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Basin and Range extensional tectonics at the latitude of Las Vegas, Nevada

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ABSTRACT

The Basin and Range province at the latitude of Las Vegas, Nevada (approximately 36°N), is ideally suited for reconstructing Neogene extension owing to an abundance of structural markers, primarily Mesozoic thrust faults, developed within the generally conformable Cordilleran miogeocline. In map view, extension is heterogeneous and is divisible into two major extensional domains, the Las Vegas and Death Valley normal fault systems, that lie east and west (respectively) of a relatively unextended median block. We determined horizontal relative-motion vectors between pairs of reference points across the province, chosen so as to best allow geologic markers to constrain the relative motion of the pair during extension. We recognize three sequences of pairs, two in the Las Vegas system and one in the Death Valley system, that define an unbroken path across the entire province. The vectors along these paths sum to give 247 ± 56 km of net extension oriented N73° ± 12°W.

Timing considerations indicate that extension occurred principally during the past 15 m.y. Westward motion of the Sierra Nevada away from the Colorado Plateau occurred at a rate of 20–30 mm/yr in the interval 10–15 m.y. ago, but was no greater than 10 mm/yr over the past 5 m.y. Strike-slip faulting was an important component in the extending system and absorbed perhaps 40–50 km of north-south shortening of the region during extension, indicating a constrictional strain field for the crust as a whole. If one assumes no major rotations of the Sierra Nevada during Cenozoic extension, and about 100 km of pre–15-m.y.-ago extension in the central portion of the northern Great Basin, the crust in the Las Vegas region extended by a factor of 3–4, whereas the wider Great Basin region extended by only a factor of 2. This difference may explain the contrast in regional elevation between the two areas (the northern Great Basin is on average about 1,000 m higher) and the constrictional strain in the Las Vegas region. The more widely distributed extension to the north may not have kept pace with the larger extension to the south, such that the south lost gravitational potential more rapidly. Thus, comparatively buoyant northern Great Basin lithosphere was (and continues to be) forced down the potential gradient into the Las Vegas region. Resolved parallel to the northern San Andreas fault, our reconstruction accounts for 214 ± 48 km of right-lateral shear along the Pacific–North America transform plate boundary.

INTRODUCTION

In the two decades since R. Ernest Anderson's first studies of large-magnitude Cenozoic extensional tectonism in the Lake Mead area of the Basin and Range, the significance of the structures described in his initial report on the area, published in the Geological Society of America *Bulletin* (Anderson, 1971), has grown from that of a freak occurrence of local importance to a benchmark in the recognition of how the crust extends. Simultaneous work in the late 1960s and early 1970s of Lauren Wright and Bennie Troxel (1973) in the Death Valley region, and of John Proffett (1977) in the Yerington mining district in west-central Nevada, led them independently to the same conclusion as that of Anderson, that in at least some regions of the Basin and Range, shallowly dipping normal faults separating steeply tilted fault blocks had accommodated large-magnitude extension of the continental crust in Neogene time. In addition, Armstrong's (1972) perceptive synthesis of low-angle faults in east-central Nevada showed that Cenozoic low-angle faulting was important over a wide region of the Basin and Range.

These studies were not the first to recognize structures now widely believed to accommodate large-magnitude extensional tectonism. Ransome and others (1910) recognized the structural style of imbricate normal faulting of Tertiary volcanic strata in the Bullfrog mining district near Death Valley and the fact that the entire faulted package lay tectonically upon a metamorphic complex. They developed the hypothesis that the basal fault was normal and of significant displacement but favored the interpretation that the fault was an overthrust. They considered it unlikely that the force of gravity alone could have moved the volcanic strata on such a shallowly dipping fault and suggested that the undulose geometry of the basal fault facilitated extensional faulting in the hanging wall of the overthrust. Subsequent works, notably Noble (1941), Longwell (1945), Curry (1954), Young (1960), Misch (1960), Pashley (1966), and Hunt and Mabey (1966), all recognized similar features in the Basin and Range, yet as in the case of Ransome and others (1910), none of them concluded that the deformation resulted from large-magnitude Cenozoic extension (although Hunt and Mabey argued for Mesozoic extension). In reading these older works today, one simultaneously feels a sense of loss over how long it took to begin to understand extensional tectonism, and elation over the fact that there is still so much exciting work to be done.

In the wake of Anderson's and other studies that compellingly demonstrated large-magnitude extension, strong sentiment against low-angle

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normal faulting as a major mechanism of extension in the Basin and Range province and elsewhere remained, with a tendency to consider areas of shallowly dipping normal faults as exceptional to an over-all style of steeply dipping, widely spaced normal faults that accommodated relatively modest crustal extension (10%–30% increase over original width). Explanations excluding crustal extension, including surficial gravity sliding, special circumstances during thrust faulting, or a combination of the two, were still often invoked to explain the enigmatic low-angle faults. These explanations defended the classical view (for example, of G. K. Gilbert) of a Basin and Range that was folded and thrust faulted in Mesozoic time, blanketed by ignimbrite in early to middle Tertiary time, and block faulted on steep faults in the late Tertiary. The observations of large faults that place high crustal levels on low, and the involvement of steeply dipping Tertiary strata along them, flew in the face of Gilbert's Basin and Range. The lukewarm reception of a decidedly non-Gilbertian Basin and Range is exemplified in the citations of Anderson (1971), Armstrong (1972), Wright and Troxel (1973), and Proffett (1977, but submitted to the *Bulletin* in 1972) shown in Figure 1. Most of this work had been completed and reported on at Geological Society of America meetings by 1972; yet, it was nearly a decade before its importance was widely realized in the geological community. Stewart (1978) best summarized the thinking on the province in the late 1970s, when it was thought that locally large-magnitude extension had been accommodated in areas such as the Yerington district but that most of the province probably had not extended more than about 10%–35%. In contrast to the prevalent view, the mobilistic syntheses of Hamilton and Myers (1966) and Hamilton (1969) argued for a doubling in width of the province, based on crustal structure and thickness, the possible distortion of pre-Cenozoic tectonic belts by extension, and the possibility that steep range-bounding faults flatten downward.

The sudden appreciation of the significance of these early studies was catalyzed by the 1977 Penrose Conference on Cordilleran Metamorphic Core Complexes, at which a number of workers argued that widespread regions of metamorphic tectonite in the Basin and Range were Tertiary in age and related to low-angle faulting and crustal extension (Davis and Coney, 1979; articles in Crittenden and others, 1980), first hypothesized for east-central Nevada by Armstrong (1963, 1972). The provocative reflection profiles from the starved passive margin of the Bay of Biscay (for example, de Charpal and others, 1978) and the arguments of McKenzie (1978) for major crustal extension in rifts and on passive margins also contributed to this appreciation. These insightful field studies had ushered in a new era of Basin and Range *observation*, unencumbered with the need to explain away, case by case, the first-order field relations of the province as flukes.

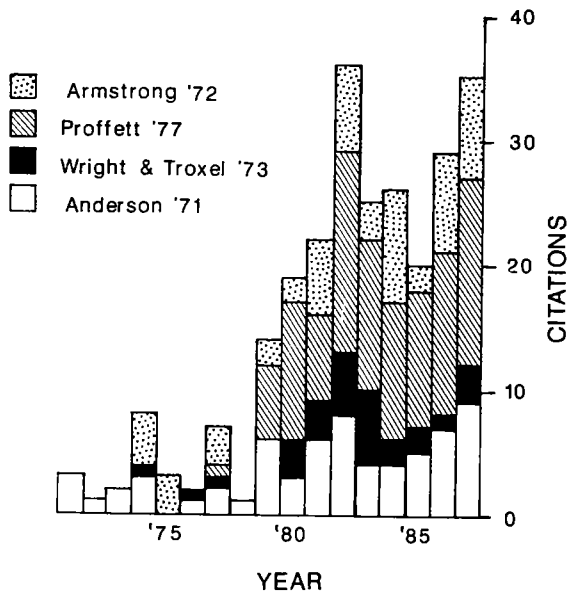

Figure 1. Number of citations in refereed journals of articles discussed in text. Note tenfold increase from 1979 to 1982, roughly a decade after studies were completed.

This paper is a progress report summarizing the results and implications of field studies in the southern Great Basin region by the Program in Extensional Tectonics at Harvard. We build on Anderson, Wright, and Troxel's studies in the region to present for the first time measurement of Cenozoic extensional strain across the entire Basin and Range, accurate to two significant figures, and demonstration of the slowing of extension between 10 m.y. ago and the present. In an earlier report on the region, Wernicke and others (1982) used offsets on selected strike-slip faults to reconstruct the extension, concluding that at least 140 km of extension had occurred. In this report, we incorporate new data into reconstructing the province, in particular an improved understanding of the northern Death Valley–Furnace Creek fault zone, which was not considered in Wernicke and others (1982). Using a similar but more detailed approach, we present our reconstruction as a series of vectors that describe the motion between fiducial points such that we can quantify (that is, bound uncertainties on) the magnitude, direction, and rate of extension of a number of subregions of the province, then sum the vectors and their uncertainties to place bounds on the westward motion of the Sierra Nevada block relative to the Colorado Plateau in Neogene time (compare with Minster and Jordan, 1984, 1987). The vectors are based on palinspastic reconstructions of areas mapped by us and many other workers. Documentation of the field relations of each reconstruction is well beyond the scope of a single paper and is presented elsewhere (Axen and Wernicke, 1987; Wernicke and others, 1988a, 1988b; J. K. Snow and B. Wernicke, unpub. data; numerous reports by other workers cited below), but the key structural markers in each are outlined below. Our results indicate substantially more extension than the 140-km minimum determined by Wernicke and others (1982) and have important implications for the nature of extensional tectonism.

GEOLOGIC SETTING

The Basin and Range province near the latitude of Las Vegas has an ideal regional tectonic setting for a province-wide reconstruction of Cenozoic extension (Fig. 2). The pre-extension geology is more straightforward than at other latitudes because the regionally conformable Cordilleran miogeocline is exposed across the entire width of the province (Figs. 2 and 3). The miogeocline is disrupted by east-vergent Mesozoic thrust faults that make local reconstructions more complicated than they might be in the absence of faults. The thrusts, however, are distinctive enough and the extensional separation of crustal blocks great enough that they provide the markers necessary to tightly constrain large-scale reconstructions. The thrusts are thus more an aid than a complication, for discrete markers within the miogeoclinal section are few, and in most cases, the determination of fault offsets based purely on isopachs and facies trends is limited by the uncertainty in their precise location and with the assumption of their initial geometry (for example, Stewart, 1983; Prave and Wright, 1986). The great thickness of the thrust-faulted miogeocline gives excellent depth control on cross-sectional reconstructions, locally in excess of 15 km (Fig. 3). Exposure is generally excellent in the region because it lies at low elevation and in the rain shadow of the Sierra Nevada and more southerly

ranges. In addition, much of the geology is developed within carbonate rocks, which crop out well in desert regions.

The regionally averaged topographic pattern of the Basin and Range at the latitude of Las Vegas is one of high flanks, comprising the Sierra Nevada on the west and the Colorado Plateau on the east, and two broad, low-lying areas on either side of a median high (Eaton and others, 1978). This pattern resembles that of the northern Basin and Range, but at smaller scale because the province here is half the width (Fig. 2). The median high is centered on the Spring Mountains, Sheep Range, and Las Vegas Range, whereas the lows include the Colorado River trough/Lake Mead area on the east and the Death Valley region on the west (Fig. 4). As discussed below, the two low-lying areas are highly extended, whereas the median high is less extended.

Basement, Proterozoic Basin, and Miogeoclinal Wedge

Precambrian Y (mostly ca. 1.7–1.4 Ga) crystalline basement in the region lies nonconformably beneath unmetamorphosed sediments of Precambrian Y (?), Precambrian Z, or Cambrian age (Fig. 3). Precambrian Y (?) and Z strata of the Pahrump Group (Fig. 3) are locally present in ranges west and southwest of the Spring Mountains between basement and regionally persistent Precambrian Z to Cambrian strata that form the base of the Cordilleran miogeocline in the region (Stewart, 1970, 1972). Although the lower portion of the Pahrump is probably Precambrian Y in age, the upper part appears to be in gradational contact with the Cordilleran miogeocline, and thus is probably Precambrian Z in age (Miller, 1987). The west-thickening Precambrian Z and Paleozoic miogeocline (Figs. 2 and 3) is overlain disconformably or with mild angular unconformity by locally thick accumulations of Mesozoic strata (Fig. 3).

The most significant stratigraphic feature beneath the miogeoclinal strata is the northward pinchout of the Pahrump Group in the southern Death Valley region (Wright and others, 1974, 1981). South of the pinchout, as much as 3,000 m of Pahrump strata is present below the basal units of the miogeocline in the southern Black Mountains, Kingston Range, and Panamint Range (Fig. 4). Over a distance of less than 10 km, the basal miogeoclinal unconformity cuts downsection through the Pahrump Group and onto crystalline rocks.

Lithologically, the miogeocline is divisible into two main parts, including a Middle Cambrian and older clastic wedge and a Middle Cambrian and younger carbonate succession (Fig. 3). The clastic wedge thickens from less than 100 m on the craton to the east, where basal strata are Lower Cambrian, to more than 5,000 m in western areas, where most of the sequence lies below basal Cambrian beds. The Paleozoic sequence is entirely marine, except for some Permian strata that are partly nonmarine (Wright and others, 1981; Stone and Stevens, 1987). Westward thickening of the carbonate succession occurs in part by thickening of individual units and in part by the pinching in of Ordovician, Silurian, and Devonian strata beneath a major sub–Upper Devonian disconformity (Fig. 3). On the craton, Upper Devonian strata lie disconformably on Upper Cambrian. To the west, they lie on progressively younger strata until a fully developed Ordovician, Silurian, and Devonian section is present. The youngest marine strata in the region are Triassic and are overlain in eastern areas by nonmarine clastics locally as young as Cretaceous and in western areas by lower Mesozoic volcanics (for example, Wright and others, 1981). In sections in the transition zone between craton and miogeocline, the highest Paleozoic strata present on the craton, including the Kaibab and Toroweap Formations, pinch out westward beneath the basal Mesozoic unconform-

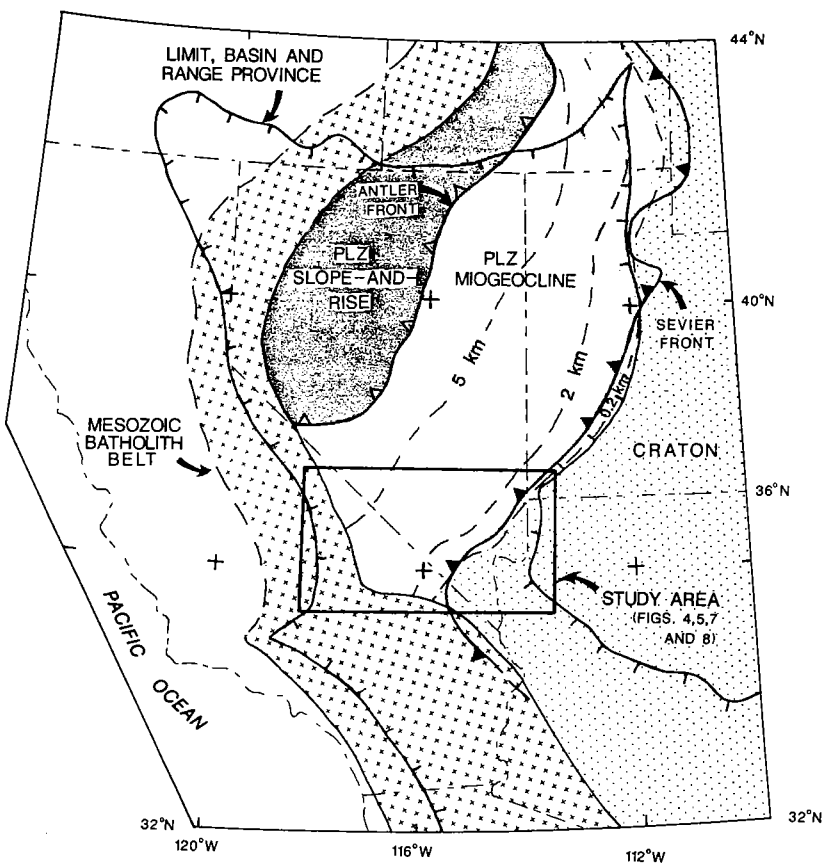

Figure 2. Regional tectonic setting of the Las Vegas area Basin and Range, showing isopach trends of the Precambrian Z–Cambrian clastic wedge of the Cordilleran miogeocline, Paleozoic Antler and Mesozoic Sevier orogenic fronts, and the position of the Mesozoic batholith belt (crosses). Note that the position of the study area of this report resides largely in the miogeoclinal prism and craton.

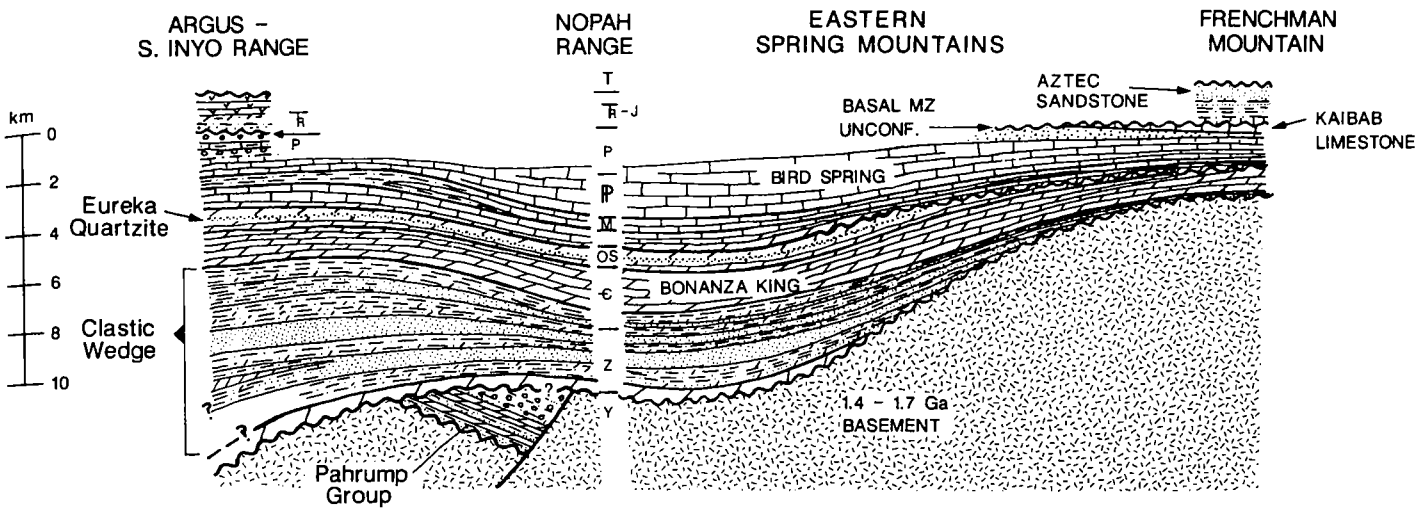

Figure 3. Highly simplified stratigraphic cross section of Mesozoic and older rocks exposed in the region. Modified from Stewart (1970) and Wright and others (1981). See Figure 4 for locations.

ity (Fig. 3; for example, Tschanz and Pampeyan, 1970; Burchfiel and others, 1974). In westernmost sections of the miogeocline, complex unconformities and facies changes in Carboniferous and Permian strata indicate Permian tectonism (Stone and Stevens, 1984, 1988a).

The primary lateral facies changes within the miogeocline include a transition from quartzite and siltstone in eastern exposures of the Precambrian Z–Cambrian clastic wedge to predominantly shale and carbonate in the west (Stewart, 1970), and a transition from mostly shallow marine limestone in Carboniferous and Permian strata in eastern areas to locally deep-marine shale, sandstone, and limestone in the west (Dunne and others, 1981; Stone and Stevens, 1988b; Fig. 3). The increase in fine clastics and carbonate in the clastic wedge indicates a transition from shelf to slope-and-rise facies (Stewart, 1972), but early Paleozoic slope-and-rise deposits are not present east of the Sierran batholith at this latitude (Fig. 2). The westward increase in clastics in Carboniferous strata probably represents the distal effects of the mid-Paleozoic Antler orogeny (for example, Dunne and others, 1981), which farther north in central Nevada is expressed by the eastward thrusting of early Paleozoic slope-and-rise facies strata onto the shelf facies rocks, forming a broad, asymmetrical foreland basin (for example, Poole and Sandberg, 1977). Structural effects of the Antler and Permian-Triassic Sonoma orogenic events may be present in western portions of the region (for example, Nelson, 1981), including possible truncation of the continental margin in Permian and Triassic time (Burchfiel and Davis, 1981; Stone and Stevens, 1988a, 1988b).

Mesozoic Thrust-and-Fold Belt

Mesozoic structures in the region are predominantly east-vergent folds and thrusts. The region lies at a major transition zone in the Mesozoic Cordilleran thrust belt. Within the region and to the north, the thrusts are developed within the miogeoclinal wedge such that many of the thrusts cut slowly through it, in many cases with ramp-flat geometries. To the south, however, the thrust belt changes trend from north-northeast, parallel to isopach and facies trends in the miogeoclinal wedge, to southwest, parallel to the east margin of the Mesozoic batholiths (Figs. 2 and 5). In so doing, the thrust belt leaves the miogeoclinal wedge, and the thrusts lose their ramp-flat geometry, tending instead to cut more steeply across the layers of cratonic sediments and involve crystalline basement (for example, Burchfiel and Davis, 1975, 1981). Thus, within any given thrust plate, the stratigraphic section thins from north to south as the thrusts curve southward out of the miogeocline. In the Spring and Clark Mountains blocks, the thrusts tend to converge upon one another to the south (Burchfiel and Davis, 1975, 1981).

Throughout most of the region, exposures of pre-Tertiary rock are separated by alluviated valleys, hampering correlations of thrusts between ranges (Fig. 5). In the median, relatively unextended block, and within a number of highly extended blocks, the nature of the thrust belt can be deduced across three or four major thrust plates (Figs. 5 and 6). Thrusts that now lie in widely separated blocks may be correlated by examining each of these fragments. As such, the thrusts represent useful markers for the reconstruction of Cenozoic tectonism. Below, we describe the principal characteristics of each of the Mesozoic thrust plates from east to west across the province (Figs. 5 and 6). Although somewhat tedious, these details are of great importance in measuring the directions and magnitudes of Cenozoic displacements discussed in the following section.

Keystone System. The lowest major thrust is the Keystone thrust and correlatives, which can be traced for more than 200 km along strike (for example, Burchfiel and others, 1982). Its hanging wall is characteristically a décollement in Middle Cambrian dolostones of the Bonanza King Formation (Fig. 3). The footwall geology of the thrust is complex, locally including blocks of a lower, older thrust similar to the Keystone. The thrust has a regionally persistent ramp and hanging-wall ramp syncline or synclinorium that is cored with Mesozoic strata. Along its entire trace, the hanging wall contains a portion of the miogeocline characterized by (1) rapid westward pinchin of Ordovician strata beneath the sub–Upper Devonian disconformity and (2) the westward pinchout of the Kaibab and Toroweap Formation beneath the basal Mesozoic unconformity (Figs. 3 and 6).

The stratigraphic and structural uniformity of the Keystone system, and the fact that it is continuous for large distances within range blocks, suggests strong stratigraphic control on the thrust. Subtle details of the miogeoclinal stratigraphy change little in character and trend along strike. The only major break in the trend of the Keystone system occurs along the Las Vegas Valley shear zone (Figs. 4 and 5), where geologic lines defined by the intersection of the ramp with footwall Mesozoic strata are appar-

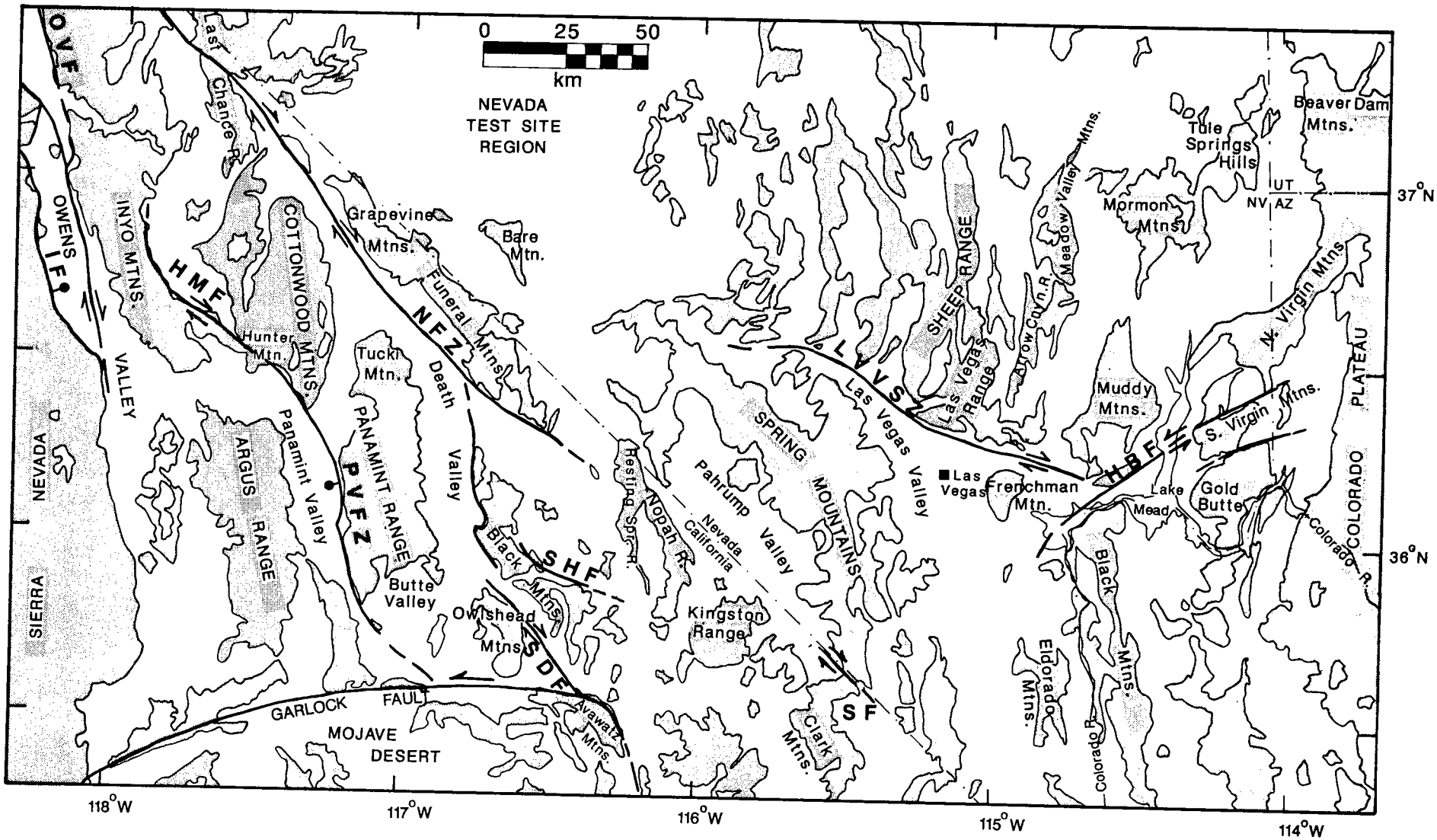

Figure 4. Location map showing geographic features mentioned in text, outcrop area of pre-Cenozoic rocks (shaded), and selected Cenozoic fault zones discussed in text (bold lines). HBF, Hamblin Bay fault; HMF, Hunter Mountain fault; IF, Independence fault; NFZ, northern Death Valley–Furnace Creek fault zone; OVF, Owens Valley fault; PVFZ, Panamint Valley fault zone; SDF, southern Death Valley fault zone; SF, State Line fault; SHF, Sheephead fault.

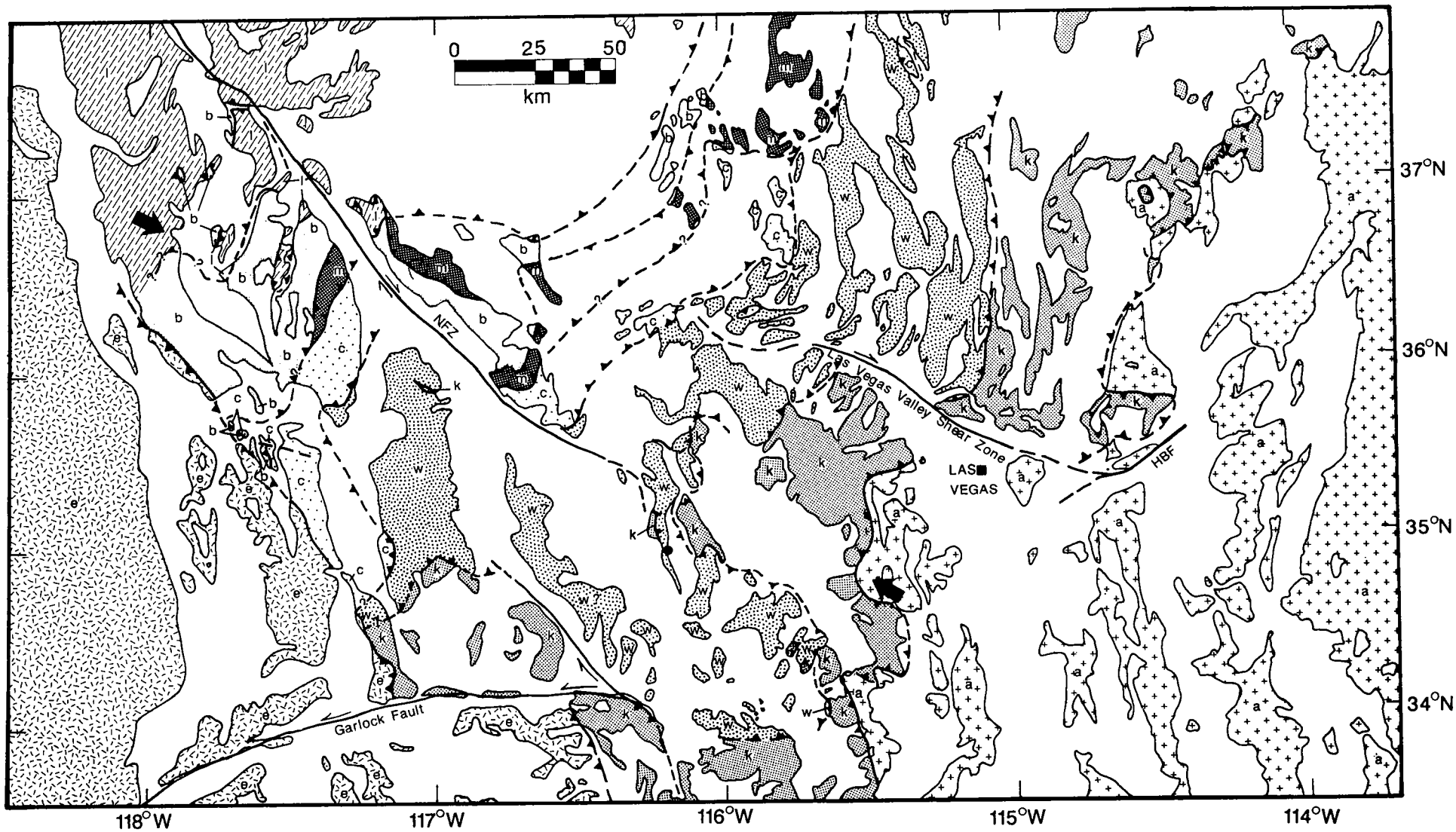

Figure 5. Map of pre-Tertiary outcrop area, showing Mesozoic structural levels. Contacts between units are not necessarily thrust faults and include later faults that juxtapose thrust plates and the axial traces of anticlines in asymmetric fold pairs; some areas in westernmost region are intruded post-tectonically. a = sub-Keystone system, autochthonous and parautochthonous rocks, k = rocks above Keystone thrust system and below Wheeler Pass system, w = rocks above the Wheeler Pass system and below the Clery thrust, c = rocks above the Clery thrust and below the Marble Canyon thrust, m = rocks above the Marble Canyon thrust and the west-vergent White Top Mountain backfold/thrust system, b = rocks below the White Top Mountain structure and the Last Chance system, l = rocks above the Last Chance system, e = rocks above the East Sierran thrust system. Bold arrows show approximate line of restored section in Figure 6. Arrow on west-vergent structure in Grapevine Mountains shows sense of rotation required to realign it with other exposures of the structure. HBF, Hamblin Bay fault; NFZ, northern Death Valley–Furnace Creek fault zone.

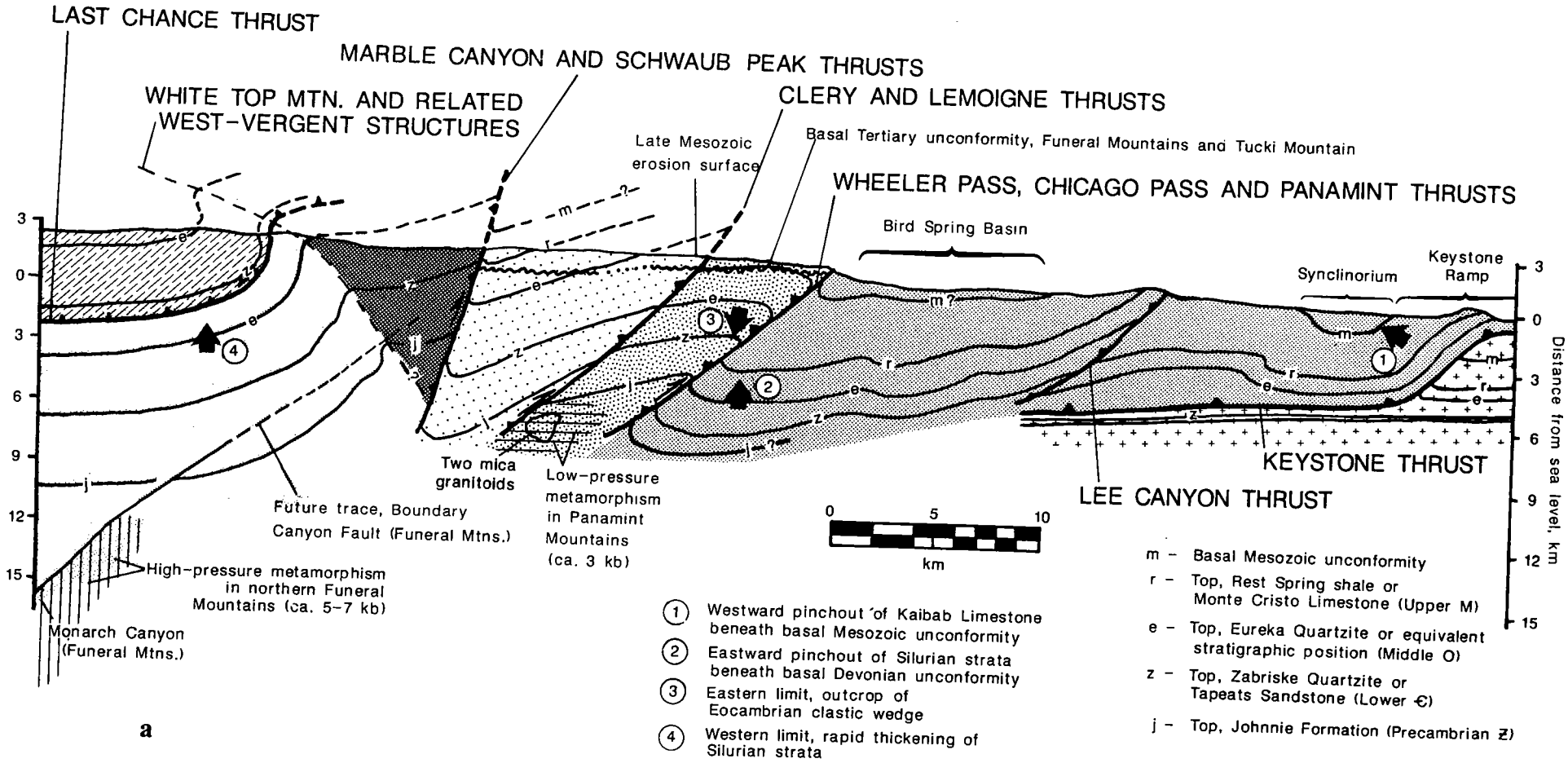

Figure 6. Restored cross section between bold arrows in Figure 5, showing geometry of thrust belt prior to Cenozoic extension. a. Patterned after Figure 5; pressure estimates of metamorphic terrains from Labotka and Albee (1988).

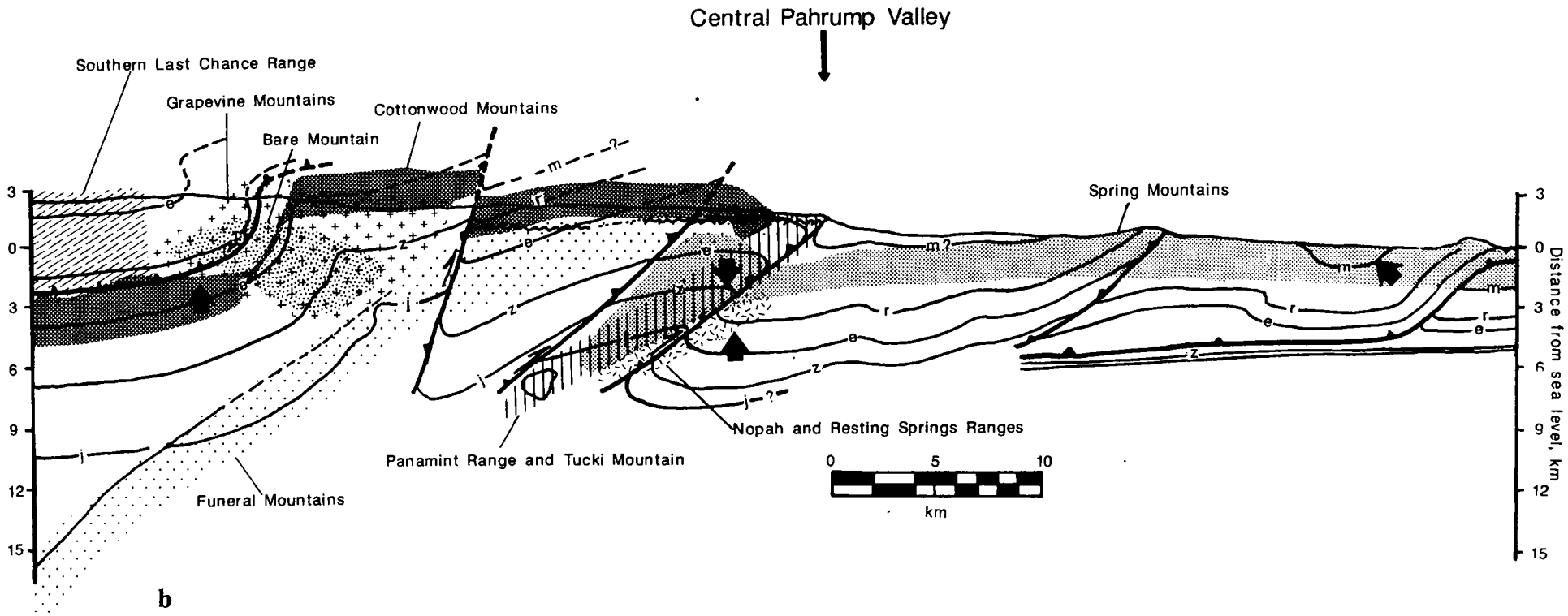

Figure 6. (*Continued*). b. Patterned to show distribution of structures within various range blocks used to construct the section. All geology is taken from within 20 km of the line between the bold arrows. Patterned areas depict structural levels exposed in ranges at present erosion levels. Overlapping ranges occur on either side of the NFZ, providing independent constraints on the section (see Fig. 4 for geographic locations).

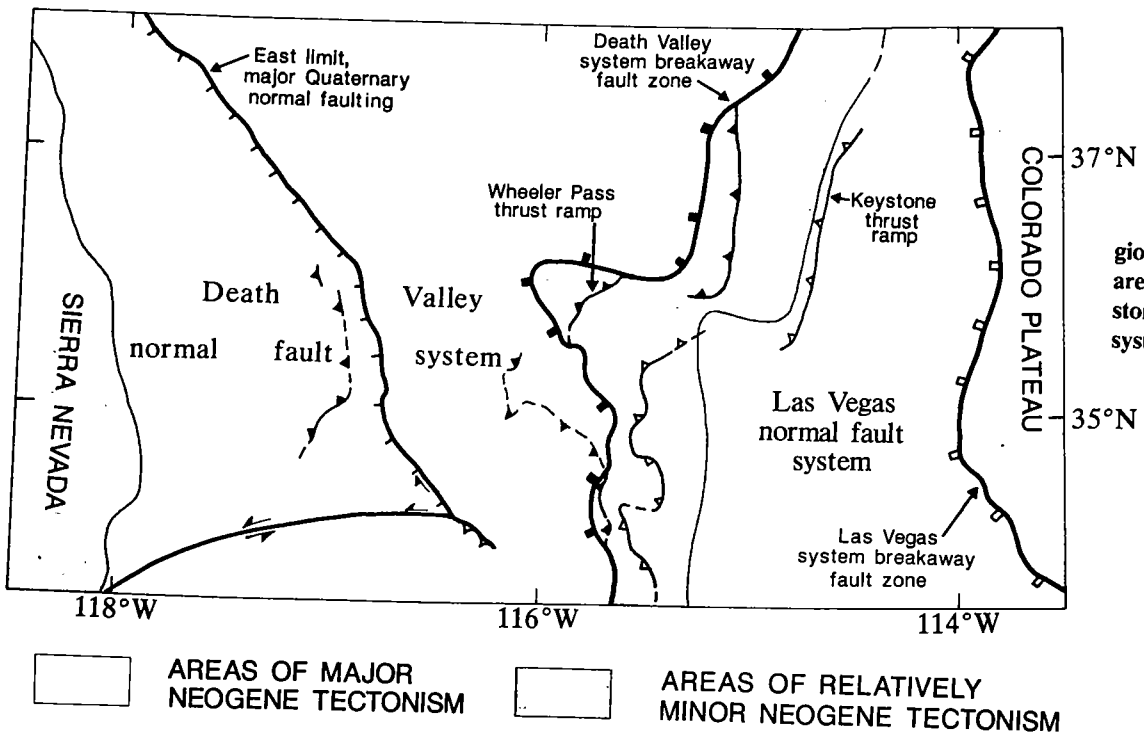

Figure 7. Map showing regional distribution of extended areas with respect to the Keystone and Wheeler Pass thrust systems.

ently offset right laterally about 50 km (Fig. 5). The likelihood that this offset has an origin as a tear structure in the Keystone thrust plate (for example, Royse, 1983) or was controlled in some way by an abrupt change in trend of isopachs in the miogeocline seems remote in view of the similarities of the thrust system north and south of the shear zone. The strong control of facies and isopach trends on thrust geometry observed in most thrust belts suggests that major paleogeographic anomalies are usually associated with major transverse structures, with a change in both the character of the faulted sediments and in the number and spacing of thrusts across them (for example, Price, 1981).

West of the synclinorium in the hanging wall of the Keystone, the Paleozoic section thickens by thickening within individual units and also by the appearance of the Middle Ordovician Eureka Quartzite, Upper Ordovician Ely Springs Dolomite, and Silurian strata below the sub-Devonian unconformity and of Middle and Lower Devonian strata above the unconformity (Figs. 3 and 6; Burchfiel and others, 1974). Within this region, there are a number of relatively small east-vergent folds and thrust faults that carry units as old as the Middle Cambrian Bonanza King Formation over rocks as young as the Carboniferous-Permian Bird Spring Formation, such as the Lee Canyon thrust (Fig. 6a). These structures tend not to be as laterally persistent as the Keystone system. West of the synclinorium, the Bird Spring Formation thickens from about 600 to >2,000 m (Figs. 3 and 6; Burchfiel and others, 1974).

Wheeler Pass System. The next highest major thrust system is the Wheeler Pass system. It can be traced for nearly 120 km along strike in the Sheep Range, Las Vegas Range, and Spring Mountains, interrupted only by the Las Vegas Valley shear zone (Fig. 5; Longwell and others, 1965; Burchfiel, 1965; Guth, 1981). Unlike the Keystone system, the Wheeler Pass system is not continuously exposed within the Spring Mountains–Clark Mountains block (Fig. 5). In its northern exposures in the Spring Mountains, the thrust strikes at high angle to the boundary between the Spring Mountains block and the highly extended area to the west (Fig. 7), projecting into the alluvium of Pahrump Valley (Figs. 4 and 5). Southwest of the Spring Mountains, in the highly extended Death Valley region, the thrust is present in a number of range blocks and is found as far west as the Panamint Range (Wernicke and others, 1988a, 1988b; Figs. 4 and 5). To the south, the thrust system reappears in the Clark Mountains block (Fig. 5).

At present erosion levels, the thrust usually carries Precambrian Z clastics over the Bird Spring Formation (Figs. 3 and 6). The thrust is probably in part a décollement in northern areas (for example, Guth, 1981; Burchfiel and others, 1974), but south of the Spring Mountains, it typically is not a hanging-wall décollement within miogeoclinal strata, as it cuts rapidly through the miogeoclinal section (for example, Burchfiel and Davis, 1971; Burchfiel and others, 1983; Wernicke and others, 1988b). At structurally deep levels, where hanging-wall basement overrides Precambrian Z clastics, the thrust has a décollement geometry (Burchfiel and Davis, 1971).

The Wheeler Pass system is the lowest thrust plate to contain exposures of the Precambrian Z clastic section and underlying basement and Pahrump strata (Fig. 6). It carries the thinnest sections of these strata, which thicken rapidly westward from about 2,000–2,500 m immediately above the thrust to more than 5,000 m to the west (Figs. 3 and 6). Silurian strata pinch in just beneath the thrust in the Nopah Range area (Figs. 4 and 6), but farther north, the Silurian is present well to the east of the thrust plane (for example, Langenheim and others, 1962). Maximum thicknesses of Silurian strata are found only in higher thrust plates (Fig. 6a).

The features that distinguish the Wheeler Pass system include its structural style, position in the miogeocline, and the fact that it is the only thrust in the region that emplaces rocks as old as Precambrian Z on top of post-Mississippian strata, with the exception of portions of the Last Chance system (described below), which is clearly a structurally higher system. The stratigraphic throw of the Wheeler Pass system is consistently about 5,000 m.

Higher Thrusts and Backfold. Above the Wheeler Pass system, we recognize a belt of two east-vergent structures (Clery-Lemoigne and Marble Canyon–Schwaub Peak thrusts) and a third, structurally higher west-vergent structure (White Top Mountain and related west-vergent

structures, Figs. 5 and 6a; Reynolds, 1974; J. K. Snow and B. Wernicke, unpub. data). Correlations of these structures are difficult because they all lie west of the unextended, median topographic high and are obscured by extensive Tertiary volcanic cover in the Nevada Test Site region (Figs. 4 and 5). Nonetheless, the sequence is found continuously exposed in each of two range blocks in the Death Valley region, the Funeral-Grapevine Mountains block and the Cottonwood Mountains block (Figs. 4 and 5; J. K. Snow and B. Wernicke, unpub. data). Despite the large distance between the two blocks, the three structures have similar stratigraphic throw, position in the miogeocline, and relative spacing. As shown in Figure 6b, the geology of the Funeral-Grapevine Mountains block and the Cottonwood Mountains block fits together well on the same pre-extensional cross section. The presence of the west-vergent structure is particularly diagnostic of their correlation, as it is the only major west-vergent structure in the Death Valley region (J. K. Snow and B. Wernicke, unpub. data). All three structures characteristically cut upsection rapidly in both hanging wall and footwall and have a stratigraphic throw of 2,000 to 3,000 m. They occupy a position in the miogeocline characterized by rapid increase in thickness of the Silurian section as it pinches in beneath the sub-Devonian unconformity, and the transition from carbonate facies to shale, sandstone, and limestone facies in Carboniferous strata (Figs. 3 and 6a).

Last Chance System. Structurally above the west-vergent system, a major thrust system carries the thickest sections of the Precambrian Z and Cambrian clastic wedge over Carboniferous shale, sandstone, and limestone (Stewart and others, 1966; Reynolds, 1974). The Last Chance system differs from structurally underlying thrusts in the Death Valley region in that it commonly is a décollement in both hanging wall and footwall and has nearly twice the stratigraphic throw (generally 5,000–6,000 m). There are numerous windows into Carboniferous strata throughout the Last Chance Range–Inyo Mountains area that show that the thrust cuts gradually downsection in Precambrian Z strata to the west (Stewart and others, 1966; Nelson, 1981). The transition from quartzite and siltstone facies to shale and carbonate facies of the Precambrian Z clastic wedge (Fig. 3) occurs within the hanging wall of the thrust, although the onset of the change may occur in the footwall (Stewart, 1970).

East Sierran Thrust System and Sierran Batholith. The eastern margin of the Sierra Nevada batholith and a coincident zone of thrust faults trend about N30°W across the western part of the region, cutting obliquely across the northeast-trending isopachs, facies lines, and thrust faults developed in the miogeoclinal wedge (Fig. 5; Dunne, 1986). The East Sierran system was apparently localized by the thermal contrast between the batholith and cooler lithosphere to the east (for example, Burchfiel and Davis, 1975). It is younger than the higher thrusts developed in the miogeocline, as a suite of Early Jurassic alkalic plutons cuts the miogeoclinal thrusts, whereas younger plutons of the batholith are cut by strands of the East Sierran system (Dunne, 1986). The hanging wall of the system seems to override progressively lower thrust plates southward, but the large proportion of plutonic rock in the hanging wall of the thrust system precludes identification of offset traces of the older thrusts. For a discussion of relative and absolute timing constraints on the thrust systems in the region, the reader is referred to Dunne and others (1978), Burchfiel and Davis (1981), and Dunne (1986).

CENOZOIC EXTENSION

The first-order pattern of extensional tectonism is that of two topographically low regions pervaded by down-to-the-west normal faults, separated by a central unextended block (Fig. 7). The system to the east of the unextended block is herein referred to as the "Las Vegas normal fault system" and the system to the west as the "Death Valley normal fault system" (Fig. 7). The Mesozoic structure of the region is in part reflected in the position of these extended domains. The Las Vegas system is developed almost entirely below the Keystone thrust system and involves crystalline basement (Anderson, 1971); major normal faults involve the upper plate of the thrust only at highest structural levels in the northern part of the region (Wernicke and others, 1984, 1985; Smith and others, 1987). The east limit of major extension, or breakaway zone, for the Las Vegas system is developed within cratonic strata but, as is evident from extension magnitudes discussed below, initially lay no more than a few tens of kilometers east of the Keystone system.

The breakaway zone for the Death Valley system closely follows the trace of the Wheeler Pass thrust system, in some places leaving the thrust behind on the stable block, in others carrying it within extensional allochthons more than 100 km to the west (Figs. 5, 6, and 7). It is this fact in particular that affords considerable precision in reconstructing the Death Valley extensional terrane. The localization of the two principal breakaway zones near the Keystone and Wheeler Pass systems leaves a stable terrane between them composed of the Keystone and higher thrust plates that lie below the Wheeler Pass system (Fig. 7).

Our strategy for constraining both local and province-wide extension is to determine horizontal relative motion vectors V_i between the ith pair of reference points, which are chosen so as to best allow geologic markers to constrain the pair's relative motion. Any sequence of n pairs that defines an unbroken path between endpoints then defines the relative motion V_e between the endpoints (compare with Minster and Jordan, 1984, 1987).

$$\sum_{i=1}^{n} V_i = V_e$$

We ignore curvature of the Earth and vertical motions of the reference points, as they are negligible in comparison with the horizontal offsets.

The reference points and key geologic markers are shown in Figure 8. Below, we discuss constraints on the relative motion between pairs of points that define paths suitable for both local and province-wide reconstruction, summarized in Table 1. We emphasize determination of the uncertainties in each of the vectors, which in most cases are based on simple strain compatibility arguments, principally the condition that geologic markers do not overlap in the reconstruction.

Las Vegas System

The Las Vegas system is composed predominantly of southwest-directed normal faults. Geologic data allow constraint of the motion between the Spring Mountains block and the Colorado Plateau using two independent paths, including point pairs A1A2, A2A3, and A3A4, in the Lake Mead area, and C1C2, C2B1, and B1B2, in the Mormon Mountains–Las Vegas Valley area (Fig. 8).

Reconstruction via Lake Mead Fault System. Anderson (1971) and Anderson and others (1972) first recognized the large-magnitude extension in the Lake Mead area and concluded that most of the deformation occurred between 15 and 11 Ma, although the precise magnitude was not determined. Anderson (1973) and Bohannon (1979, 1984) also suggested that large-magnitude strike-slip faulting was present in the region, indicating about 65 km of translation of the Frenchman Mountain block southwestward away from the Virgin Mountains area (V_{A2A3}; Figs. 4 and 8). The evidence for offset includes (1) a distinctive stratigraphic sequence in basal Tertiary (Miocene) sedimentary rocks (Bohannon, 1979, 1984); (2) the geometry of the basal Tertiary unconformity, which at Frenchman Mountain and in the Virgin Mountains, gradually cuts out Mesozoic section when followed from north to south (Bohannon, 1979, 1984); and (3) the presence of numerous landslide breccias within the Miocene section in the Frenchman Mountain block, composed of at least 14 differ-

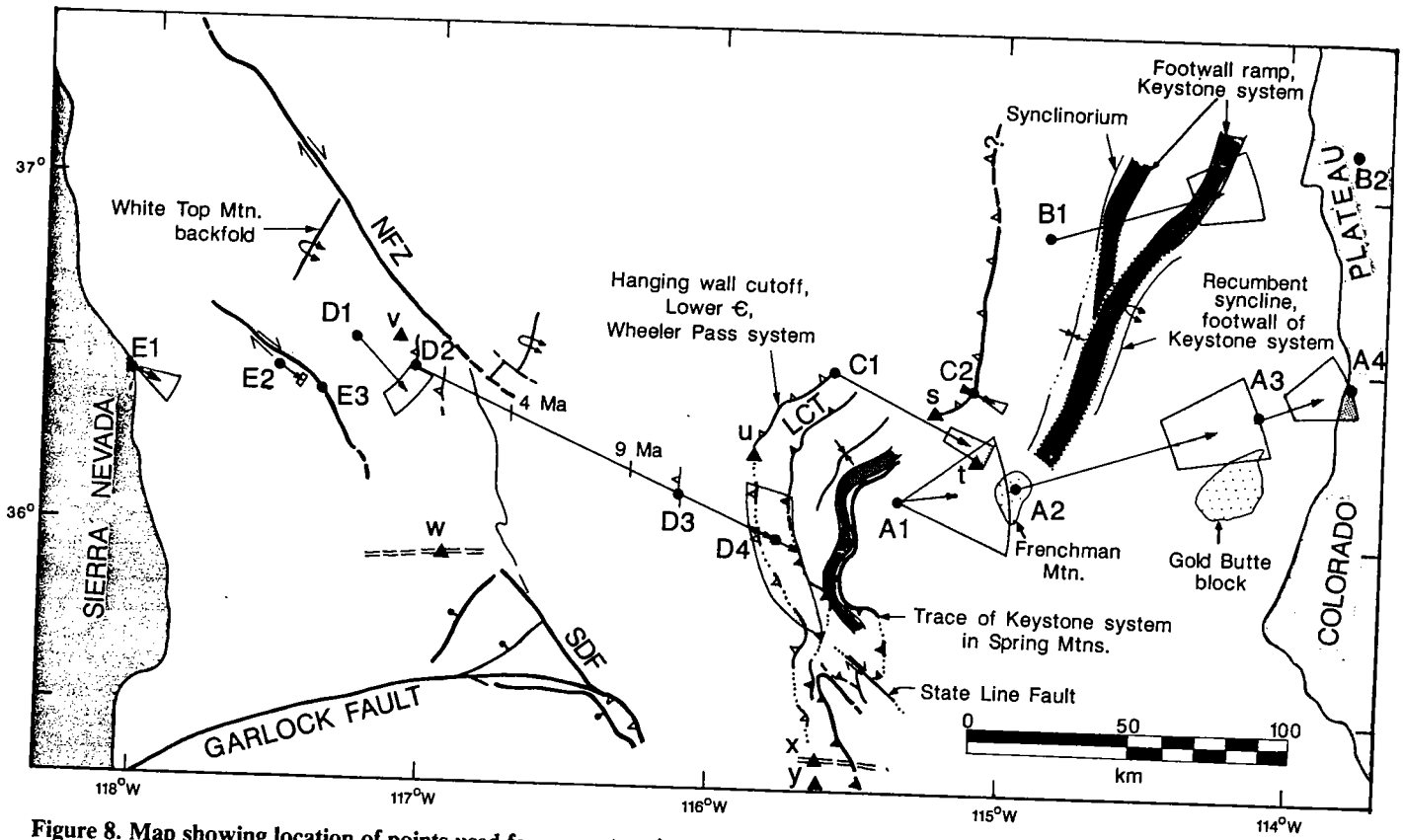

Figure 8. Map showing location of points used for reconstruction paths, key structural features used to constrain offsets, and "best-fit" restoration vectors and their individual uncertainties. Small-lettered points (triangles) are reference points for strain compatibility arguments discussed in text. Capital-lettered points (filled circles) each belong to one or more point pairs used for reconstruction paths. Dotted lines, structural features projected beneath Quaternary alluvium; light shading, uncertainty regions of vectors in Table 1; note irregular geometry of V_{D2D4}; double dashed lines, northern limits of abundant Mesozoic plutons in southern Panamint Range and Clark Mountains area; NFZ, northern Death Valley–Furnace Creek fault zone; SDF, southern Death Valley fault zone; LCT, Lee Canyon thrust. Doubling of northernmost exposures of footwall ramp in Keystone system is due to offset on Mormon Peak detachment. Vertical lines on segment D2D3 show western limit of Tertiary strata that overlap major extensional structures at times indicated (from Wright and others, 1984). See Table 1 for numerical values of vectors, and text for discussion of their derivation.

ent rock types that match those seen in the Gold Butte block, many of which are not common in other crystalline blocks exposed in the region (Figs. 3 and 8; Anderson, 1973; Longwell, 1974; Parolini and others, 1982; Parolini and Rowland, 1988).

The Frenchman Mountain block must have been in a position close enough to the Gold Butte block to receive megabreccia deposits for which the transport direction was to the north (Anderson, 1973; Longwell, 1974; Parolini and Rowland, 1988). Restoration of Frenchman Mountain to a position north of the Gold Butte block is also supported by the southward pinchout of the Jurassic Aztec Sandstone and Triassic Chinle Formation beneath the basal Tertiary unconformity in the Frenchman Mountain block and in the fault blocks in the South Virgin Mountains just to the north of Gold Butte (Bohannon, 1979, 1984). On the basis of the proximal-channel facies of the megabreccias (Parolini and Rowland, 1988), a distance of no more than 10 km between the Frenchman Mountain and the Gold Butte blocks prior to extension is likely, giving a minimum of 50 km of west-southwest relative motion between the two (lower bound of displacement for V_{A2A3}; Fig. 8). Palinspastic reconstruction of the Gold Butte and other blocks in the South Virgin Mountains (for example, Wernicke and Axen, 1988), however, requires that they restore at least 10 km east toward the Colorado Plateau, but no farther than the edge of the plateau itself, giving a minimum westward translation of the Frenchman Mountain block relative to the plateau of 60 km (minimum length of $V_{A2A3} + V_{A3A4}$; Fig. 8, Table 1). The maximum possible translation is 90 km, the current distance from Frenchman Mountain to the plateau.

The azimuth of displacement suggested by matching the southward pinchouts of the Jurassic Aztec sandstone beneath the basal Tertiary unconformity (Bohannon, 1979) is S70°W, consistent with other kinematic studies of extension direction in the region (for example, Anderson, 1971; Angelier and others, 1985; Wernicke and others, 1985; Smith and others, 1987). The pinchout, however, is so gradual that its pre-extension trend is poorly constrained and is valid as a piercing point only if Frenchman Mountain restores directly atop the blocks north of Gold Butte (see analysis of Proffett, 1977, Fig. 11). An uncertainty in the azimuth of displacement of 10° for V_{A2A3} from this direction, however, places Frenchman Mountain too far north to receive the proximal megabreccias on the northern extreme, and too far south to accept crystalline detritus from the south on the southern extreme. We assume the same direction for V_{A3A4}, but assign a 20° azimuthal uncertainty, constrained on the north by the overlap of Phanerozoic strata at A3 with those in the North Virgin Mountains (Fig. 4) and on the south by the improbability that the blocks had a northward component of motion relative to the plateau (Fig. 8, Table 1).

TABLE 1. RELATIVE MOTION VECTORS BETWEEN SELECTED POINTS IN THE BASIN AND RANGE PROVINCE NEAR THE LATITUDE OF LAS VEGAS, NEVADA

	Vector	Displacement (km)	Azimuth
I. Las Vegas system			
A. Lake Mead path	V_{A1A2}	8 ± 8	N84° ± 20°E
	V_{A2A3}	65 ± 15	N70° ± 10°E
	V_{A3A4}	20 ± 10	N70° ± 20°E
B. Mormon Mountains/ Las Vegas Valley path	V_{B1B2}	54 ± 10	N75° ± 10°E
	V_{B1C2}	0 ± 10	N68° ± 0°W
	V_{C1C2}	48 ± 7	N69° ± 3°W
II. Spring Mountains rotation	V_{D4C1}	0 ± 10	N65° ± 0°W
III. Death Valley system	V_{D2D4}	125 ± 7	N65° ± 7°W
	V_{D1D2}	22 ± 3	N45° ± 20°W
	V_{E1E2}	9 ± 6	N60° ± 20°W
	V_{E2E3}	9 ± 1	N35° ± 10°W

Anderson (1973) and Bohannon (1979, 1984) proposed that the motion of Frenchman Mountain relative to the Virgin Mountains was accommodated by left-lateral strike-slip faulting and recognized that it may be kinematically associated with normal faulting (for example, Hamilton and Myers, 1966; Davis and Burchfiel, 1973). The extent to which the large translations are a product of crustal thinning versus strike slip, however, is not clear from the field relations. Although there are clearly large left-lateral strike-slip offsets present in the region (a stratovolcano centered on the fault system is offset about 20 km by the Hamblin Bay fault, Fig. 4; Anderson, 1973), it is not clear to what extent the translation of the Frenchman Mountain block away from the plateau is a product of *crustal* strike slip (deep-seated relative translation without crustal thinning) versus crustal extension. As emphasized by Anderson (1984), plane strain by sets of strike-slip faults may combine with normal faulting that is otherwise not kinematically coordinated with (or even coeval with) the strike-slip faulting to produce crustal extension. Thus, the left-lateral faults in the Lake Mead region may combine with the right-lateral Las Vegas Valley shear zone to accommodate east-west extension and coeval north-south shortening. The ambiguity of how much extension is absorbed by north-south shortening versus crustal extension is a problem throughout in Death Valley region as well. There is clearly a major component of strike slip in the extending system (for example, Wright and others, 1981; Anderson, 1984; see faults in Fig. 4), and we will attempt to quantify its contribution to the over-all strain pattern below.

In order to obtain the displacement of the Spring Mountains block relative to the plateau, the displacement between the Spring Mountains and Frenchman Mountain must be determined (V_{A1A2}; Fig. 8, Table 1). This is not well known as there are no geologic markers between the two that can be used as a basis for reconstruction. The relative stability of the Spring Mountains contrasts with the highly extended fault blocks on Frenchman Mountain (Longwell and others, 1965), and presumably there has been significant separation of the two, as suggested by the presence of Las Vegas Valley between them. The limits on magnitude of relative motion for V_{A1A2} are 0 and 36 km, assuming no extension and complete closure of Las Vegas Valley (that is, no overlap of autochthonous Phanerozoic section), respectively. The uncertainty in azimuth is difficult to evaluate; we chose an uncertainty of ±30° as a conservative limit (Fig. 8, Table 1), which assumes that it is not exceptional to known regional extension directions in either the Death Valley or Las Vegas systems.

Reconstruction via Las Vegas Valley Shear Zone and Mormon Mountains–Tule Springs Hills Area. An alternative to the path between A1 and A4 is combining a cross-section palinspastic reconstruction drawn between the Colorado Plateau and the Meadow Valley Mountains (V_{B1B2}; Axen and Wernicke, 1987, unpub. data) with the offset along the Las Vegas Valley shear zone (V_{C1C2}; for example, Burchfiel, 1965). The Mormon Mountains and Tule Springs Hills are the principal ranges along the transect and have been mapped at scales of 1:12,000 to 1:24,000 along the entire length of the palinspastic reconstruction. On the basis of varying the geometry of faults at depth that are significant to the reconstruction, Axen and Wernicke (1987) determined that there has been 54 ± 10 km of extension (Fig. 8, Table 1). The azimuth of the extension direction is constrained by mesoscopic studies of fault striae and fault mullions in the Beaver Dam Mountains, fault dips and striae in the Tule Springs Hills, and trends of tear faults in the Tule Springs Hills (Smith and others, 1987). Farther west, the orientation of an elongate dome in the highest major detachment in the system (Mormon Peak detachment); the net slip determined on one of the normal faults in the Mormon Mountains; and the bisectrix of a system of small, conjugate normal faults exposed throughout the western half of the Mormon Mountains constrain the extension direction (Wernicke and others, 1985). These indicators yield an extension direction between S60°W and S80°W for V_{B1B2} (Fig. 8, Table 1).

The Meadow Valley Mountains are structurally contiguous with the Las Vegas Range and are thus part of the central stable block (Figs. 4 and 7). Although the block comprises a number of ranges, including the Meadow Valley Mountains and the Las Vegas, Sheep, and Arrow Canyon Ranges (Fig. 4), their bounding faults are steep and discontinuous along strike, and Mesozoic structures within the ranges are not cut by major detachments (Langenheim, 1988). In particular, the Keystone and Gass Peak systems are continuous and maintain their relative spacing from north to south between B1 and C2 (Fig. 8; although note offset of the Keystone system ramp by the Mormon Peak detachment). A small amount of extension (2–5 km), however, is probable. In addition, the block may have rotated about a vertical axis during extension, although rotation of more than 10° in either direction seems unlikely in that it would misalign structural elements of the thrust belt from their regional north-northeast to north-south trends. This is supported by paleomagnetic studies at three sites located 7, 20, and 25 km due north of C2, indicating little rotation of the block following thrusting (Nelson and Jones, 1987). Thus, we assign a value of 5 ± 10 km of motion S65 ± 10°W to V_{B1C2} to account for the extension and possible small rotations of the entire block (Fig. 8, Table 1).

Offset on the Las Vegas Valley shear zone is the last vector needed to complete the closing of the Las Vegas system (V_{C1C2}). A displacement of 48 km best aligns the Gass Peak thrust with the Wheeler Pass thrust, also closely aligning the positions of the Keystone system ramp and its hanging-wall syncline. Displacement of more or less than about 7 km either way results in significant misalignment of these features. The displacement includes oroflexure in the Las Vegas Range area (for example, Burchfiel, 1965), which has been confirmed by paleomagnetic studies (Nelson and Jones, 1987). A northern limit for the azimuth of reconstruction is firmly set by the requirement that the thrust traces do not overlap (C1 cannot restore north of reference point s, Fig. 8). The southern limit is constrained by the maximum amount of north-south compressive strain accommodated in range blocks and in Las Vegas Valley. Although there may be minor north-south shortening in the ranges north of the shear zone, the oroflexural bending in these ranges seems best considered as a response of the brittle crust to deep-seated simple shear on vertical planes (for example, Shawe, 1965). As concluded by Nelson and Jones (1987), a model of block rotation above a smoothly shearing medium below (Jackson and McKenzie, 1983) fits paleomagnetic and structural data well. Along the portion of the shear zone between C1 and C2, a wide alluviated valley is present (Las Vegas Valley, Fig. 4), favoring net transtension over trans-

pression. Minimal north-south distance between exposures of the thrusts prior to motion on the shear zone places C1 no farther south than reference point t in Figure 8; hence, we assign a 3° uncertainty to the azimuth of V_{C1C2} (Fig. 8, Table 1).

Death Valley System

Down-to-the-west normal faulting of the Death Valley normal fault system is superimposed on the Wheeler Pass system and higher thrusts (Fig. 7). By reconstructing the Mesozoic orogen in the Spring Mountains and in ranges to the west, we have established firm correlations between individual thrust faults discontinuously exposed in the range blocks and determined their relative spacing. An important factor in the precision of the reconstruction is the correlation of the Panamint thrust fault, exposed in the Panamint Range (point D2, Fig. 8), with the Chicago Pass thrust, exposed in the Nopah Range (point D3, Fig. 8; Wernicke and others, 1988a, 1988b). Correlations of structurally higher thrusts confirm this, because they tie together the thrust stack in the Tucki Mountain–Cottonwood Mountains area with that in the Funeral and Grapevine Mountains areas (J. K. Snow and B. Wernicke, unpub. data), showing that the entire system, now exposed across an area more than 150 km wide, was initially slightly less than 30 km wide prior to extension (compare scales of Figs. 5 and 6). The principal marker constraining the reconstruction is the White Top Mountain backfold in the Cottonwood Mountains and a correlative fold system in the Funeral Mountains (Figs. 5, 6, and 8; J. K. Snow and B. Wernicke, unpub. data).

Tucki Mountain–Nopah Range Reconstruction. The Panamint and Chicago Pass thrusts exposed in these two areas share the distinguishing features of the Wheeler Pass system described above. In addition, the normal fault blocks of the two ranges reconstruct such that the ranges structurally overlap one another in map view (Fig. 6b; Wernicke and others, 1988b). The geology of both ranges is characterized by steep dips of Tertiary strata that lie with mild angular unconformity on Paleozoic strata (Burchfiel and others, 1983; Wernicke and others, 1986, 1988a). Proximity of the two ranges prior to extension was proposed by Stewart (1970, 1983), based on similarities in stratigraphy of the miogeoclinal clastic wedge between them. Stewart's (1983) reconstruction restores the strong anomaly in isopach trends across the northern Death Valley–Furnace Creek fault zone (NFZ, Fig. 4), indicating about 80 km of displacement N55°W of the Panamint Range block relative to the Nopah–Resting Springs Range block. The reconstruction was questioned by Prave and Wright (1986), who argued that the isopach trends could be reasonably restored with only 50 km of displacement.

Identification of the Panamint thrust at Tucki Mountain confirms Stewart's (1983) placement of the Panamint Range adjacent to the Nopah–Resting Springs Range block because it provides a structural marker that can be used to precisely determine the relative offset. As discussed above, the Panamint and Chicago Pass thrusts have about 5,000 m of stratigraphic throw, cut steeply (40°–60°) across miogeoclinal layering, and occur in identical positions in the miogeocline. In addition, the structural details of the exposure of the two thrusts permit cross-section reconstruction of the two range blocks *directly adjacent* to one another without holes or overlap (Fig. 6b). In the Nopah Range, the thrust places lowest Cambrian on Devonian-Mississippian strata (at point D3), whereas at Tucki Mountain, the thrust places Middle Cambrian strata on Permian (at point D2). In both blocks, now tilted owing to extension, a normal fault system dips slightly more shallowly than the thrust, such that it cuts downward to the west from the footwall into the hanging wall of the thrust, moving the footwall westward over the hanging wall. In both ranges, the normal fault crosses the thrust at very shallow angle where the thrust emplaces high Lower or low Middle Cambrian on Pennsylvanian or Permian strata (at points D3 and D2, respectively). Figure 6b shows that the two ranges fit directly against one another, restoring into a crustal sliver only 2–3 km wide that contains the trace of the thrust.

If it is assumed that the thrusts do correlate and require juxtaposition of the two ranges, they do not specify azimuthal control on V_{D2D3}. We suspect that points D2 and D3 fit directly against one another; otherwise, the structural details of the normal fault system relative to the thrust in the two ranges would have to persist for significant distances along strike. As we show below, however, an azimuth similar to the one suggested by Stewart (1983) is indicated from independent strain-compatibility arguments.

Closing Pahrump Valley. The placement of the combined Nopah–Resting Springs–Panamint crustal sliver in its position with respect to the Spring Mountains (V_{D2D4}, Table 1) can be done with precision by considering the trace of the Wheeler Pass relative to other thrusts in the Spring Mountains block. When followed from north to south, (1) the thrusts curve from northeast strikes to due north or north-northwest, (2) individual thrust plates carry progressively thinner sections of the miogeoclinal prism, and (3) the relative spacing between major thrusts decreases. At large scale, the Death Valley system breakaway fault zone makes a concave-west scoop across the west side of the Spring Mountains such that the Wheeler Pass system projects beneath the alluvium of Pahrump Valley for a distance of 60 km and reappears in the Clark Mountains area as the Winters Pass thrust (Figs. 4, 7, and 8; Burchfiel and Davis, 1971, 1981). In the Clark Mountains area, spacing between the three major thrusts, the Keystone, Mesquite Pass, and Wheeler Pass thrusts, is only about a kilometer or two, whereas to the north in the Spring Mountains, the spacing is about 10–15 km (Figs. 6 and 8). Thus, a geologic line associated with the Wheeler Pass thrust (for example, the intersection of the thrust plane with footwall Mississippian strata) projected southward into Pahrump Valley would lie within 30 km of the trace of the Keystone system and within 10 km of the Lee Canyon thrust, the major thrust between the Keystone and Wheeler Pass (Figs. 6 and 8). The choice of geologic line is not critical, as the thrust cuts steeply through the miogeoclinal section where exposed, and thus, any fault-bed intersection originally lay within a few kilometers of any other along the thrust. These arguments suggest that D2 and D3 restore to a position at least as far east as that line, *regardless of their initial position relative to one another* (Fig. 8).

The Panamint Range block, although tilted and extended, is a relatively coherent homocline of miogeoclinal strata and underlying basement that contains the trace of the thrust. It has experienced little, if any, north-south internal strain (for example, Wernicke and others, 1988a; Albee and others, 1981). At the southern end of the block, exposures of upper Paleozoic and Mesozoic strata in Butte Valley (Fig. 4) may represent the footwall of the thrust, as they are juxtaposed against basement and Pahrump strata along the steeply dipping Butte Valley fault zone (Johnson, 1957). We favor the interpretation that the Butte Valley fault juxtaposes the hanging wall of the Wheeler Pass system with its footwall (Figs. 5 and 7), such that the Precambrian is downthrown along the fault relative to the younger rocks. If so, then the Panamint Range homocline is everywhere within a kilometer or two of the thrust plane, tightly constraining the position of the Panamints with respect to the Spring Mountains block, in light of the western limit on the original position of the Wheeler Pass system in Pahrump Valley discussed above. Even if the upper Paleozoic and Mesozoic strata in Butte Valley are not part of the footwall of the thrust, the shallow dip of the thrust where exposed at Tucki Mountain (Wernicke and others, 1988b) suggests that it cannot have strayed too far beneath the Panamint homocline south of Tucki Mountain.

An independent argument supporting correlation of the Panamint thrust with the Wheeler Pass system is the apparent structural continuity of rocks in the Panamints with those exposed in the hanging wall of the Winters Pass thrust in the Clark Mountains area. As indicated by Burchfiel and others (1983), the region between the Nopah–Resting Springs Range

block and the Panamints is devoid of exposures of older-over-younger faults. It is composed of a number of steeply east-dipping, north-striking homoclines repeated on numerous low-angle normal faults (Wright and Troxel, 1973, 1984). These blocks are apparently fragments of a once-contiguous homocline exposed from the Winters Pass thrust to the Panamints, including the southern Nopah Range, the Kingston Range, the southern Black Mountains, and other smaller blocks (Fig. 4). Northward pinchouts of the Pahrump beneath Precambrian Z strata are preserved within each of the blocks and fall on a single west-northwest–trending line between the Kingston and southern Panamint Ranges (Wright and others, 1974), and the basal Tertiary unconformity in most places rests on Lower or Middle Cambrian strata, which throughout the area are fairly uniform in thickness (for example, Stewart, 1970). These relations indicate that it is unlikely that a thrust fault with 5 km of stratigraphic throw disrupts the blocks between the Winters Pass area and the Panamints. All of the blocks likely belong to the same Mesozoic thrust plate, and their lower bounds (Chicago Pass, Winters Pass, and Panamint thrusts) are thus parts of the same thrust system (Fig. 7).

Azimuthal limits on V_{D2D4} are prescribed on the northern extreme by the condition that Cambrian strata intersecting the thrust plane at Tucki Mountain (reference point v, Fig. 8) do not overlap those at the southern limit of exposure of the Wheeler Pass thrust (reference point u, Fig. 8). On the south, a limit is set by the presence of a roughly east-west–trending boundary between both hanging-wall and footwall strata intruded by Mesozoic plutons to the south and a pluton-free area to the north (Fig. 8). If the upper Paleozoic and Mesozoic strata in Butte Valley are not in the footwall of the thrust system, the reconstruction is less tightly constrained, but not seriously compromised. The Panamint block must restore to a position such that unintruded hanging-wall rocks do not overlap intruded hanging wall in the Clark Mountains area. The precise location of the thrust plane in the Panamints is not critical to this constraint, as in both areas, the northern limit of the plutonic belt is laterally persistent for tens of kilometers. In Figure 8, this constraint means that point w in the Butte Valley area cannot restore to a position south of point x in the Clark Mountains area. The possibility exists, however, that there has been north-south shortening in the southern part of the Spring Mountains block, in which case the southern limit of the plutonic belt would shift farther south. The shortening may be accommodated by as much as 20 km of right-lateral movement along the State Line fault (Fig. 8; Hewitt, 1956), although the geology on both sides of the fault does not require major displacement. Because significant displacement is possible, however, we consider a reference point y 10 km to the south of x as a southern limit for the azimuth of the reconstruction. These constraints give an azimuth of N65 ± 7°W for V_{D2D4}. This corresponds closely with the azimuth inferred for restoration of Tucki Mountain and Chicago Pass discussed above, but is based on independent constraints.

Distance limits on V_{D2D4} are given by the need to restore D2 at least as far east as the projection of the Wheeler Pass system into Pahrump Valley, but not so far east that it would overlap the interpolated trajectory of the next-lower thrusts (Lee Canyon and Green Monster thrusts) into Pahrump Valley (Fig. 8). These limits place D2 at D4, 125 ± 7 km S65° ± 7°E of its present location, but with the distance uncertainty skewed such that the minimum follows the western limit for the position of the Wheeler Pass system (Fig. 8).

The final vector needed to close Pahrump Valley is that for the restoration of the Cottonwood Mountains relative to Tucki Mountain (V_{D1D2}, Table 1). Palinspastic reconstruction of the Cottonwood Mountains relative to D2 on Tucki Mountain requires restoration of the Emigrant fault system on the east side of the Panamint block. Detailed mapping and structural analysis of this area (see Wernicke and others, 1986, 1988b; Hodges and others, 1987) show that 20–25 km of extension has occurred between these points, oriented N45° ± 20°W, assuming liberal uncertainty in the extension direction from several hundred measurements of fault striae and mylonitic stretching lineations in the extended blocks at Tucki Mountain (Walker and others, 1986).

Cottonwood Mountains to the Sierra Nevada. Extension between the Cottonwood Mountains and the Sierra Nevada is modest and is best constrained by closing the northern part of Panamint Valley along the Hunter Mountain fault (Burchfiel and others, 1987). Piercing points across this structure indicate 9 ± 1 km of motion oriented N55° ± 10°W (Burchfiel and others, 1987) for V_{E2E3} (Table 1). The area between D1 and E3 is occupied by the Hunter Mountain batholith terrain and does not appear to be highly extended or rotated about a vertical axis.

The last vector to be considered is one that connects the northern Argus–southern Inyo Range area with the Sierra Nevada, which takes into account extension related to the opening of Owens Valley (V_{E1E2}, Table 1). There are no major detachments or major normal faults other than those bordering Owens Valley between E1 and E2 (see Dunne, 1986, for a review of extensional structures in this region). The geometry of faulting at depth in Owens Valley is not known but is probably of the type that is steep and fairly deeply penetrating (for example, Anderson and others, 1983). Two major known faults, the Owens Valley and Independence faults (Fig. 4), are steeply dipping at the surface and show evidence of oblique slip, being primarily right-lateral strike slip for the Owens Valley fault and dip slip for the Independence fault (Zoback and Beanland, 1986; Gillespie, 1982). We assume that the region is generally pervaded by high-angle faults with modest offset; an estimate of 15% ± 10% extension reasonably bounds the extension, giving a displacement of 9 ± 6 km. An over-all extension direction of N60° ± 20°W, parallel to the geodetically determined direction (Savage, 1983; see also review of pertinent data in Jones, 1987), bounds the azimuth.

Rotation of the Spring Mountains Block. The two vector paths within the Las Vegas system and the one in the Death Valley system have different endpoints in the Spring Mountains block that are separated by about 50 km. Thus, although the Spring Mountains block is negligibly extended internally (for example, Burchfiel and others, 1974; Axen, 1984), rotation about a vertical axis of the entire range block could significantly affect the relative position of two widely separated points. In particular, rotation of segment C1D4 introduces significant error into determination of east-west extension. The amount of rotation has yet to be constrained paleomagnetically, but the over-all north to north-northeast trend of the thrust faults in the area suggests that major rotation (in excess of 10°) has probably not occurred. This uncertainty will be considered in the discussion of putting the other vectors together into a whole-province reconstruction.

DISCUSSION

Vector Addition

A whole-province reconstruction may be obtained by adding the displacements of the Las Vegas system to those of the Death Valley system. Because we have two paths in the Las Vegas system based on independent constraints, we can narrow the uncertainty limits on the Las Vegas system reconstruction to that region of uncertainty common to both paths. We can then define a new vector for the Las Vegas system and add it to the Death Valley system to obtain a whole-province reconstruction, taking into account possible rotation about a vertical axis of the Spring Mountains block.

In adding vectors with uncertainties, we assume constant probability distribution within each uncertainty domain. We generate the combined uncertainty region by sweeping one uncertainty region around the other, placing the "best-fit" vector of one on the perimeter of the other. The new area is that in which any combination of the summed vectors may lie. This

area is a conservative estimate of the uncertainty, as two randomly chosen vectors from the original uncertainty fields are less likely to sum to a point on the perimeter of the combined uncertainty field than to a point near the center (see Monte Carlo simulation, below). The uncertainty is also considered conservative to the extent that the probability distribution within each of the uncertainty regions to be summed is not everywhere equal, but generally concentrated in the center near the "best-fit" vector. Because it is difficult to quantify the probability distribution for each vector, we have chosen an even distribution in order to provide an upper limit on the uncertainty.

Adding the vector paths between the Spring Mountains and the Colorado Plateau shows that the uncertainty for the Lake Mead path is considerably larger than that for the Mormon Mountains area path (Figs 9a and 9b, respectively); however, there is a relatively small area of overlap between the two (Fig. 10). The Lake Mead path suggests a more northerly over-all extension direction; most of the uncertainty lies in the positioning of the Frenchman Mountain block (Fig. 8). The more easterly trend of the Mormon Mountains path is due primarily to the southeasterly motion on the Las Vegas Valley shear zone. The two paths are thus best reconciled by southeasterly motion between the Frenchman Mountain block and the Spring Mountains, parallel to the Las Vegas shear zone. If

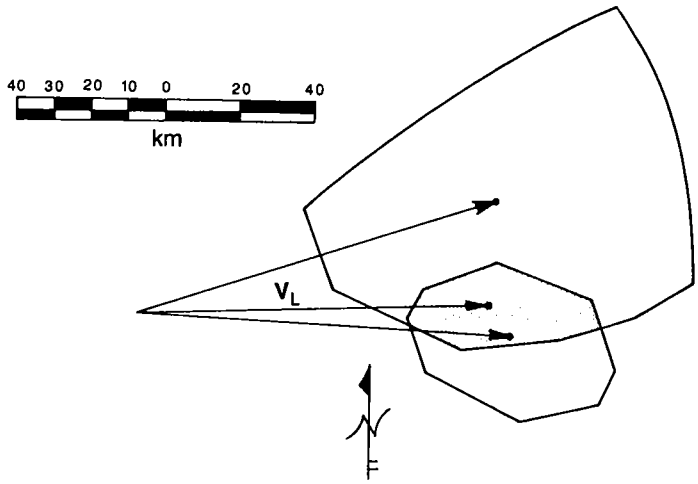

Figure 10. Derivation of Las Vegas system vector V_L using overlap of combined uncertainty fields of two independent paths (Figs. 9a and 9b).

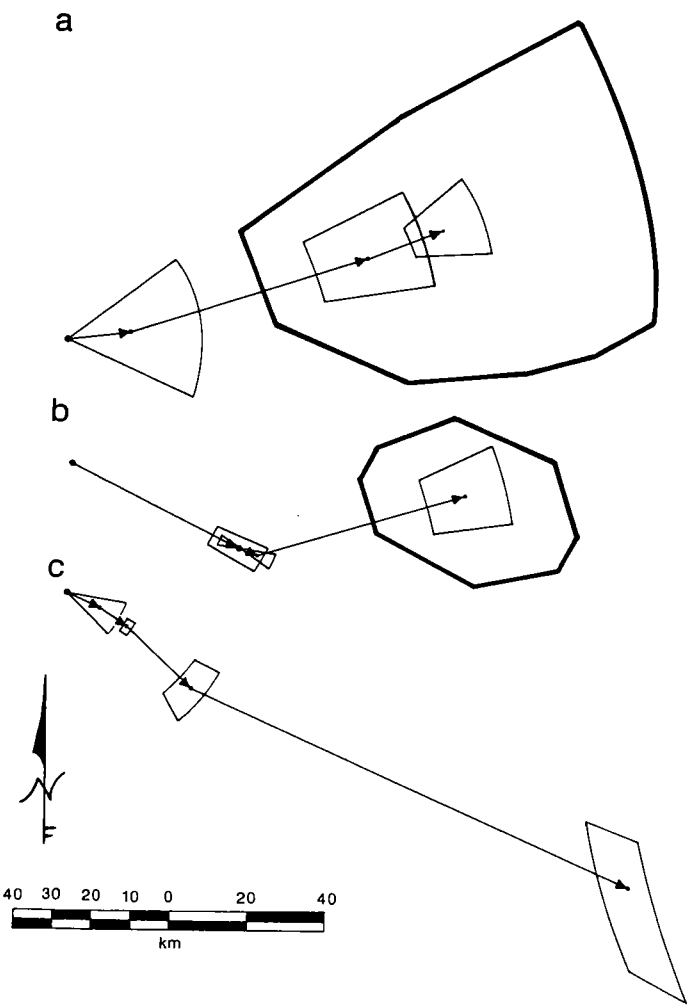

Figure 9. Summation of vectors in reconstruction paths. a. Las Vegas normal fault system via Lake Mead area path; b. Las Vegas system via Mormon Mountains–Las Vegas Valley area path; c. Death Valley system. Bold outlines in a and b show limits of combined uncertainties. See text for discussion.

so, it appears likely that a strand of the shear zone passes to the south of Frenchman Mountain (for example, Anderson, 1973). The area of overlap defines a new vector V_L with an uncertainty region that comprises all the combinations of vectors from the two paths that are consistent between the two sets. We chose a "best fit" that lies in the center of this combined uncertainty field (Fig. 10).

Combining the vector derived from the two paths in the Las Vegas system with those in the Death Valley system must account for possible rotation of the Spring Mountains block about a vertical axis, because the paths for the Las Vegas and Death Valley systems have distant endpoints within the Spring Mountains (Fig. 8). Although such a rotation of the whole block is probably small, a significant differential displacement of points D4 and C1 with respect to the plateau is possible. To account for as much as 10° of rotation of either sense, we let $V_{C1D4} = 0 \pm 10$ km N65°W (Fig. 8, Table 1).

The total displacement of the Sierra Nevada relative to the Colorado Plateau V_t is

$$V_t = V_{E1E2} + V_{E2E3} + V_{D1D2} + V_{D2D4} + V_{D4C1} + V_L \qquad (1)$$

We depict the result of this addition in the form of an error cloud produced by a Monte Carlo simulation of 10,000 runs of equation 1, where vectors were chosen randomly from the uncertainty regions of each vector (Fig. 11). A curve parallel to the density distribution of points which excludes 5% of them is taken as an estimate of the uncertainty in the province-wide reconstruction. From this, we obtain $V_t = 247 \pm 56$ km S73° ± 12°E, using the "best-fit" vectors and considering the extremes of the error curve in Figure 11.

Strain Rate

The timing of extensional tectonism in the region is constrained to have occurred principally between 20 m.y. ago and the present. Deposition of orogenic conglomerates in the oldest preserved Tertiary strata began in Oligocene time (for example, Reynolds, 1974), but major extension, as indicated by angular unconformities within the Tertiary section and the depositional overlap of extensional structures, appears not to have begun until after 20 m.y. ago (Bohannon, 1984; Anderson and others, 1972; Cemen and others, 1985; Wright and others, 1983). On the basis of the age of strata cut by major extensional features, the bulk of the exten-

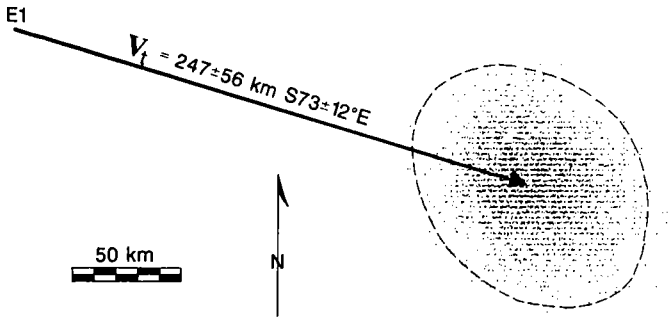

Figure 11. Neogene displacement vector and uncertainty region of point E1 (Fig. 8) with respect to the Colorado Plateau. See text for discussion.

sion appears to be post–15 m.y. ago. The peak period of extension on the Las Vegas system occurred between 15 and 11 m.y. ago (Anderson and others, 1972; Bohannon, 1984; Smith and others, 1987). In the Death Valley system, most of the deformation in the eastern part of the path occurred between about 14 and 4 m.y. ago (Cemen and others, 1985), but in western areas, tens of kilometers of displacement have probably occurred since 4 m.y. ago (for example, Wernicke and others, 1986; Burchfiel and others, 1987; Butler and others, 1988), including motion on major low-angle normal faults. If it is assumed that most of the translation of the Sierra Nevada away from the plateau occurred after 15 m.y. ago, the average displacement rate for the past 15 m.y. is 16.7 ± 4.5 mm/yr.

Given that the province has at least doubled in width (lower limit of displacement of about 190 km compared with a current width of 360 km) and may have experienced a sixfold increase in width (about 300 km of extension or 500% increase over original width), the time-averaged strain rate of the lithosphere as a whole is in the range 2.1×10^{-15} to 1.9×10^{-14} s^{-1}. The lower bound of our average displacement rate is greater than the upper bound of Minster and Jordan's (1987) Holocene opening rate of 9.7 ± 2.1 mm/yr derived from considerations of geodetic data, the RM2 plate model, and strain west of the San Andreas fault. The azimuth of opening from this study and that derived by Minster and Jordan (1984, 1987) are similar. Combined, these data indicate that Basin and Range extension has slowed significantly over the past 15 m.y. Such slowing would require faster displacement rates during earlier parts of the Neogene. For example, if the opening rate of the Basin and Range has been 8 mm/yr for the past 5 m.y., then the average rate between 5 and 15 m.y. ago would have been 21 ± 7 mm/yr.

The concept of slowing of Basin and Range extension with time has been hypothesized based on the observation that widely spaced, steep normal faults commonly overprint younger, more closely spaced normal-fault systems (for example, Zoback and others, 1981), but measurements supporting such a hypothesis have heretofore been lacking. The difference in timing between extension in the Las Vegas system and the Death Valley system places bounds on how the displacement rate of the Sierra relative to the plateau varied in time. Figure 8 shows points on V_{D2D4} that constrain the timing of movement of the Panamint Range block northwestward relative to the Spring Mountains block. Tertiary overlap of major extensional structures occurred by 9 m.y. ago in the Resting Springs Range area and by 4 m.y. ago in the Furnace Creek Wash area (Fig. 8; Wright and others, 1983, 1984; Cemen and others, 1985; McAllister, 1973). Thus, of the 125 km of motion represented by D2D4, at least 50 km had occurred by 10 m.y. ago, at least 90 km had occurred by 5 m.y. ago, and no more than about 35 km occurred between 5 m.y. ago and the present. If it is assumed that all motion on vectors west of D2 occurred in the past 5 m.y., the Death Valley system has accommodated no more than 50 km of extension in the past 5 m.y., giving a rate of 10 mm/yr, in good agreement with the Holocene rate of Minster and Jordan (1987). Because it is clear that extension in the Las Vegas system was complete by 5 m.y. ago (although it was probably mostly complete by 10 m.y. ago; Anderson and others, 1972; Bohannon, 1984), slowing of extension with time is required.

The most likely displacement history, neglecting the effects of locally variable extension direction, includes about 150 km of extension accommodated on both systems between 10 and 15 m.y. ago (100 km on the Las Vegas system, 50 km on the Death Valley system, for a total of 30 mm/yr) and an additional 100 km accommodated on the Death Valley system in the past 10 m.y. (10 mm/yr), with over-all slowing occurring between 5 and 10 m.y. ago (Fig. 12, curve 1). Alternatively, if extension in the Las Vegas system were evenly distributed across the time interval 15–5 m.y. ago, then the displacement rate would be 20 mm/yr for that interval, slowing to 10 mm/yr for the past 5 m.y. (Fig. 12, curve 2). The actual displacement rate probably slowed from a value in excess of 20 mm/yr to one near 10 mm/yr over the interval 5–15 m.y. ago (Fig. 12, curve 3). Timing data are not yet precise enough to meaningfully bound displacement rates at greater precision than over 5-m.y. intervals.

Although the slowing documented in this study is in accord with that previously suggested on the basis of changing structural style with time in the Basin and Range, we stress that our results are independent of assumptions of structural style. In fact, we note that the Holocene opening rate of 9.7 ± 2.1 mm/yr (Minster and Jordan, 1987; Fig. 12) must be accommodated principally in the western part of the Death Valley system at the latitude of Las Vegas (Fig. 7), which has a structural style similar to that of earlier extensional regimes (Wernicke, 1981; Wernicke and others, 1986; Hamilton, 1987; Burchfiel and others, 1987). According to the analysis above, neither the Death Valley system nor the Las Vegas system individually needs to have spread at a rate in excess of 10 mm/yr at any time in their histories. Thus, our results do not necessarily support the concept that slowing of Basin and Range extension is associated with a change in structural style or that such change, if any, constrains the displacement rate.

The magnitudes of strain rate and the changes of strain rate with time are all broadly consistent with the physical model presented by Sonder and others (1987) and Wernicke and others (1987), suggesting that extension is controlled by the gravitational collapse of crust overthickened during Mesozoic time (for example, Coney and Harms, 1984). According to the calculations of Sonder and others (1987), the peak magnitude of strain rate

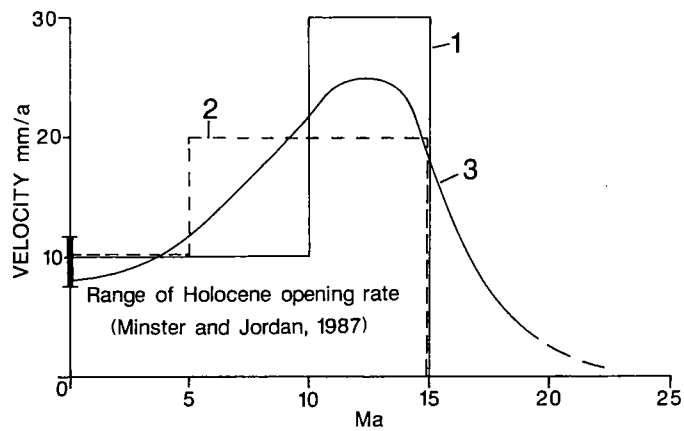

Figure 12. Opening rates of Las Vegas area Basin and Range averaged over 5-m.y. intervals. Curve 1, assuming Las Vegas system ceased moving by 10 m.y. ago; curve 2, assuming Las Vegas system ceased moving by 5 m.y. ago; curve 3, possible smoothed path. See text for discussion.

is approximately $1.0–3.0 \times 10^{-15}$ s^{-1} with a time scale of slowing on the order of a factor of 2–3 per 10 m.y. following the peak of extensional strain rate. The calculations therefore broadly agree with these observations, but of course do not rule out forces other than gravitational ones for the origin of extension.

An important difference between the results of Sonder and others (1987) and of this report is the magnitude of strain. Figure 5 shows that the total shortening across the thrust belt is about 100 km or roughly a factor of 2. Given an initially cold Moho temperature (<600 °C) for the Las Vegas region (Sonder and others, 1987) and shortening on the order of a factor of 2 during thrusting, the gravitational-collapse model is difficult to reconcile with the "best-fit" value of $\beta = 3.5$, although it is consistent with the lower bound (extension factor $\beta > 2$). The results of Sonder and others (1987) show that in general, it is possible to have a greater magnitude of extension than of compression because of the excess potential added to the lithosphere by upward advection of heat during extension. For a broad range of assumptions as to how such advection occurs and the mechanical properties of the lithosphere, however, the extension in the Las Vegas area, if significantly greater than the lower bound of $\beta = 2$, may not be entirely explained by the gravitational collapse mechanism. As we discuss below, two factors may contribute to the possible discrepancy between our measurements and the gravitational collapse model: (1) a large component of constrictional strain in the evolution of the Las Vegas area Basin and Range and (2) the availability of a driving force for extension other than gravitational collapse, possibly the tangential shear traction exerted on the west margin of the North American plate by the Pacific plate in Neogene time.

Constrictional Strain Component

A substantial percentage of the east-west extensional strain in the region may be absorbed by north-south crustal shortening rather than crustal thinning, resulting in an over-all constriction of the crust during extension. The effect of constriction on the gravitational collapse model, which is one dimensional and assumes that all crustal extension is accommodated by plane-strain crustal thinning, is to allow the crust to extend more than the driving force of gravity alone might permit. For example, if 30% of the total extensional strain is balanced by north-south shortening, then sufficient energy may be available from gravitational collapse to account for the remaining component of extension. Such a solution to the problem is not entirely satisfactory, however, because it does not specify the driving force for the constriction itself. A detailed assessment of this problem is beyond the scope of this paper, but we can make some statements about the likely importance of constriction in the region and propose a model for its origin.

As pointed out by Wright (1976), the presence of numerous, large strike-slip faults in the Las Vegas area relative to other parts of the Basin and Range may indicate that the region has been extended more than have areas to the north and south, responding to the difference in extension by accommodating much of it along conjugate strike-slip zones in addition to

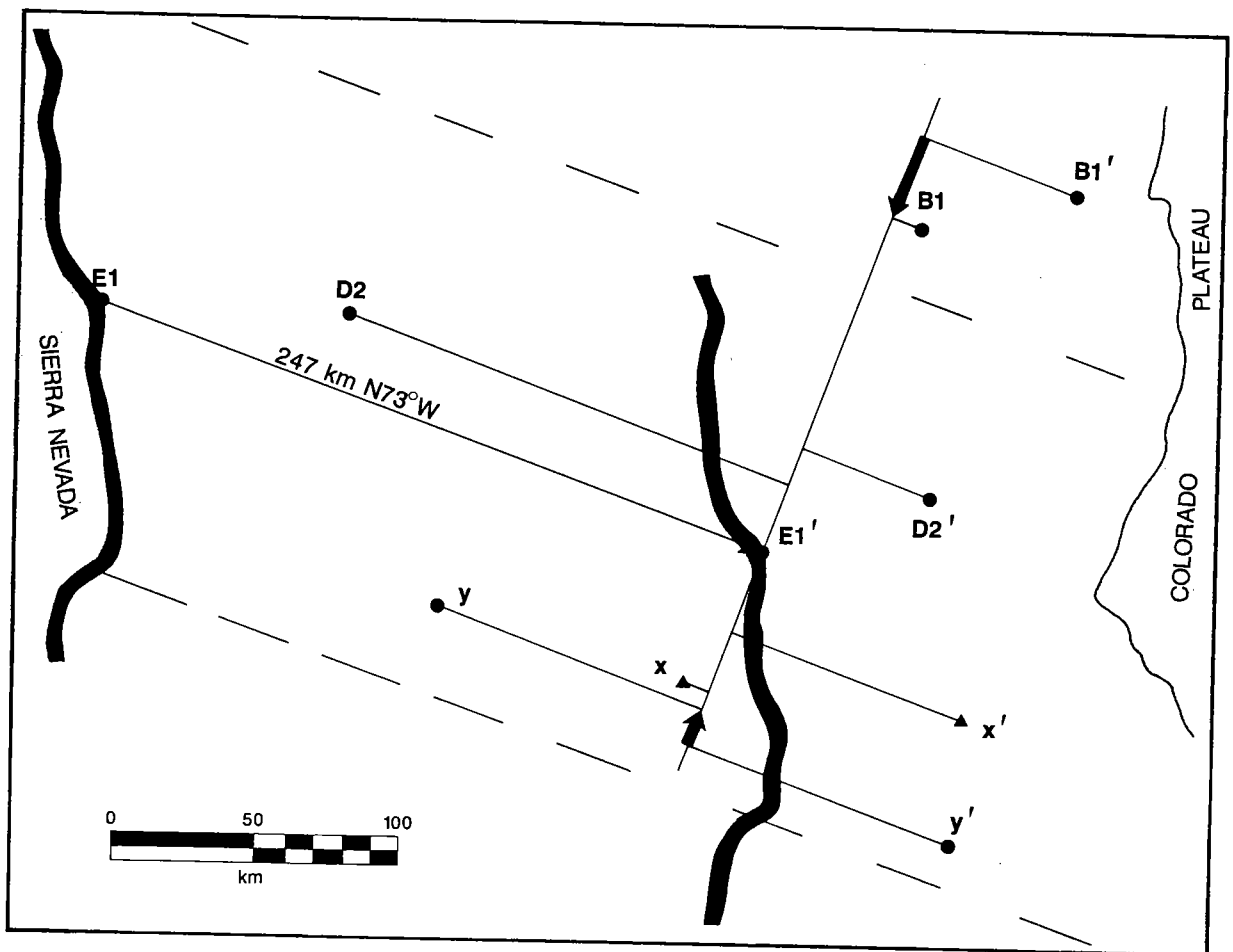

Figure 13. Map showing selected points from Figure 8 (unprimed) restored to their pre-extension configurations (primed) relative to the Colorado Plateau, using "best-fit" vector V_t from Figure 11. See text for discussion.

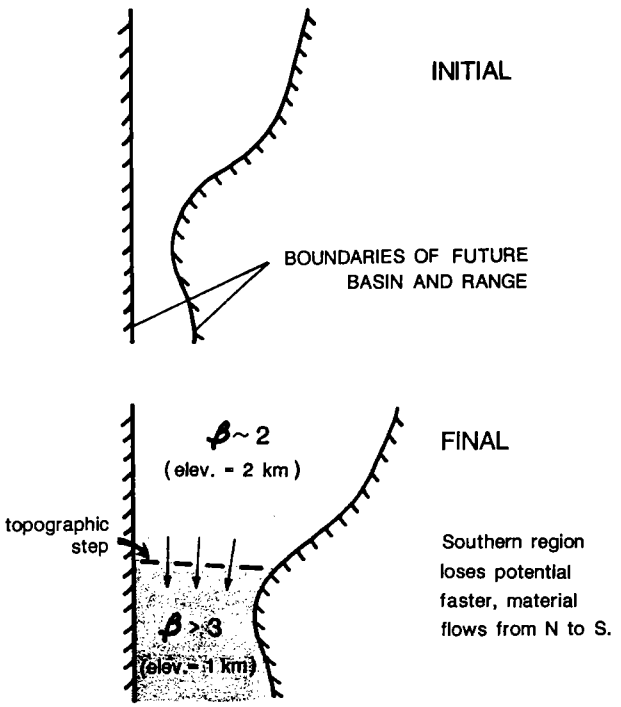

Figure 14. Model for origin of constrictional strain in the Las Vegas area Basin and Range. See text for discussion.

normal faulting. Hamilton and Myers (1966), Davis and Burchfiel (1973), and Lawrence (1976) viewed strike-slip faulting at the northern and southern extremes of the Great Basin as large tear or transform faults bounding regions of relatively large extension from those of little or no extension. The major strike-slip faults in the Great Basin region characteristically have apparent left-lateral offset where northeast striking and right-lateral offset where northwest striking (Shawe, 1965). This is particularly clear in the Las Vegas region, where the Lake Mead and Garlock fault systems are left lateral and the Las Vegas Valley shear zone and Death Valley fault zone and several other northwest-striking faults are right lateral (Fig. 4). Shawe's (1965) theory of Basin and Range extension ascribes a fundamental role to deep-seated conjugate strike-slip faults in the development of Basin and Range structure. Such an origin implies a predominantly map-view plane-strain pattern in which the crust need not thin appreciably in the accommodation of extension, thus representing an end-member case in which conjugate shears accommodate most or all of the extension. At the other extreme, the "intracontinental transform" (Davis and Burchfiel, 1973) or "transfer" (Gibbs, 1984) fault model of these strike-slip faults implies that none of the strike slip be attributed to north-south shortening. These end members are herein referred to as "transfer" and "conjugate" faulting.

The relative importance of conjugate faulting can be assessed by determining the component of north-south motion of crustal blocks relative to the net extension direction of N73° ± 12°W. Consider two points B1 and y in the extending system (Figs. 8 and 13), which initially lie at the northern and southern extremes of the region characterized by strike-slip faulting. Point y lies in the northern Mojave Desert and thus accounts for approximately 65 km of left slip on the Garlock fault in the reconstruction (vector not listed in Table 1). The initial positions of selected points from Figure 8 are shown in Figure 13 as primed points. Measured relative to the net extension direction, points B1 and y converge 43 km, using the "best-fit" vectors. Because of the uncertainty in the net extension direction and in the individual displacements, there is a great deal of uncertainty in this estimate. In addition, if extension directions were different at different times or in different areas (for example, Zoback and others, 1981), it is possible to produce apparent shortening perpendicular to the net extension direction in the absence of true conjugate faulting. For example, a combination of southwest extension of points B1 and y followed by northwest extension involving only y would produce apparent convergence of B1 and y without conjugate faulting or shortening perpendicular to the net extension direction. Considering these difficulties, we can estimate, but not reasonably bound, the constrictional component. It does not seem likely, however, that the shortening is substantially more than 43 km, which represents about a 20% decrease in original north-south distance between B1 and y. Thus, the component of crustal extension accommodated by horizontal shortening may be significant but is probably relatively small. For an extension factor $\beta = 3.3$ and north-south shortening factor of 0.8, the crustal thinning factor is 0.38, as opposed to a thinning factor of 0.30 in the absence of horizontal shortening.

The modest amount of constriction is probably related to the narrowness of the province at the latitude of Las Vegas (for example, Wright, 1976). If we consider the motion of two rigid blocks relative to one another with unequal widths of material accommodating the strain between them, crustal thinning will be greater in the narrow areas than in the wide ones (Fig. 14). The thin-sheet calculations of Sonder and others (1987) indicate that significant differential strain between two regions (say, thinning by a factor $\beta = 3.5$ versus $\beta = 2$) may lead to a significant gradient in gravitational potential of the lithosphere between them. We suggest that such a gradient may have led to flow of material parallel to the gradient (Fig. 4). Such a contrast in buoyancy of the lithosphere may still be affecting the region, where a strong north-south gradient in regional topography is present all across the province (for example, Eaton and others, 1978) and north-south shortening is active, as in the Death Valley region (Fig. 7). In addition, the Intermountain Seismic Belt (for example, Smith, 1978) is developed along the topographic discontinuity and is perhaps driven by the current buoyancy contrast. Such a contrast is not present to the south of the region, but the contrast in width of the province to the south is negligible compared with that to the north.

The mode of strain accommodation for the constrictional component seems to be similar to that proposed by Hill (1982), whereby a system of conjugate faults bounds rigid, crudely hexagonal blocks. In this model, the strike-slip faults are both conjugate and transform, because they transfer extensional displacement from one zone of pull-apart to another, while simultaneously accommodating extensional strain in a manner similar to that proposed by Shawe (1965). The join between the Las Vegas Valley shear zone and the Hamblin Bay fault (Fig. 4) may represent the southern apex of a relatively stable block that moved southward into the region of severe pull-apart between the Spring Mountains and the Colorado Plateau, as predicted by Hill's (1982) block model (see also Anderson, 1984).

CONCLUSIONS

We conclude that the Sierra Nevada moved 247 ± 56 km N73° ± 12°W relative to the Colorado Plateau in Neogene time. The rate of motion appears to be on average more easterly and of greater magnitude (20–30 mm/yr) in the early phases of extension between 10 and 15 m.y. ago, slowing to its current rate of less than 10 mm/yr. Although the implications of these measurements for the Cenozoic tectonics of western North America and for processes of extension in general are significant, we emphasize their contribution to the model of Atwater (1970), who proposed that Basin and Range extension was in part the result of diffuse extensional shearing of the continent in response to the growing right-lateral San Andreas transform. Since her study, global plate reconstructions and improved knowledge of the San Andreas fault have confirmed that Pacific–North America relative plate motion during the past 20 m.y.

exceeds the offset on the San Andreas fault by more than a factor of 2 (for example, Stock and Molnar, 1988).

Resolved parallel to the northern San Andreas, our reconstruction adds at least 166 km and as much as 262 km of right-lateral motion along the plate boundary at the latitude of the central Colorado Plateau, coeval with growth of the transform plate boundary. Within the uncertainties of all the data bearing on the probem, the extension magnitude proposed herein supports the hypothesis that all of the right-lateral displacement between the growing San Andreas transform and North America was accommodated as relatively diffuse deformation within the North American plate. Much work remains to be done, however, on reconciling the details and timing of the deformation with plate-tectonic constraints. For example, the west-southwest direction of extension within portions of the Las Vegas normal fault system may be the last phases of extension prior to a province-wide reorientation of extension direction to west-northwest at about 10 m.y. ago, thought to signal the influence of the San Andreas transform on Basin and Range extension (for example, Zoback and others, 1981). Major extension (probably on the order of 100 km or more) in the northern Great Basin region occurred in Eocene to mid-Miocene time (for example, Coney and Harms, 1984; Wernicke and others, 1987), and as such, the 15- to 10-m.y.-ago extension in the Las Vegas region may have an origin more closely related to pure gravitational collapse than to forces applied to the edge of North America by the San Andreas transform system. Continuum models of the San Andreas transform as a finite-width deformation zone attached firmly to the edge of North America have yielded promising comparisons with the actual zone of plate boundary deformation (Sonder and others, 1986).

Our results are one of a long series of attempts to constrain timing and magnitude of Basin and Range extension. We feel that the strength of our approach is to introduce quantitative rigor to the problem, including the major challenge of developing criteria for bounding uncertainties in geologic reconstructions. Although construction of balanced and restored cross sections yields possible strain fields (and spares us the embarassment of proposing an impossible one), we have not yet developed a general set of techniques for determining all possible strain fields from a given body of geologic data. This study represents a crude attempt at doing so, but there is considerable room for debating, point by point, our methods for determining uncertainties, and we hope that from such debate a set of more general principles on how to handle these problems might emerge. Because the principal basis for assigning uncertainties has been the simple compatibility condition that no two points occupy the same place upon reconstruction, we think that we have not grossly underestimated the uncertainties. New geologic, geochronologic, and paleomagnetic data may expand, but one hopes, mostly contract, the uncertainty regions presented in Figure 8 and Table 1. In addition, new data to the north and south may provide more independent paths across all or parts of the province that will serve to reduce uncertainties. We envision an advanced stage of research in the Basin and Range where individual displacement vectors form a tight grid over the province, ultimately allowing inverse modeling of the strain field to obtain the stress field.

Our results fully confirm earlier suggestions that extension in the Basin and Range province is quite large. Hamilton and Myers (1966) proposed that the province may have doubled in width, based in part on the possibility that range-bounding faults may flatten with depth. Since then, a major geologic revolution has fundamentally changed how geologists view extension of the Earth's lithosphere, based principally on Basin and Range field studies. The lower limit of our measurement of extension (which is likely too small) exceeds those made by a handful of workers in the 1960s and 1970s (with the exception of Hamilton and Myers), who at that time were regarded by most other workers as the liberal fringe.

The magnitude of the shift in thinking shows the importance not only of these early field studies on the problem, but also of field-oriented geologic research in general. The revolution in thinking about how the crust extends is likely only a minor sampling of what geologic field relations have yet to teach us about the continental lithosphere. We are probably in a period of development of research into the nature of the continental lithosphere where we know a great deal about large areas of the continents in reconnaissance but have yet to ask the right questions of the rocks to shed proper light on major processes; this is certainly the feeling one gets from comparing the voluminous pre-Anderson literature on the Basin and Range with that of today. Far from the nineteenth-century descriptive science that field geology is perceived to be by the growing number of earth scientists comfortably insulated from terrestrial reality by machines and elegant calculations, it seems to us that major advances in understanding the continental lithosphere will be unlikely in the absence of well-posed, field-oriented research.

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