affected continental tectonics in the western United States. Slab paleodip is difficult to determine, so S values for a subhorizontal slab are plotted by Severinghaus and Atwater (1990); corrections for slab dip shift S contours west. S contours at 50 Ma are probably near their westernmost possible locations, due to the lack of heating expected on the top of a flat slab (Severinghaus and Atwater, 1990).

The subducted slab may have been buoyant and relatively flat throughout Tertiary time, so that large westward corrections of the S contours for slab dip may not be required. Oceanic lithosphere younger than 40–50 Ma is less dense than asthenosphere if the basalt-eclogite transition is prevented (Oxburgh and Parmentier, 1977; Sacks, 1983). In middle Tertiary time, the subducted slab under the Great Basin was <40–50 m.y. old when subducted (Fig. 12; Severinghaus and Atwater, 1990). Its previously flat aspect may have prevented the heating necessary for eclogite formation except at the edges.

S contours shift westward and become more closely spaced between 50 and 35 Ma as the model slab heats rapidly ($S = 1$, Fig. 12a). A west-southwest shift of the east edge of strong slab material ($S = 1$) under the Great Basin is inferred at about the time that extension began in the belts described here. The correlation with $S = 2$ is much worse (Fig. 12b). Strong, cold lithosphere between the Surveyor and Mendocino fracture zones apparently extended eastward to the present west edge of the Colorado Plateau until after 30 Ma (Fig. 12a). This flap may have prevented extension south of the belts outlined here until mid-Miocene time: by 20 Ma, the strong model slab extended only to the western edge of the present Great Basin (Fig. 12; Severinghaus and Atwater, 1990).

The northwest sweep of extension in the southern Basin and Range–Death Valley area correlates well with the northward passage of the Mendocino triple junction (Glazner and Bartley, 1984). This may be due to lithospheric flexure above the subducted north-migrating Mendocino fracture zone and stress drops near the plate margin (Glazner and Bartley, 1984), but thermal weakening or destruction of the slab south of the Mendocino fracture zone also provides an explanation (Glazner and Supplee, 1982; Severinghaus and Atwater, 1990). By 20 Ma, young subducted lithosphere south of the Mendocino fracture zone was probably weak or nonexistent and formed a "slab gap" (Fig. 12; Severinghaus and Atwater, 1990; compare with Dickinson and Snyder, 1979). Northward drift of the fracture zone with respect to the overlying North American plate probably triggered the northward sweep of extension and volcanism suggested by Glazner and Bartley (1984). Rapid westward retreat of the east edge of slab material south of the Mendocino fracture zone between 35 and 20 Ma preceded formation of the slab gap (Severinghaus and Atwater, 1990) and may explain the westward migration of volcanism in southern Arizona preferred by Spencer and Reynolds (1989) without steep subduction angles that are unlikely for such a short young slab.

Thermal weakening of the subducted slab may have triggered extension of North American lithosphere either by allowing upward transfer of heat or by removing elastic support of the thickened crust above. While the subducted slab was cold, strong, flat, and buoyant, it may have partly supported the crustal welt relic from Mesozoic shortening and insulated the overlying lithosphere from hot asthenosphere below (for example, Dumitru and others, 1991). Upon heating, the slab may have lost the elastic strength needed to support the overlying load and may have allowed extension to begin. This would be enhanced by thermal weakening of the overlying lithosphere as heat was transferred upward.

This concept is consistent with the increasingly documented Mesozoic extension in the western United States (for example, Allmendinger and Jordan, 1984; Applegate and others, 1991; Hodges and Walker, 1990; Hodges and others, 1992). Normal subduction allowed gravitational collapse continuously to limit ongoing crustal thickening. The change to a flat slab geometry shifted thickening to the east and impeded further extension in the Great Basin. Eocene extension proceeded unhindered in Canada and the northwestern United States north of the inferred flat slab (Fig. 10), continuously thinning the crust and lowering its gravitational instability to the point that further extension never occurred there. Eocene extension in Nevada was locally intense, but widespread extension occurred only after the underlying slab weakened in the late Paleogene.

The model slab edge beneath the Great Basin retreated in a west-southwestward direction in present coordinates (Fig. 12; Severinghaus and Atwater, 1990). This correlates only moderately well with the south-southwestward sweep of effusive volcanism and ignimbrite eruption in central and southwestern Nevada and correlates poorly with the southward sweep of ignimbrite eruptions in the eastern Great Basin (Fig. 12). This discrepancy may simply be related to the effects of slightly extended to unextended crust inhibiting the upward passage of magma to near-surface levels. By 20 Ma, the slab beneath the southern Great Basin had almost disappeared (Fig. 12), which is about the time of onset of alkalic volcanism, initial rapid extension in the Lake Mead–Mormon Mountains area, and renewed extension in the Dry Lake Valley area.

The southward migration of volcanism in the Great Basin has also been attributed to a southward steepening slab geometry that bounds the leading edge of a migrating asthenospheric wedge (for example, Cross and Pilger, 1978; Best and Christiansen, 1991). Warming of the flat slab may have triggered the basalt-eclogite transformation, causing the formerly buoyant slab to become more dense and sink (for example, Sacks, 1983). A flat slab can only be heated from below and around its edges, so eclogite formation might be expected to migrate in from the edges. An asthenospheric wedge probably already existed along the north edge of the flat slab where normal subduction was occurring throughout the Paleogene; slab weakening ($S > 1$) and eclogite formation may have occurred at about the same time there, with resultant extension and volcanism being synchronous. Weakening of the late Paleogene slab farther south beneath the Great Basin may have allowed extension before heating was sufficient for eclogite formation. If eclogite formation migrated south in front of an asthenospheric wedge, then the consequent volcanism would appear southward progressively later than extension, as observed.

In our view, the asthenospheric mantle beneath the Basin and Range province was largely passive and only began to influence crustal tectonism after the subducted lithosphere began to lose its thermal identity. In contrast to passive rifting as proposed by Sengor and Burke (1978), however, rifting in the Basin and Range was probably driven by gravitational collapse of previously overthickened crust rather than by changes in plate motions or interactions at plate margins. Crust east of the thrust belt experienced slab weakening without extending, because it lacked the gravitational instability of thick crust and the thermal weakening caused by thickening (for example, Glazner and Bartley, 1985).

CONCLUSIONS

Late Paleogene extension was common in the Great Basin and apparently formed two north- to north-northeast–trending extended belts before mid-Miocene time. Late Paleogene volcanism, and by inference, crustal magmatism, migrated generally southward so that extension and magmatism coincided only locally in both time and space. Thus, simple active or passive rifting models do not apply to the whole Great Basin, although magmatic inflation and weakening of the crust may be important on length scales of as little as 10–100 km. Interacting crustal and lithospheric processes are necessary to explain the disparate space-time patterns of extension and magmatism observed on the scale of the Great Basin or whole Basin and Range.

The structural state of the upper crust (extended or unextended) at the time of volcanism largely controlled the eruptive style of volcanism in the Great Basin. Differences in density profile of the crust, lithostatic stress at the levels of magma chambers, and structural condition of the crust probably control these differences. Within the late Paleogene extended belts, volcanism that was preextensional or early synextensional was dominated by effusive lava eruption. Postextensional or late synextensional volcanism in the southern parts of the belts was dominated by voluminous explosive ignimbrite eruptions. Southward migration of volcanism stalled against unextended crust south of the belts. Unextended crust east and west of the extended belts may have slowed the southward migration of volcanism and led to the convex-south form of the volcanic front. Space-time patterns of volcanism and extension in the southern Basin and Range also are consistent with a structural state of the crust that controlled eruptive style.

The timing and spatial distribution of late Paleogene extension in the Great Basin is consistent with extension driven by gravitational collapse of an over-thickened crustal welt relict from Mesozoic–early Tertiary time. This collapse may have been triggered by thermal weakening of subducted lithosphere north of the Mendocino fracture zone, as modeled by Severinghaus and Atwater (1990). Northward passage of the subducted Mendocino fracture zone probably triggered extension in the southern Basin and Range (Glazner and Bartley, 1984), because the subducted slab south of the fracture zone was younger, and it weakened or was assimilated into the asthenosphere earlier than the slab north of the fracture zone (Severinghaus and Atwater, 1990). Thermal weakening of the slab between the Mendocino and Surveyor fracture zones probably occurred last, in mid-Miocene time. This part of the slab lay beneath the southern Great Basin, which first extended in mid-Miocene time, later than areas to the north and south.

The geometry and strength of the subducted lithospheric slab beneath the Basin and Range may have been important controls on timing of tectonic and magmatic events there. Correlation of events in the Great Basin with the geometry of strong seismic(?) slab ($S = 1$; Severinghaus and Atwater, 1990) is good, but correlation with weak aseismic(?) slab ($S \geq 2$) is poor, suggesting that strength of subducted lithosphere may have been important in controlling crustal tectonism. The change from calc-alkaline volcanism to alkaline volcanism correlates well with slab geometry, but poorly with the onset of extension. These correlations provide geologic support for the slab geometries reported by Severinghaus and Atwater (1990).

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