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Chapter 12

Cenozoic extensional tectonics of the U.S. Cordillera

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INTRODUCTION

The U.S. Cordillera is typical of orogenic belts in its preservation of multiple episodes of extensional, strike-slip, and compressional deformation. Widespread, latest Proterozoic extension established an early Paleozoic passive margin (*Stewart, 1972*; *Bond and Kominz, 1984*). Other events, probably of lesser overall magnitude and extent, include mid-Proterozoic rifting in the Pacific Northwest, resulting in the accumulation of the Belt Supergroup, late Paleozoic rifting along the continental margin arc (*Miller and others, 1984*) and perhaps also within the craton and miogeocline (*Kluth, 1986*), and Mesozoic extensional tectonism in the "hinterland" of the Mesozoic fold and thrust belt (*Allmendinger and Jordan, 1984*).

The most clearly expressed extensional event occurred during Cenozoic time, creating one of the Earth's few well-exposed regions of large-magnitude intracontinental extension. While the crust is thinned as often as it is thickened, major extension and thinning ultimately result in subsidence and burial. Most of the planet's extensional tectonic record lies beneath thick syn- and post-rift sediments, or has been overprinted by compressional events. The Cenozoic extended region in the Cordillera remains above sea level, despite large extensional strains recorded in upper crustal rocks. This situation results from thin lithosphere in the region and the fact that most of the extended terrain is still underlain by continental crust 25 to 35 km thick (*Thompson and Zoback, 1979*; *Braile and others, 1989*), which may have been twice as thick prior to extension. The well-exposed, unmodified, and widely developed Cenozoic extensional tectonism in the Cordillera allows study of extensional processes that are difficult to observe in outcrop elsewhere, such as on passive continental margins.

During the 1980s, there was a major rethinking of Cordilleran-extension concepts. In the mid-1970s, after nearly a century of doctrine attributing Basin and Range physiography to block faulting (*Gilbert, 1874*; *Davis, 1903*), Cordilleran extension was viewed as being modest (10 to 15 percent widening), accommodated in the upper crust mainly on widely spaced, steeply dipping fault planes (Stewart, 1971; Thompson and Burke, 1974). Based on concepts independent of this doctrine, large-magnitude Cenozoic extension (100 percent widening), mainly predating the development of basin-range fault-block structure, had been proposed as early as the mid-1960s (Hamilton and Myers, 1966). The 1980s witnessed a shift of consensus toward this latter view. With this shift came renewed skepticism that all Basin and Range physiography is controlled by block faulting (as it is described in Stewart, 1971, for example), reminiscent of dissenting views expressed around the turn of the century (e.g., *Spurr, 1901*). The principal catalyst of the new consensus was the discovery of large, shallowly dipping normal faults (detachments) that often juxtapose upper crustal rocks on metamorphic tectonites from the middle crust (a structural association known as a Cordilleran metamorphic core complex; Crittenden and others, 1980). The impact of this consensus has been felt worldwide, prompting reappraisal of long-studied field relations not only in the Cordillera, but in orogens such as the Alps and Caledonides (e.g., *Selverstone, 1988*; *Serrane and Siguret, 1987*).

This chapter synthesizes the results of this upheaval for the Cordillera. The discussion first covers observations of extensional tectonism in the upper crust, constrained primarily by geologic data, and summarizes geophysical observations relevant to understanding the deeper structure of the lithosphere. These observations are synthesized to present an interpretive cross-sectional model of extension of the Cordilleran lithosphere in the central part of the Basin and Range province, which emphasizes fluid behavior of the middle part of the crust during the early stages of extension. Finally, the distribution of extensional tectonism in space and time is reviewed to provide a basis for discussion of the causes of extension. For readers with limited time, the figures and summary section constitute a self-contained overview of the chapter.

UPPER CRUSTAL OBSERVATIONS

Pre-Cenozoic framework and extensional provinces

The pre-Cenozoic Cordillera is divisible into three main paleogeographic elements (Fig. 1): cratonic North America, the Paleozoic Cordilleran miogeocline, and rocks accreted to North

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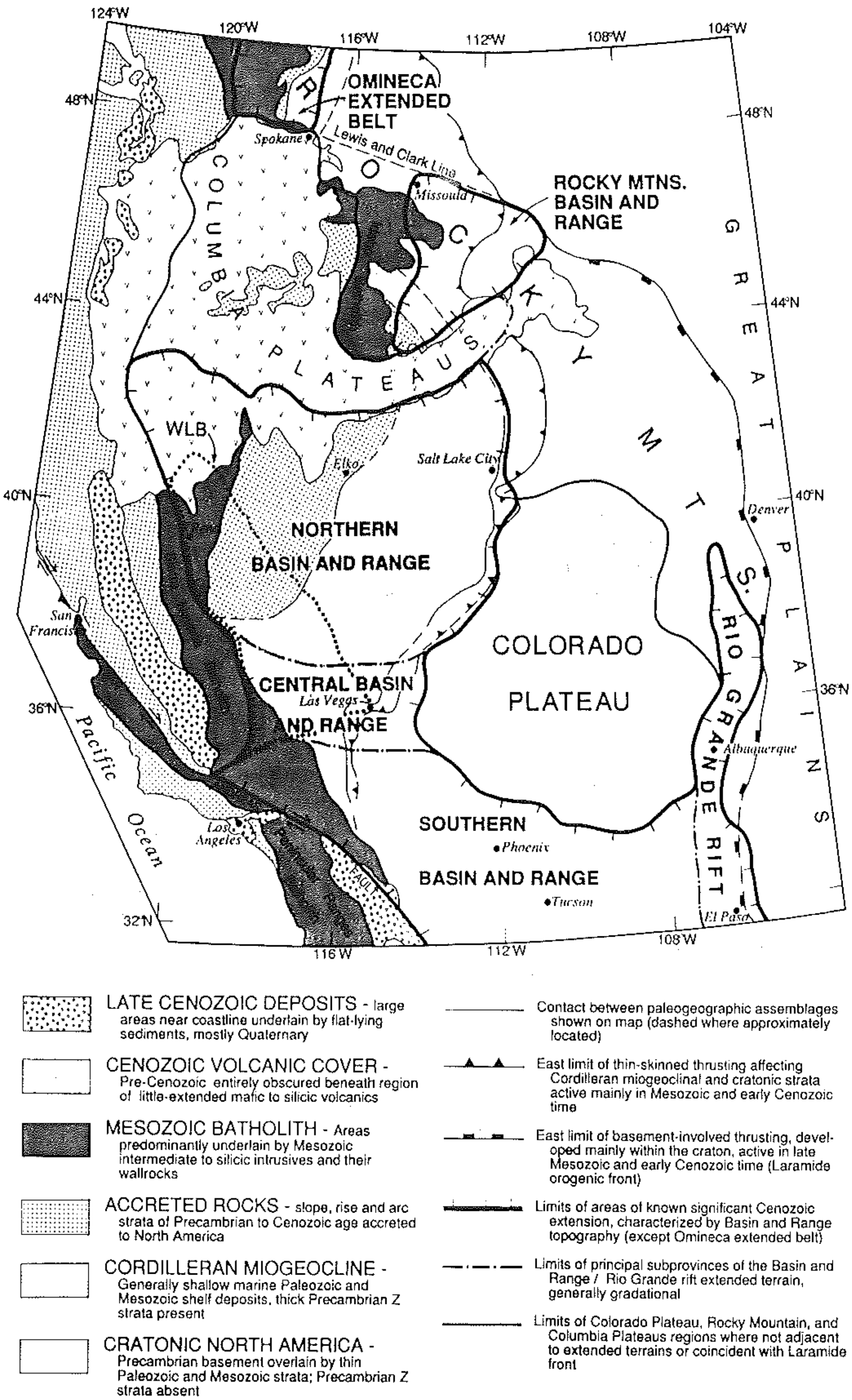


Figure 1. Map showing distribution of extensional provinces relative to major pre-extensional tectonic elements. See text for discussion.

America during various Paleozoic and Mesozoic orogenies. Following most of the accretion, during Cretaceous and early Tertiary time, large areas of cratonic North America and previously undeformed portions of the miogeocline experienced crustal shortening (*Allmendinger, this volume*), broadly synchronous with the emplacement of granitic batholiths along the western side of the orogen (Fig. 1). Where developed within the miogeocline, this shortening was accommodated by east-vergent thin-skinned thrusting of the miogeoclinal wedge and adjacent cratonic cover, forming the Cordilleran thrust and fold belt. In cratonic areas, shortening was accommodated by basement-involved thrusting of variable trend and vergence in latest Cretaceous and early Tertiary time (Laramide orogeny; Fig. 1).

Upper crustal extensional tectonism in the Cordillera is best developed within areas of pre-Cenozoic tectonism. While Cenozoic normal faults occur throughout the Cordillera, four subregions or extensional provinces have been identified that have experienced substantial extensional strain, generally in excess of 10 to 15 percent elongation over original width. These include the Basin and Range province, Rio Grande rift, Rocky Mountains Basin and Range, and Omineca extended belt (Fig. 1). The latter two names are introduced here, although extension in these areas has long been recognized. Within the Basin and Range province, which has traditionally been subdivided into a northern and southern part (the northern being mostly coincident with the Great Basin, a large region of interior drainage), a threefold division into a northern, central, and southern Basin and Range now seems appropriate based on contrasting histories and structural styles, elaborated below (Fig. 1). Another commonly distinguished subprovince in the Basin and Range includes a broad zone in the western parts of the central and northern Basin and Range known as the Walker Lane belt (e.g., Stewart, 1988; Fig. 1, WLB), wherein the trends of basin ranges are highly irregular, in contrast to the more-uniform northerly trends of ranges to the east.

Regions outside of these provinces include: (1) those that have not experienced major extension; and (2) those in which major extension might have occurred, but is either obscured by younger deposits or not yet recognized owing to structural complexity. The Colorado Plateau and most of the Rocky Mountains have been locally mildly extended, but are mainly intact. The Columbia Plateaus comprise a large region of the Pacific Northwest underlain by little-extended Neogene volcanic strata, beneath which substantial extension may have taken place (Hamilton and Myers, 1966) but is not yet documented (Fig. 1). Areas west of the Mesozoic batholiths (Cascades, Coast Ranges, and Continental Borderland off southern California) are all areas where significant extensional tectonism has been proposed (e.g., *Platt, 1986*; *Jayko and others, 1987*), but there is no consensus as to the age, distribution, or magnitude of extension in these areas.

A factor in the pre-Cenozoic framework of noncratonic regions that appears to predispose the upper crust to remain stable during Cenozoic time is the presence of intense Cretaceous plutonism, as in the Peninsular Ranges, Sierra Nevada, and Idaho batholiths (Fig. 1). The western boundaries of the extensional provinces lie within or near the eastern margins of the batholiths. Thus, the Mesozoic magmatism has apparently strengthened the upper crust (but not necessarily the whole lithosphere) in areas where it was most intense, creating a ribbon of firm ground that was later pulled away from cratonic North America, leaving the extensional provinces in their wake (Hamilton and Myers, 1966). The eastern boundaries of the extensional provinces follow specific pre-extensional boundaries locally, but do not have a simple overall pattern. Thus, although upper crustal extension generally does not affect cratonic areas that were not shortened in Cretaceous or early Tertiary time, the eastern limit of extension may lie anywhere from internal portions of the miogeoclinal thrust belt (Omineca extended belt) to the outermost areas of Laramide deformation bordering the Great Plains (Rio Grande rift).

Domainal character and asymmetry of large-scale extensional tectonism

Within the extensional provinces, the magnitude of extension and upper crustal thinning is strongly heterogeneous, partitioned between areas of low and high strain (e.g., *Davis and Burchfiel, 1973*; Proffett, 1977; Guth, 1981; Wernicke and others, *1982*, 1988; Chamberlin, 1983; Miller and others, 1983), especially in the early stages of extension. Plate 8 subdivides the extensional provinces of the Cordillera into two main tectonic elements: strongly extended domains and stable blocks (or simply extended domains and blocks). Cordilleran metamorphic core complexes, described in more detail in the next section, lie within the extended domains, but not all of the extended domains contain core complexes.

The partitioning of strain within the upper crust is documented in many areas, although portions of Plate 8 are interpretive. The distribution of extended domains and stable blocks in some areas is highly uncertain, particularly in the western part of the Basin and Range. Strongly extended domains in the southern Basin and Range are best understood in the northeast, including the Pinaleño, Catalina, South Mountains, Whipple, and Mojave extended domains (Pl. 8), active mainly in Oligocene and early Miocene time. Areas to the southwest are also strongly extended, but the details of their distribution, particularly near the Mexican border and southward, are still emerging. The distribution of extended domains in the central Basin and Range is well understood across the entire province, including the Sheep Range and Death Valley domains to the west and the Lake Mead domain to the east (Pl. 8), active principally in mid-Miocene and later time. As in the southern Basin and Range, the distribution of extended domains is better understood in the eastern part of the northern Basin and Range, especially in a transect across the Sevier Desert, Snake Range, and Ruby Mountains domains (Pl. 8). Activity in these areas ranges in age from Oligocene to Recent. Farther north in the Rocky Mountains Basin and Range, the Pioneer Mountains and Bitterroot extended domains form a narrower belt of strong extension than in the northern Basin and Range (Pl. 8), active principally in Eocene and Oligocene time. A wider belt of strong Eocene extension exists farther north in the Omineca extended belt, including the Okanagan and Newport extended domains (Pl. 8).

Significant extension is by no means restricted to the strongly extended domains. The magnitude of extension within the stable blocks varies, but is generally less than a few tens of percent increase over original width. Within the extended domains, elongations are at least an order of magnitude greater, and in many cases are so large that the entire upper crust, except for a few volumetrically minor tectonic fragments, has been entirely removed from the domain. Thus, many of the extended domains shown on Plate 8 are nearly as wide as the magnitude of extension within them. Plate 8 suggests a scale of partitioning such that stable blocks and extended domains alternate with a wavelength of 100 to 200 km, or a distance as wide as 3 to 5 typical basin-range pairs.

A second major property of the extensional provinces is that, within a given extended domain, the normal faults tend to dip in the same direction. The asymmetry suggests a distinction of their across-strike boundaries according to whether the normal faults dip away from or toward the adjacent stable block (Pl. 8). Boundaries where the faults dip away from the stable block will be referred to here as proximal, and those where they dip toward them distal (Wernicke, 1985), in reference to the position of the adjacent stable block with respect to the structurally lowest normal fault in the system. Proximal boundaries, where the structurally lowest fault breaches the Earth's surface, are also referred to as breakaway zones (e.g., Howard and John, 1987). The patterns of asymmetric faulting correspond approximately to the tilt domains outlined by Stewart (1979) on the basis of Tertiary-strata dip direction in the Basin and Range. The tilt domains include both extended domains and stable blocks shown on Plate 8. Stewart (1979) identified across-strike tilt boundaries as synformal (strata dip toward the boundary) and antiformal (strata dip away from the boundary), which generally lie within stable blocks whose boundaries are both either proximal or distal, respectively.

In map view, the strongly extended domains typically are elongate perpendicular to their extension direction, and tend not to continue with constant net extension for large distances along strike, often terminating quite abruptly (Pl. 8). Domains with opposite asymmetry of normal faulting may closely interact with one another, both along and across strike of normal faults within the domains. Three principal end-member classes of termination and interaction of extended domains along strike appear to be most common (Fig. 2). One is a transform- or transfer-type boundary (*Burchfiel and Stewart, 1966*; *Davis and Burchfiel, 1973*), in which the extension direction is parallel to strike-slip faulting. Another is a mixture of transform and pure wrench faulting, in which crudely hexagonal blocks separate along strike-slip faults (Hill, 1982), such that the overall horizontal extension direction does not parallel the strike-slip faults. With this mechanism, the strain field for the crust as a whole is constrictional, with local, upper crustal shortening possible with an axis perpendicular to the extension direction. The third principal class of interaction, which may or may not exhibit strike-slip faulting, is referred to as an accommodation zone (Fig. 2), consisting of two main types: high relief and low relief, depending on the opposed-asymmetry geometry of interacting domains (Fig. 2; terminology after Rosendahl, 1987).

It is possible that these mechanisms may act in concert near the ends of domains, although there appear to be good examples

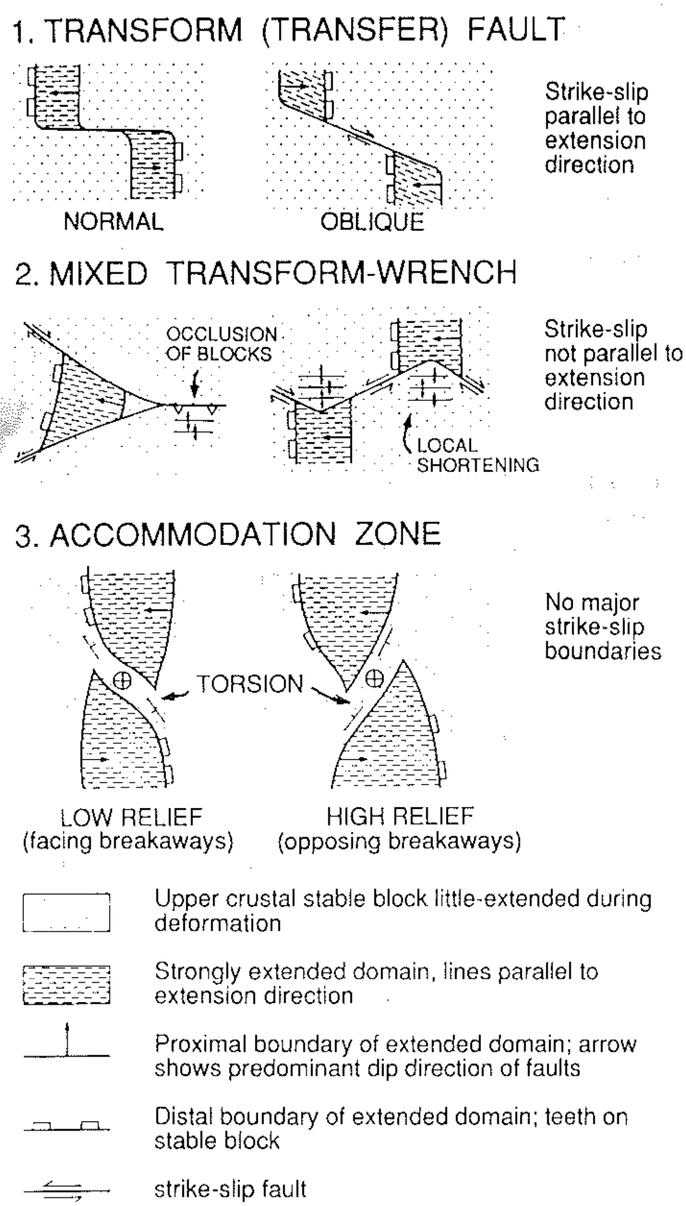


Figure 2. Models of along-strike termination of strongly extended domains and interactions with stable blocks. Examples of oblique transfer and mixed transform-wrench faulting appear to be common in the central Basin and Range and Walker Lane belt, while more-diffuse accommodation-zone terminations may typify the remainder of the extended Cordillera (Pl. 8).

proximal distal

of all of the types shown in Figure 2 at various places in the Cordillera, with the exception of the normal transfer boundary (Fig. 2; Pl. 8). The best examples of transform and mixed transform-wrench boundaries, at a variety of scales, are found in the central Basin and Range and a portion of the northern Basin and Range that includes the Walker Lane belt (e.g., *Anderson, 1973*; Stewart, 1988). Strike-slip faulting in association with extension is pervasive there, and is one of the criteria that distinguishes the central Basin and Range and much of the Walker Lane belt from the other subprovinces designated on Figure 1 (e.g., *Wright, 1976*). The strike-slip faulting appears to accommodate regional constriction of the upper crust in the central Basin and Range (e.g., Wernicke and others, 1988), although examples of transfer-type faults are found in that area as well (e.g., Burchfiel and others, 1987). Outside of the central Basin and Range/Walker Lane belt subprovinces, the occurrence of discrete strike-slip faults along the boundaries of strongly extended domains is limited, with most of the interactions dominated by accommodation-zone-style tectonism (Fig. 2; e.g., *Faulds and others, 1990*). One exception is the Lewis and Clark line, interpreted to be in part an oblique transfer structure (e.g., *Rehrig and Reynolds, 1981*; Pl. 8). A number of transfer structures have been described from mildly extended areas, such as in the mildly extended volcanic cover in the northwestern corner of the Basin and Range (*Lawrence, 1976*).

A potentially important deformation style within the extended Cordillera involves rotation of crustal blocks about vertical axes. Such rotations are suggested by the accommodation zone geometry shown in Figure 2, where the ribbon of crust separating domains is not only bent by torsion about a horizontal axis, but is also rotated about a vertical axis. An end-member case would be a domain wherein strain was accommodated by domino-style rotation of blocks and bounding strike-slip faults about vertical axes with no vertical strain (*Freund, 1974*). This mechanism and derivatives have been suggested to explain mapped fault patterns and paleomagnetic declination anomalies in the southwestern U.S. Cordillera, especially in the western Transverse Ranges area near Los Angeles (*Luyendyk and Hornafius, 1987*). Deformation of this type has been proposed for the central Basin and Range and Walker Lane belts on a more local scale (*Ron and others, 1986*; *Nelson and Jones, 1987*; Stewart, 1988), however the relative importance of this mechanism in the domains outlined on Plate 8 is highly uncertain.

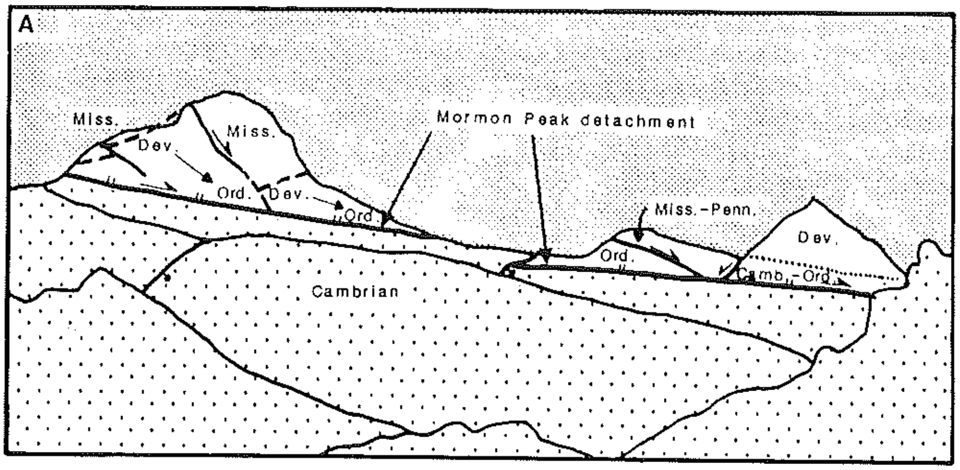
Cross-sectional kinematics of extended upper crust

Generalities. The classical model of extensional tectonics, derived from observations of faults bounding a number of the present-day ranges in the Basin and Range province (e.g., *Gilbert, 1874*; *Davis, 1903*), holds that the upper crust extends via brittle fracturing on moderately to steeply dipping (typically 60°, mostly >40°), widely spaced, major, normal faults, whose net displacements are a small fraction of spacing between range blocks. This model implies crustal extension to be about 10 to 15 percent increase over original width (*Thompson, 1960*; Hamilton and Myers, 1966; Stewart, 1971; Thompson and Burke, 1974). Increasing documentation of shallowly inclined Cenozoic normal faults during the 1970s and 1980s (many of which had already been described and attributed to either compression or mass-wasting phenomena during the first 70 years of the century) has required significant revision of this model as a complete description of Cenozoic extensional tectonism in the Cordillera (e.g., Hamilton and Myers, 1966; Anderson and others, 1983; *Coney, 1987*).

The field observation most significant to the new conceptual framework is the widespread occurrence of large subhorizontal faults of Cenozoic age across which stratigraphic or structural succession is omitted, and which serve as basal dislocations for mosaics of imbricate normal faults, first documented by *Ransome and others (1910)* in the central Basin and Range. A number of studies in the early 1970s (presaged by *Gilbert, 1928*) identified shallowly inclined normal faults of limited extent (less than 10 km of exposed width parallel to slip direction for a single fault) as an important mode of accommodation of large extensional strain in the crust, at least in some areas (e.g., Anderson, 1971; Armstrong, 1972; Wright and Troxel, 1973; Proffett, 1977). By 1980, consensus held that in large areas of the Basin and Range and Pacific Northwest extension was accommodated by a class of faults that are typically exposed as single subhorizontal fault zones across distances of a few tens to as much as 50 to 60 km parallel to their transport directions. These faults emplace strongly extended upper crustal fault blocks similar to those mapped by Anderson (1971) above deeper crustal rocks.

These faults, known as detachments or detachment faults (e.g., Davis and others, *1980*, 1986), reveal in outcrop how upper crustal fault blocks interact with deeper structural levels (Fig. 3). The fundamental observation is that the upper crustal faults either merge with, or are truncated by, the detachments. Figure 3 illustrates a typical example of these faults, and some of the complexities typically encountered where steep faults are truncated by shallow ones.

Footwall metamorphic tectonite. The footwalls of detachments are often composed of metamorphic tectonites that yield Cenozoic cooling ages for thermochronometers with closure temperatures in the range of 100° to 500 °C (e.g., *Armstrong and Hansen, 1966*; *Keith and others, 1981*; *Dokka and others, 1986*; *Miller and others, 1988*), and contain metamorphic assemblages that suggest depths of Mesozoic and Cenozoic metamorphism in excess of 10 km, up to as much as 30 km (e.g., *Labotka, 1980*; *Davis and others, 1980*; *Hamilton, 1982*; *Wernicke, 1982*; Bartley and Wernicke, 1984; *Labotka and Albee, 1988*; *Anderson, 1988*; *Hodges, 1988*; *Hodges and Walker, 1990*). The idea that the tectonites may to some extent be genetically related to extension was first suggested by *Armstrong (1963)* in his discussion of the Raft River, Ruby Mountains, and Snake Range areas of



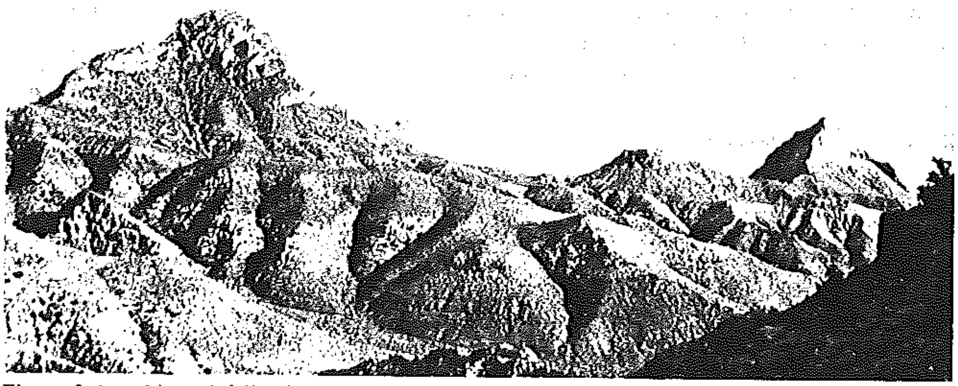
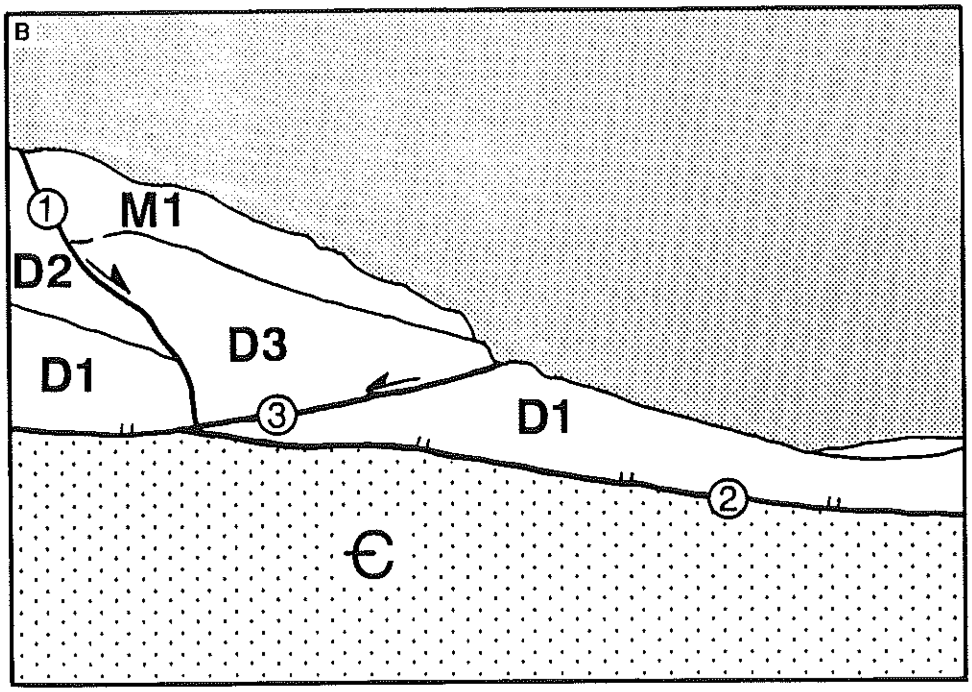


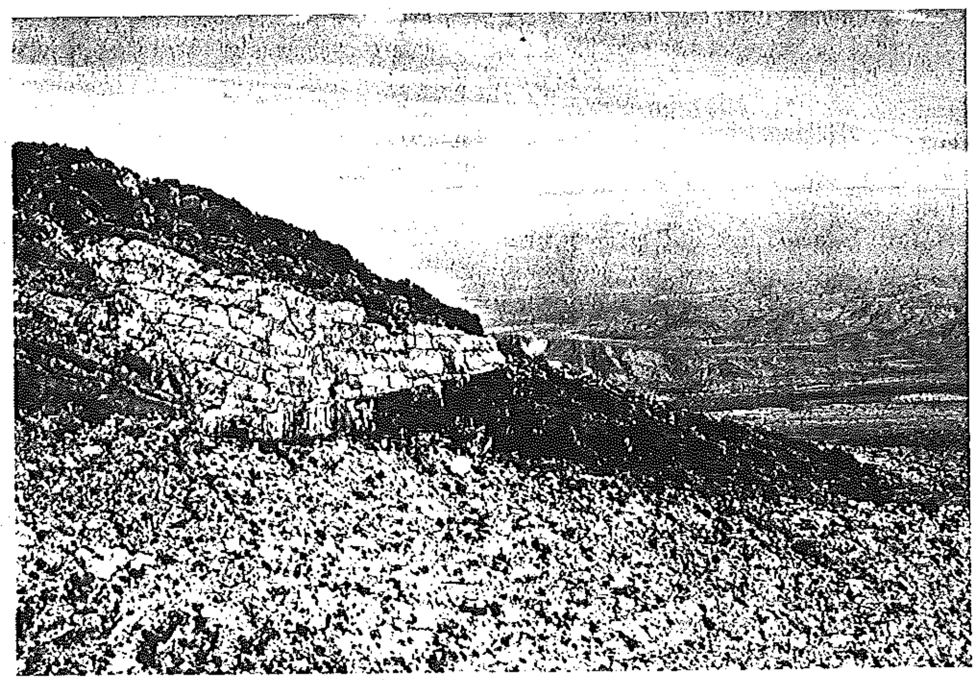
Figure 3 (on this and following two pages). Examples of detachment systems, central Basin and Range. (A) View south of Mormon Peak detachment (see Fig. 4a, lowest frame). Tilted hanging wall blocks are cut by normal faults that do not cut the detachment. Detachment cuts gently (about 10°) downsection to the west through unmetamorphosed footwall strata. Vertical relief is about 800 m. (B) View north of Mormon Peak detachment (Fig. 4a, lowest frame), showing timing relations among associated faults. Fault 1 places highest Devonian (D3) and lowest Mississippian (M1) strata against lower Devonian units (D1, D2) and is truncated by fault 3, which merges with the detachment at very low angle. The truncation of high-angle faults by low-angle faults creates a situation where hanging wall imbricate faults sharply abut the detachment, making them difficult to restore palinspastically (e.g., Davis and Lister, 1988). Vertical relief is about 100 m. (C) View north of Tucki Wash fault (arrows show trace), Death Valley region. Hanging wall has moved to the left, relative to the footwall. The fault cuts steeply downsection through hanging wall Cambrian strata on the right, curving into parallelism toward the left. Eocambrian strata in the footwall are subparallel to the fault along both steep and shallow portions, apparently monoclinally warped as they were dragged out from beneath the hanging wall (analogous to point C in second frame from the top of Figure 5b). Near upper arrow, a younger subhorizontal fault cuts the Tucki Wash fault, which may have been flexurally rotated once in the footwall of younger, structurally higher faults (e.g., as in Fig. 5b). Curvature of the fault plane lay at about 6 km below surface while fault was active.

east-central Nevada. He referred to these ranges as "arch ranges," stating (p. 133):

> The three best examples of arch ranges are composed of rocks which were probably deep below the surface before normal faulting. Because of this, the rocks of which they are composed would have been more plastic than the shallow superstructure. As a consequence, these ranges could have formed during uplift, perhaps developing beneath horsts formed at shallower tectonic levels.

The tectonites are usually lineated and foliated mylonites whose foliation is broadly parallel to overlying detachments, and whose stretching directions are observed to be roughly parallel to both upper plate extension direction and the slip direction on the detachment (Davis and Coney, 1979; G. H. Davis, 1980; *Rehrig and Reynolds, 1980*). Further, S-C fabrics (*Simpson and Schmid, 1983*) in the tectonites in some instances suggest that they devel-





oped in part via simple shear in the same sense as displacement on the overlying detachments (e.g., *Snoke and Lush, 1984*; Davis and others, 1986; *Malavieille, 1987*; *Gaudemer and Tapponnier, 1987*; Lee and others, 1987; Parrish and others, 1988).

Areas containing the threefold association of detachment, hanging wall imbricate normal fault blocks, and footwall metamorphic tectonite are referred to as Cordilleran metamorphic core complexes, or simply core complexes (Davis and Coney, 1979; articles in Crittenden and others, 1980; Armstrong, 1982). Their origin is widely believed to result from tectonic denudation of deep structural levels along normal shear zones that initially penetrated to depths of 10 to 15 km and have net displacements measured in tens of kilometers (e.g., *Howard and others, 1982*; Spencer, *1982*, 1984, *1985*; *Davis, 1983*; Allmendinger and others, 1983; Bartley and Wernicke, 1984; Reynolds, 1985; Reynolds and Spencer, 1985; *Wernicke and others, 1985*; Dokka, *1986*, 1989; Davis and others, 1986; *John, 1987*; *Hodges and others, 1987*; Davis and Lister, 1988; Snoke and Miller, 1988;

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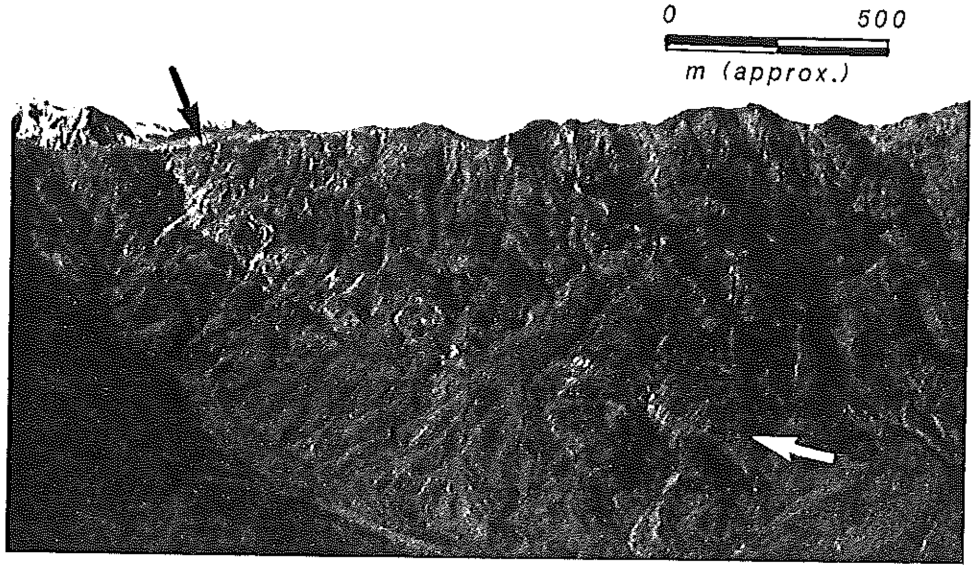


Hamilton, 1988; Parrish and others, 1988; *Glazner and others, 1989*). In this manner, some of the tectonites may form along the deep-seated portions of the detachment fault zones early in their history, and be progressively overprinted by brittle deformation and finally the detachment fault itself during ascent to the surface (e.g., Davis and others, 1986).

A major complication in interpreting the tectonites as products of extension is that because they are uplifted from mid-crustal levels, they often contain older metamorphic fabrics developed during Mesozoic or even Precambrian time that are difficult to date because they were at high temperatures in the Cenozoic (e.g., *Armstrong and Hills, 1967*). Thus, while the age and tectonic significance of many of these tectonites have been controversial, geochronological and structural analyses in many of the core complexes have resulted in substantial progress in testing kinematic models for their origin (e.g., *Reynolds and Rehrig, 1980*; Reynolds, 1985; *Dokka and others, 1986*; *Wright and others, 1986*; *Wright and Snoke, 1986*; Lee and others, 1987; *DeWitt and others, 1986*; *Miller and others, 1988*; Parrish and others, 1988; Walker and others, 1990; *Asmerom and others, 1990*). The 1990s promise to be a decade of major progress in understanding the relationship between upper crustal and deep crustal levels as more data from the core complexes is gathered. In particular, understanding the role of pre-mylonitic, relatively high-temperature fabrics found in some core complexes that are of Tertiary age may add significant new insight into deep crustal processes active during extension (*Wright and Snoke, 1986*; *Wright and others, 1986*; Lee and others, 1987), as discussed below.

The strongly extended domains on Plate 8 are characterized by core complexes and detachment faults, but some detachments do not expose metamorphic tectonite in their footwalls (e.g., Figs. 3a and b). In addition, some highly extended domains do not contain detachments at current levels of exposure, because of insufficient erosion of hangingwall imbricate fault mosaics, inefficiency of tectonic denudation, or both. Thus the regional distribution of core complex tectonites (e.g., Coney and Harms, 1984; Wust, 1986) does not completely reveal the distribution of strongly extended crust.

Magnitude of extension. Sequentially restored regional cross sections across two extended domains, the Mormon Mountains and Snake Range domains (Fig. 4; Pl. 8), illustrate the salient characteristics of upper crustal fault kinematics encountered in the extended domains. The amount of extension in these examples is a large fraction of the exposed width of the domains (>80 percent in both cases), which is typical of domains for which reconstructions have been attempted. Despite roughly equivalent amounts of extension, the Snake Range domain contains a core complex (the Snake Range) while the Mormon Mountains area does not, reflecting the greater depths brought to the surface by the normal faults in the Snake Range domain. Only a small volume of the upper 7 to 10 km of the crust is preserved within the domains relative to their widths, with overall extension of the upper crust in excess of 300 to 400 percent increase over original width. In contrast, the stable domains are extended a few tens of percent or less, generally along steep normal faults. In other words, the two stable blocks on either side of the domain are separated along a system of faults whose pre-extension trajectories through the upper crust occupy a volume that is small relative to that of the stable blocks.

Total extension across the Basin and Range province is about 250 km based on reconstruction of the extended domains in the central Basin and Range (Wernicke and others, 1988). Assuming the large batholithic blocks on western margins of the

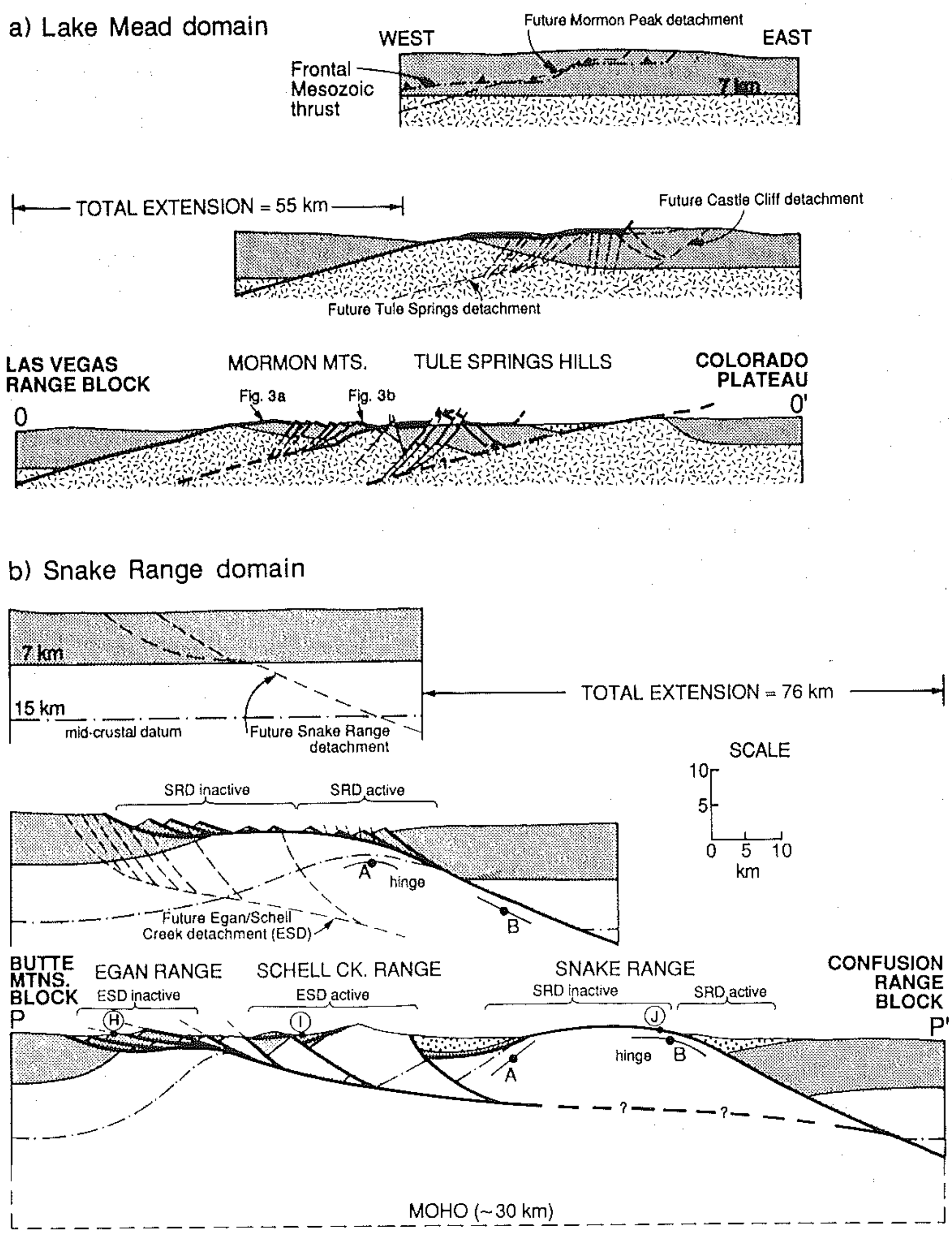


Figure 4. Stepwise-restored cross sections through two highly extended domains. See Pl. 8 for locations of sections. Scale is the same for both sections. (a) Mormon Mountains domain, showing position of frontal thrust fault of the foreland fold and thrust belt, which in the Mormon Mountains area is a decollement in Middle Cambrian strata. Pre-extension datum at 7 km corresponds approximately to the nonconformity between Proterozoic crystalline basement and Phanerozoic sediments (after Axen and others, 1990). (b) Snake Range domain. Pre-extension depths of 7 and 15 km are not intended to correspond to any particular stratigraphic horizon, although most of the faulting is developed within regionally conformable miogeoclinal sediments (after Bartley and Wernicke, 1984, with some modifications based on Hauser and others, 1987). Note migrating hinge in footwall of Snake Range decollement. Localities H, I, and J correspond to those on Figure 6. See text for discussion.

extensional provinces have not appreciably rotated with respect to the interior of North America during extension, the magnitude of extension in the northern and southern Basin and Range may be comparable (e.g., *Bogen and Schweickert, 1985*; *Frei, 1986*; Levy and Christie-Blick, 1989). This gives elongations of a factor of 3.5 for the central Basin and Range and 2.0 for the northern Basin and Range.

Initial dip of detachments. Detachment faults are typically regionally subhorizontal, but their initial trajectory through the crust has been a subject of debate. On one hand, studies of earthquake focal mechanisms from rapidly extending areas, including the Basin and Range, indicate that seismic slip on low-angle (<30°) normal faults seldom if ever occurs (Jackson, 1987). On the other hand, geologic examples of some detachments (e.g., Allmendinger and others, 1983; Bartley and Wernicke, 1984; Reynolds, 1985; *Spencer, 1985*; *Wernicke and others, 1985*; Burchfiel and others, 1987; *John, 1987*; Howard and John, 1987; Davis and Lister, 1988; Dokka, 1989; Axen and others, 1990; Foster and others, 1990) suggest that they initiated at dips of 30° or less, and were active at even shallower dips in the upper crust throughout their history of movement.

The footwalls of many normal faults contain indicators showing increasing structural depth in the direction of hanging wall transport, such as downcutting through stratified rocks, increasing metamorphic grade, and decreasing cooling ages. For example, the Mormon Peak detachment initiated with an average dip of 20 to 25° through the upper 7 to 8 km of crust (*Wernicke and others, 1985*; Axen and others, 1990), cutting gently (15 to 20°) downsection to the west across the basal decollement of the Mesozoic fold and thrust belt and into autochthonous Precambrian basement (O–O′, Pl. 8; Figs. 3a, b, and 4a). Though controversial (cf. Miller and others, 1983; Bartley and Wernicke, 1984; *Gans and others, 1988*), the Snake Range detachment probably initiated with an average dip of 30 to 40° through the upper 10 km of the crust (Fig. 4b), based on the author's interpretation of: (1) the angle its subsurface trace makes with the adjacent Confusion Range block (Allmendinger and others, 1983); (2) the average fault-bed angle in hanging wall rocks that restore palinspastically against the west side of the Confusion Range block (angles measured from Fig. 6 *in* Miller and others, 1983, and Figs. 16 and 20 *in Rogers, 1987*); and (3) metamorphic and thermochronometric gradients in its footwall (Miller and others, 1983; Lee and others, 1987). However, a large proportion of normal faults within the upper 5 to 10 km of the crust have steeper average initial dips, such as the Egan–Schell Creek detachment (Fig. 4b; Armstrong, 1972; *Wernicke, 1981*; *Gans and others, 1989*) and many of the faults in the central Basin and Range (Figs. 3c and 7; *Wernicke and others, 1989*).

While detachments may initiate with steep dips in the upper 5 to 10 km of the crust, the width of exposed footwall of most core complex detachments relative to changes in structural depth across strike seem too large to permit average initial dips through the upper 15 km of crust to exceed 30 to 40° (e.g., *Compton and others, 1977*; John, 1988; Wernicke and others, 1988; Spencer and Reynolds, 1989; Richard and others, 1990; Foster and others, 1990). If we consider detachment footwalls to show a range of between 10 and 15 km of downcutting from the breakaway zone to deepest exposed metamorphic tectonite, across a 30 to 60 km width of relatively intact footwall, then initial dips would lie in the range of 10 to 27°. If the detachments characteristically dipped 60° in at least the upper 5 km, then average initial dips would range from 5 to 21° below 5 km. Combined with seismic reflection data from the Basin and Range and passive margin settings (e.g., Allmendinger and others, 1983; Smith and Bruhn, 1984; *Cheadle and others, 1987*; *Tankard and Welsink, 1987, 1990*), the weight of evidence suggests that, in the depth range of 5 to 15 km, initial dips may vary from less than 10° up to 60°, but are commonly 30° or less.

Footwall uplift and flexure. Strongly extended domains are characterized by uplift and flexure on their margins, apparently driven by the isostatic forces that accompany footwall unloading (e.g., Spencer, *1982*, 1984, *1985*; *Howard and others, 1982*; Bartley and Wernicke, 1984; *Wernicke and Axen, 1988*; *Buck, 1988*). In cases where the detachment is inclined steeply through a large fraction of the upper crust, flexure of footwall strata accompanied by steep faulting may accommodate uplift on the proximal boundaries (e.g., the Colorado Plateau and Butte Mountains blocks, Fig. 4; *Wernicke and Axen, 1988*). In cases where the faults are initially gently inclined, the uplifts may form broad domes (such as beneath the Mormon Peak detachment, Fig. 4a, or the Snake Range detachment, Fig. 4b), which characterize core complex detachments (Davis and Coney, 1979; Spencer, 1984).

Sequential development. Strongly extended domains usually display complex overprinting relationships among multiple generations of normal faults, both within individual hanging walls of detachments, and among multiple generations of detachments (Fig. 4). Two kinematic processes are commonly suggested (Fig. 5). In one (Fig. 5a), arrays of fault blocks rotate synchronously, as in a row of toppling dominos (e.g., *Chamberlin, 1978*). When the first array reaches too low a dip for continued movement, a new one develops, rotating the earlier faults to still shallower dips, possibly even reversing their dip direction (e.g., *Morton and Black, 1975*; Proffett, 1977; Chamberlin, 1983; *Gans and Miller, 1983*). Within such a system, rotation of fault blocks occurs simultaneously across a domain, and a large fraction of the rotation of initially steep faults to shallow dips is accomplished while the faults are inactive.

In another process (Fig. 5b), normal fault blocks sequentially detach, meatslicer style, from the distal block, with rotation occurring principally as a result of isostatic rebound (Spencer, 1984; Bartley and Wernicke, 1984; *Wernicke and Axen, 1988*; *Hamilton, 1988*; *Buck, 1988*). Consequences of this process are that within a given hanging wall, cessation of normal faulting progresses from proximal to distal areas (i.e., diachronous rotation), and that nearly all of the rotation of individual faults may occur without the development of another generation of faults. An additional consequence is the migration of a monoclinal flexure through the footwall, such that the footwall hinge about which rotation of the detached slices occurs (points A and B, Fig. 4b; points A through D, Fig. 5b) propagates in wave-like fashion from proximal to distal sides of the domain. Structural analysis of a core complex in the Mojave extended domain (Pl. 8) supports the passage of such a monoclinal flexure through footwall metamorphic tectonite during extension (*Bartley and others, 1990*).

These two processes are not mutually exclusive, and likely interact at a variety of scales. For example, each sequentially detached block in Figure 5b could be divided into splays such as those shown in Figure 5a, which could then experience two periods of domino-style rotation. This is likely the case for the development of blocks in the upper plate of the Snake Range decollement in the Snake Range (Miller and others, 1983). However, within the same system, there is evidence for eastward migration of fault block rotation above the Snake Range decollement, in that the fault mosaic above the decollement in the Schell Creek Range area is unconformably overlapped by strata dated at 27 Ma (locality I, Figs. 4b and 6b), yet thermochronology on footwall rocks in the Snake Range suggests significant movement on the decollement until at least 20 to 22 Ma (locality J, Figs. 4b and 6b). A similar pattern is apparent for the structurally lower Egan Range system; the structurally lowest faults in that system are cut by 35 Ma granites (locality J) while higher splays to the east in the Schell Creek Range involve younger rocks (locality I). Correlations of these fault systems across the three ranges are, however, partly interpretive (Bartley and Wernicke, 1984; Hauser and others, 19??; cf. *Miller and others, 1988*).

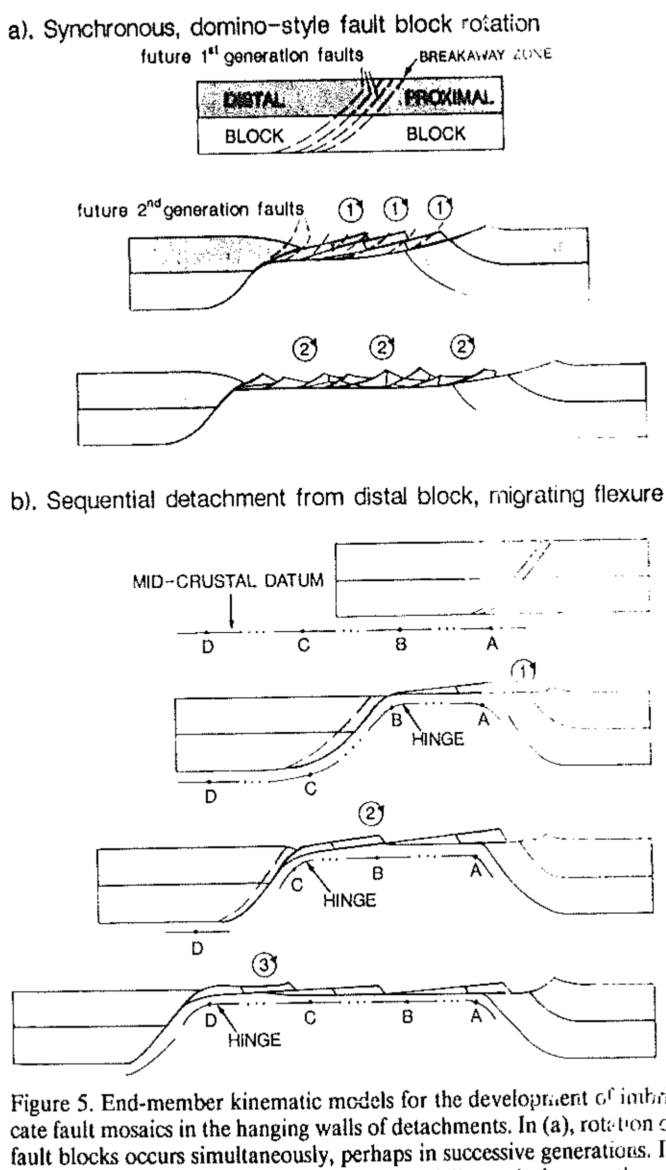


Figure 5. End-member kinematic models for the development of imbricate fault mosaics in the hanging walls of detachments. In (a), rotation of fault blocks occurs simultaneously, perhaps in successive generations. In (b), thin slivers of crust are detached sequentially, such that rotation of each block occurs at different times. Depending on the details of subdetachment strain, a hinge may or may not migrate through the lower plate in (a). Kinematics in (b) seem to require new material to shear out from beneath the distal block for each successive sliver.

An example of regional sequential detachment is found across the central Basin and Range (Fig. 6a), where the cessation of faulting and times of rotation of range blocks migrate eastward (*Wright and others, 1981*; *Hamilton, 1988*; Wernicke and others, 1988). Even at the scale of individual ranges, a progression of faulting within narrow fault blocks in the Panamint Range (Locality A, Fig. ?) apparently upward, such that the footwalls of some faults rotate prior to their hanging walls (Fig. 3c; *Wernicke and others, 1989*).

At present it is not clear whether one or the other of these mechanisms predominates within the extended domains, although the bulk of evidence suggests that the domino mechanism may be restricted to smaller subdomains and that the sequential detachment mechanism is important domain-wide. However, in both of the domains shown in Figure 4, more than one detachment system affects them. In both cases the detachment systems propagate downward, such that the unloaded footwalls of earlier detachments become incorporated into the hanging walls of later ones, opposite to the pattern in Figure 5b. The breakaway zones of the younger systems tend to be within a few kilometers of those of the older ones. Thus, in map view, extension does not propagate very far into the proximal block, but incorporates a large volume of the unloaded footwall of the earlier fault system into its hanging wall. This progression causes younger imbricate faults to rotate (and thereby deactivate) portions of older systems, amplifying the domes created by isostatic rebound. These kinematics create a large-scale pattern of thick fault blocks resembling dominos, but rotation of adjacent blocks is not synchronous. In both the Mormon and Snake Range areas, however, the oldest detachments may have been active in their most distal segments while the younger structurally lower systems were active.

Rate and duration of extension. In several of the strongly extended domains, the timing and magnitude of extension are well enough constrained to permit assessment of the spreading rate. While the cooling histories and upper crustal unconformities associated with block tilting typically show that individual blocks are up-ended over an interval of 1 to 5 m.y. (e.g., *Anderson and*

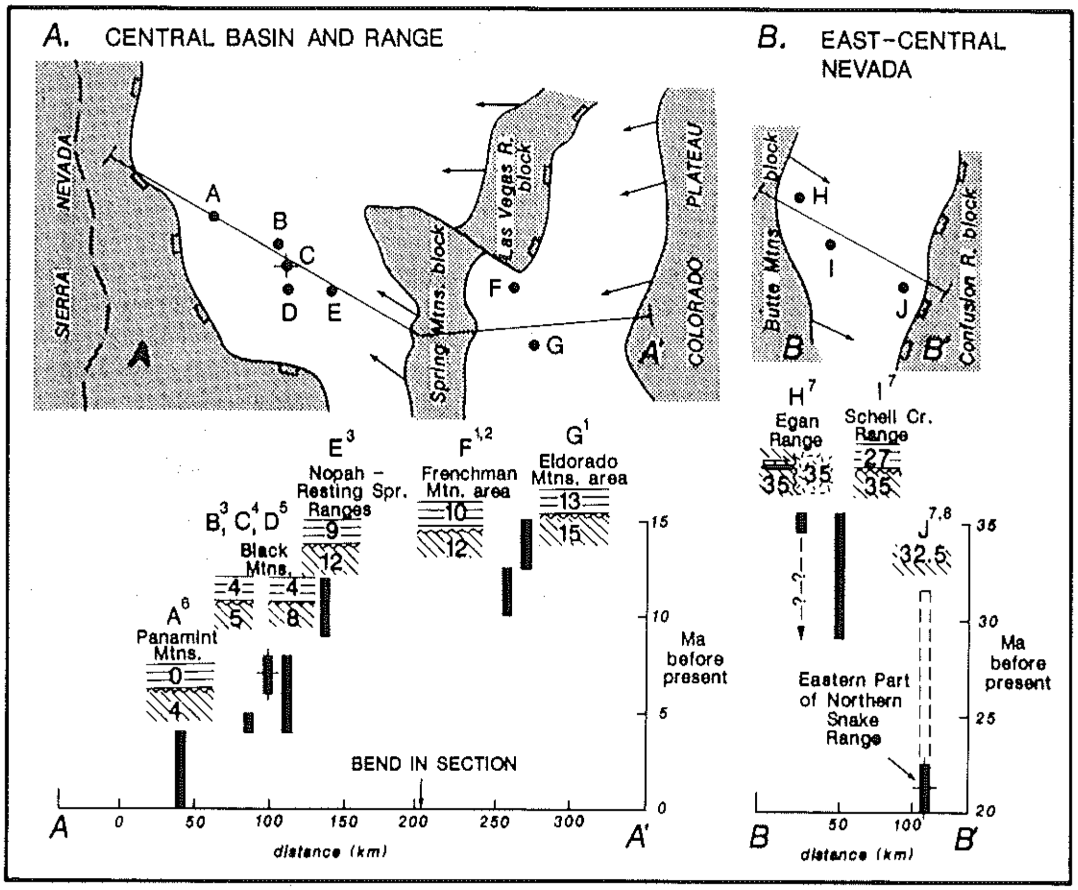


Figure 6. Timing constraints on extension for (A) central Basin and Range domains and (B) Snake Range domain in east-central Nevada (Pl. 8). Constraints are projected onto lines of section, which are not precisely coincident with lines of section shown on Plate 8. Black bars indicate intervals of rotation of fault blocks based on ages of tilted strata, faulting, and intrusion as schematically shown next to symbol. Black bars ornamented with thin tick marks show interval at which deepest portions of exposed core complexes cooled through the blocking temperatures of Ar in micas (about 250 to 350 °C). See text for discussion. Sources: 1, *Anderson and others, 1972*; 2, *Bohannon, 1984*; 3, *Wright and others, 1984*; 4, *Fleck, 1970*; 5, *Holm and others, 1989*; 6, Hodges and others, 1989; 7, *Gans and others, 1989*; 8, Lee and others, 1987.

others, 1972; Dokka and others, 1986; *Holm and others, 1989*; *Davis, 1988*), the duration of major extension across a given domain may be significantly longer, particularly if the domain contains several major blocks up-ended in sequence (Fig. 6). For example, tilting of individual fault blocks in the Lake Mead, Death Valley, and Snake Range domains occurred in as little as 1 m.y., but rapid extension appears to have taken place for 5 to 10 m.y. in the Lake Mead domain, 10 to 12 m.y. in the Death Valley domain, and 10 to 15 m.y. in the Snake Range domain, because the major block rotations are not necessarily synchronous across the domains.

The spreading of these extended domains, assuming 75 km of extension in the Snake Range domain, 100 km in the Lake Mead domain (southern part), and 150 km in the Death Valley domain, are thus 5 to 8 mm/a, 10 to 20 mm/a, and 10 to 13 mm/a, respectively. In the central Basin and Range, the timing and magnitude of extension suggest that the total rate of westward motion of the Sierra Nevada away from the Colorado Plateau was at least 20 mm/a in Miocene time (ca. 10 Ma), but has since slowed to approximately 10 mm/a in the last 5 m.y. (Wernicke and others, 1988).

Complexities in the coupling between upper crustal strain and that of deeper layers (discussed below) and in the three-dimensional kinematics of extension preclude any simple relationship between upper crustal displacement rates and the strain rates of deeper layers. Further, if strain in deeper layers is localized into relatively narrow shear zones, strain rates there would be far greater than those calculated for the lithosphere as a whole. Nonetheless, if we consider the case of the central Basin and Range, averaging the westward displacement rate of the Sierra Nevada across a uniformly extending lithosphere yields strain rates as great as 2×10^{-14} sec^{-1}, although conspiring uncertainties in timing and rate in the other direction yields values an order of magnitude lower (Wernicke and others, 1988). These rates are comparable to or greater than an earlier estimate of strain rates in the strongly extended domains of about 3 to 6 × 10^{-15} sec^{-1} (Zoback and others, 1981). For active deformation, particularly in the northern Basin and Range where strain is spread out over a broad area and is relatively slow, the average rate is in the 10^{-16} sec^{-1} range (Zoback and others, 1981).

Relation of extended domains to Basin and Range topography. In a number of areas, both stable blocks and strongly extended domains contain widely spaced, domino-style faults that control modern Basin and Range topography. For example, a number of regularly spaced ranges in the western part of the northern Basin and Range formed in late Miocene to Recent time as large, simultaneously rotating dominos, whose bounding faults penetrate to substantial depths but accommodate relatively little extension (classical Basin and Range style; e.g., Stewart, 1978; Okaya and Thompson, 1986; Allmendinger and others, 1987).

However, within most of the remainder of the province, Basin and Range topography is not as clearly an expression of deep-seated steep block faults (e.g., *Blackwelder, 1928*). Although partly due to block faulting, much of the topography may result from structures associated with large-magnitude extension, such as the eroded remnants of domiform detachment footwalls (e.g., the Snake Range and Mormon Mountains, Fig. 4), or the topography that may result from the isolation of thin, sequentially detached slices atop rebounded middle crustal rocks (Fig. 5b). This includes the eastern part of the northern Basin and Range, the central Basin and Range (especially the Death Valley region; Fig. 7), and the southern Basin and Range (Wernicke, *1981,* 1985; *Hamilton, 1982*; Anderson and others, 1983; Allmendinger and others, 1983; Burchfiel and others, 1987; Hamilton, 1987; *Dickinson and others, 1987*; Hodges and others, 1989). Thus, the widespread belief that Basin and Range topography represents a late event overprinting the strongly extended domains, and thus predominantly expresses late-stage moderate crustal extension (e.g., Zoback and others, 1981; Eaton, 1982; Coney and Harms, 1984; *Coney, 1987*), explains some of the physiography well, but such block faulting may not be the primary control on the topography in much of the province (e.g., Anderson, 1989).

Areas of large-magnitude extension need not develop Basin and Range topography. For example, the topography within the Omineca extended belt, while generally consisting of northerly trending ranges, lacks the wide alluvial valleys characteristic of the Basin and Range province.

DEEP-CRUSTAL OBSERVATIONS

The geophysical framework of extended portions of the Cordillera, in particular the Basin and Range province, was well established by the end of the 1970s (Thompson and Burke, 1974; Smith and Eaton, 1978; *Lachenbruch and Sass, 1978*; Eaton and others, 1978; *Thompson and Zoback, 1979*). Nonetheless, concomitant with the progress in understanding upper crustal fault kinematics through geologic studies, the 1980s saw major progress in imaging lower crustal and upper mantle structure, through the acquisition of deep seismic-reflection profiles (e.g., Allmendinger and others, 1983, 1987; Smith and Bruhn, 1984; Potter and others, 1986; Okaya and Thompson, 1986; Klemperer and others, 1986; Hauser and others, 1987, *1987b*; *Goodwin and Thompson, 1988*; McCarthy and Thompson, 1988; *Valasek and others, 1989*) and combined seismic-reflection and refraction profiling (*Mooney and Brocher, 1987*; *Catchings and others, 1988*). These studies provide a considerably improved basis for developing models of extensional tectonism at lithospheric scale, which became particularly numerous with the recognition of the core complexes in the late 1970s and early 1980s (e.g., Davis and Coney, 1979; *Rehrig and Reynolds, 1980*; Eaton, *1979,* 1982; *Wernicke, 1981*; *Hamilton, 1982*). A complete survey of all deep geophysical measurements is beyond the scope of this chapter, but there are several observations, in particular that bear on deep crustal structural evolution, that warrant discussion here. These include the distribution of long-wavelength variations in topography and gravity with respect to the strongly extended domains, and the interpreted position of the seismic Moho beneath the extended areas, particularly near the boundaries of strongly extended domains.

The domainal character of upper crustal extension, and the observation that the upper 7 to 15 km of crust has been partially or completely removed from the strongly extended domains, raises important questions about how such a geometry relates to strain at deeper crustal levels. For example, if normal faulting completely denuded the middle crust, then the amount of isostatic rebound of the denuded crust, and hence the magnitude of the topographic anomaly produced by such denudation, should be a function of the density of the medium upon which the lithosphere floats. Similarly, one might expect the tectonic removal of the upper one-third to one-half of the crust to lead to upwarping of the Moho, with an amplitude roughly equal to the amount of material denuded, or perhaps greater if the lower portions of the crust extend preferentially beneath the strongly extended domains. These responses make simple predictions about the density structure of the crust in and near the extended domains (e.g., *Cady, 1980* Wernicke, *1983,* 1985, *1989,* 1990; Spencer and Reynolds, *1984,* 1989; Okaya and Thompson, 1986; *Thompson and McCarthy, 1986, 1990*; Block and Royden, *1988,* 1990).

Regional gravity and topographic patterns in the Basin and Range province show variations at wavelengths of 30 km (scale of basins and ranges) and 150 to 200 km (Eaton and others, 1978; *Fletcher and Hallet, 1983*; *Froidevaux, 1986*) showing a rough correlation with the distribution of Stewart's (1979) tilt domains (Zuber and others, 1986), and thus also with the distribution of extended domains and stable blocks (Pl. 8). For example, in the Omineca extended belt, the broad region of crustal denudation is associated with depressions of 100 to 300 m (*Cady, 1980*), while in the central Basin and Range, a transect from the Colorado Plateau to the Sierra Nevada (Q–Q′, Pl. 8)

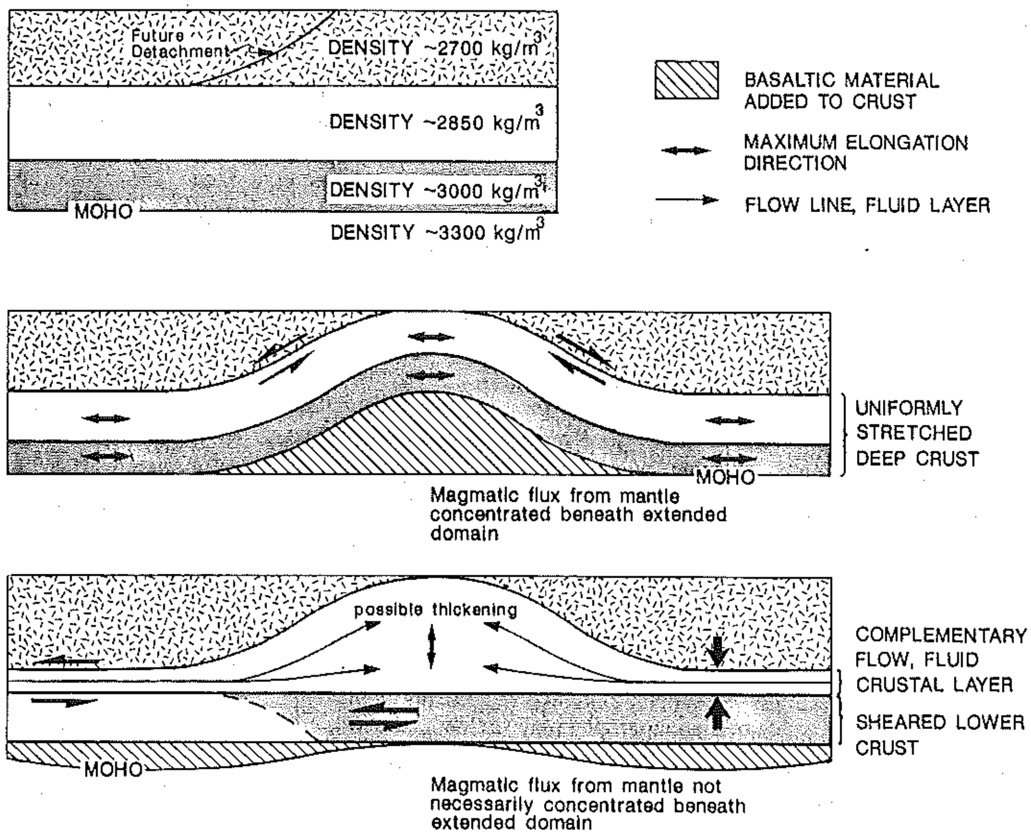


Figure 7. Contrasting modes of deep-crustal response to extension. (a) Unstrained crust. (b) Deep crust stretches uniformly, causing displacement of stretched layer out from beneath stable blocks and into extended domains. Mantle magmatic flux is concentrated beneath extended domain (possibly mixing with and remobilizing old deep crust), thereby smoothing the Moho. (c) Deep crust is divided into an upper, quartzose, fluid layer and a lower, mafic, relatively viscous layer. Fluid layer flow is the primary mechanism for keeping relief on the Moho low, and is governed in large part by flotational equilibrium of upper crustal stable blocks as they separate. Thus, fluid layer may thicken substantially beneath extended domains. Mantle magmatic flux need bear no particular relation to the position of the extended domain or stable blocks on either side, but flotational flow may serve as a mechanism to concentrate either magmatic additions or remelted crust into the extended domains relative to stable blocks via simple shear. Situation in (c) may typify early extensional history of thickened crust (e.g., northern Basin and Range), while the latest stages of extension prior to sea-floor spreading may be more similar to the situation in (b) (e.g., the Salton trough).

shows two broad depressions of 500 to 800 m centered on the strongly extended domains. Similar patterns are apparent across many of the other domains (Wernicke, *1989,* 1990; cf. *Thompson and McCarthy, 1986, 1990*). Stable blocks tend to stand higher and have broad negative Bouguer anomalies across them. However, stable blocks that have distal boundaries on either side, such as the Confusion Range block, tend to lie near topographic lows (Pl. 8; Wernicke, 1990).

Deep seismic-reflection profiling of the Basin and Range and the Omineca extended belt shows a prominent set of reflections at the Moho (Klemperer and others, 1986; Potter and others, 1986), which is locally corroborated by combined reflection-refraction imaging in the northern Basin and Range province (*Catchings and others, 1988*; *Thompson and others, 1990*). In addition, studies in the Basin and Range and other extended terrains suggest that the Moho lies at or near an abrupt downward cutoff of pervasive strong reflections characterizing much of the lower crust. This cutoff typically occurs at nearly constant depth over large regions. Throughout most of the Basin and Range, this depth is about 30 ± 3 km (Klemperer and others, 1986). The reflective lower crust contrasts with the upper crust and upper mantle, which tend to be transparent, except for discrete panels of dipping reflections. Such a pattern appears to typify the extended crust around the British Isles (*Matthews and Cheadle, 1986*; *Cheadle and others, 1987*) and the Basin and Range, except prominent upper mantle reflections have not been imaged in the Basin and Range (Allmendinger and others, 1987).

The subhorizontal Moho has been imaged across several of the boundaries between extended domains and stable blocks, in particular on either side of the Snake Range domain (Fig. 4b; Klemperer and others, 1986), across the northwestern boundary of the Whipple domain (*Hauser and others, 1987b*), and across most of the domain boundaries in the Omineca extended belt (Potter and others, 1986). Further, in some of the domains, the strongly reflective character of the lower crust is observed in the upper crust, coincident with regions of core complex tectonites, implicating extensional tectonism as the cause of the reflectivity (e.g., *Hurich and others, 1985*; Frost and others, 1987; McCarthy and Thompson, 1988; *Valasek and others, 1989*; *Goodwin and Thompson, 1988*). In the western part of the northern Basin and Range, where strongly extended domains are apparently scarce and core complex tectonites absent, the upper crust lacks pervasive reflections (Allmendinger and others, 1987).

LITHOSPHERE-SCALE KINEMATICS AND DYNAMIC IMPLICATIONS

The observations discussed in the preceding two sections are synthesized below in the context of a large-scale model for lithospheric extension in the U.S. Cordillera. Previous large-scale models have emphasized three main effects. One is the consequences of adding magma to the crust during extension, and redistributing crustal mass via magmatic transport, since extension is often broadly coeval with magmatic activity (e.g., *Mackin, 1960*; *Thompson, 1960*). A second effect is rheological layering of the lithosphere, where it is presumed that discrete brittle faulting in the upper crust is accommodated by more uniform stretching of a ductile substratum (e.g., Stewart, 1971; Eaton, *1979,* 1982; Smith and Bruhn, 1984). A third is that since core complex detachments have large offsets, the simple shear on these faults may be relayed laterally through the lower crust and lithosphere, such that thinning of various lithospheric layers may be strongly heterogeneous in any given vertical column (e.g., *Wernicke, 1981*; Davis and others, 1986). Few if any workers regard magmatism, uniform stretching, and simple shear of the deep crust as mutually exclusive processes, as they may occur in varying proportions in the history of Cordilleran extension. However, the interdependence of these processes, if any, and their implications for physical models of extension, are topics of much current debate. In this section, the debate is illuminated in the context of a recent model, advocated below, that invokes fluid behavior of the deep crust during extension. This model, while not yet generally accepted, is intended to serve as a synthetic device and focal point for discussing the merits and weaknesses of previous models.

The fluid crustal layer

The principal problem arising from synthesis of surface and deep-crustal observations is reconciling the heterogeneity of upper crustal strain with the apparent uniformity of deep-crustal structure across domain boundaries. If the deep-crustal structure of an extended domain continues beneath an adjacent stable block, then flow of deep crust out from underneath the stable block and into the extended domain is indicated (e.g., Wernicke, *1983,* 1985; Spencer and Reynolds, *1984,* 1989; Gans, 1987; Block and Royden, *1988,* 1990). This interpretation has been used from time to time to argue for the importance of magmatism, uniform stretching, and simple shear of the deep crust in extensional models (e.g., Okaya and Thompson, 1986; Gans, 1987; and Wernicke, 1985, respectively).

Existing data permit some or all of these concepts relating extension in the upper crust to deeper levels to be invoked in varying proportions. However, the observations of (1) gentle regional topographic sags (ca. a few hundred meters) and gravity highs broadly coincident with most of the extended domains, and (2) a low-relief Moho transecting strong gradients in upper crustal thinning, independently suggest an important dynamic control on the development of extension, which may reconcile these diverse concepts and observations with considerable ease: flotation of the stable blocks within a layer of intracrustal fluid, at least during the early phases of extension. This concept in essence applies the "tilted buoyant block" model for classical basin-range structure (e.g., J. K. Sales, *in* Stewart, 1978) to the strongly extended domains (e.g., Block and Royden, *1988,* 1990; *Buck, 1988*; *Kruse and others, 1989, 1991*; Wernicke, *1989,* 1990).

In the extended domains, the removal of at least the upper 10 km of crust has resulted in sags in topography in the range of 0 to 0.1 times that amount (Pl. 8). A survey of the amount of denudation in the extended domains versus the amount of topographic depression suggests a typical proportion of about 0.05, although some domains show little or no depression (Wernicke, 1990). Barring unlikely complications addressed elsewhere (e.g., Block and Royden, 1990; Wernicke, 1990; *Kruse and others, 1991*), this implies that the density of the compensating fluid lies within approximately 150 kg/m^3 of the density of the upper crust. In areas where topographic depression is minimal or absent, the medium would be nearly equal or perhaps even slightly greater than average upper crust (e.g., in the Snake Range and Ruby Mountains domains, Pl. 8; Block and Royden, 1990). Even in areas where the topographic depression is relatively large (e.g., the Death Valley and Lake Mead domains, Pl. 8), the average density of the compensating fluid would be less than 2850 to 2900 kg/m^3, and could therefore only lie within the crust. The compensating medium would be expected to contain a relatively small proportion of quartz-poor rocks thought to characterize the lower crust (e.g., mafic diorite or gabbro).

The lack of Moho relief across domain boundaries independently supports the concept of intracrustal isostasy (e.g., Block and Royden, 1990). If the separating stable blocks float in a sea of middle crust too weak to maintain shear stresses on geologic timescales, then the development of strong, upper crustal gradients in strain would not exert differential vertical loads on layers beneath the fluid (by the definition of a fluid), and they should remain undeflected unless affected from below. The crux of the argument is that the existence of a fluid layer (not simply a weak or ductile layer) requires complementarity of vertical strain between the fluid and upper crustal layers. That is, the combined thickness of the fluid layer plus upper crust would be areally constant at any given time during extension (assuming the bottom of the fluid layer is relatively flat), no matter how complex or heterogeneous the three-dimensional strain field of the upper crust might be.

Intracrustal isostasy is compatible with, but not required by, laboratory experiments on the strengths of rocks likely to exist in the deep crust. The rheology of the continental lithosphere is thought to be characterized by weak, ductile, lower crust and relatively strong upper crust and upper mantle (e.g., *Brace and Kohlstedt, 1980*), often likened to a "jelly sandwich" following *Matthews and Cheadle (1986)*. The jelly sandwich analogy is apt for a broad range of possible lithospheric strength profiles in which the lower crust is weak and ductile, but is improperly scaled with reference to the fluid layer concept, because bread does not float in jelly. If we separate two pieces of flat-lying bread resting on jelly, the horizontal pressure difference between bread-loaded and unloaded jelly is insufficient to drive flow into the void between the bread as it is pulled apart. In other words, the viscosity of the jelly is large in proportion to the imposed loads and the timescale at which even the most patient experimentalist would separate the bread, so the jelly remains far out of fluid equilibrium. A proper culinary analogy could only be drawn with substances that are impractically runny for sandwiches.

The fluid layer concept, if correct, has major implications for the roles of magmatism, ductility, and simple shear between lithospheric layers (Fig. 7). It shares the notion of rheological layering as a control of strain localization (e.g., Eaton, *1979*, 1982; *Matthews and Cheadle, 1986*), contrary to the suggestion that the lower crust may extend on inclined shear zones collinear with upper crustal detachments and upper mantle shear zones (e.g., *Wernicke, 1986*), at least during the early stages of extension. It is compatible with the concept that detachments are crustal shear zones that cut downward across the brittle-ductile transition of quartzose rocks at shallow angle (e.g., Wernicke, *1981*, 1985; Davis and others, 1986). The concept of whole-lithosphere normal simple shear (e.g., Wernicke, 1985; *Lister and others, 1986*) is also compatible with the fluid layer concept kinematically, because discrete upper mantle (e.g. *Klemperer, 1988*) and perhaps lowermost crustal shear may be transferred to the upper crust in the fluid layer. However, during the early stages of extension, discrete shear zones developed above and below the fluid layer would not be dynamically coupled in any simple way, and could develop independently of one another (e.g. *Matthews and Cheadle, 1986*; *Reston, 1990*).

Within the crust, the fluid layer model specifies (1) the relation of rheological layering to core-complex detachments, (2) a degree of weakness of the layer sufficient to cause fluid behavior on geologic timescales, (3) flow complementary to upper crustal thinning within the layer as the primary mechanism for keeping Moho relief low, and (4) deep-crustal behavior as a function of its probable strength and composition, limiting somewhat the participation of mafic rocks in the compensation process (Fig. 7b).

Earlier models stressing the importance of rheological layering and magmatism as controls on crustal strain (e.g., Eaton, *1979*, 1982; Miller and others, 1983) interpret core-complex detachments as exposures of a regional decoupling horizon between brittle deformation above and ductile deformation below, across which displacement is a small fraction of total extension, accommodating small variations in strain between the layers. For geological reasons elaborated below, it is improbable that core-complex detachments represent the interface between the fluid and nonfluid parts of the crust. Rather, they more likely represent shear zones that transect the upper crust above the fluid layer, crossing the brittle-ductile transition at low-angle (e.g., Wernicke, *1981*, 1985), although perhaps locally exhuming the fluid layer.

To the extent that mechanical thinning of a subdetachment ductile crust is uniform across domain boundaries, relief on the pre-extension Moho is equivalent to differences in upper crustal thinning between extended domains and stable blocks (Fig. 7b). Thus, if the ductile crust extends uniformly beneath strong horizontal gradients in upper crustal thinning, addition of magma from the mantle in greater proportion beneath the extended domains is required to maintain constant depth to Moho (Fig. 7b; e.g., Okaya and Thompson, 1986; *Thompson and McCarthy, 1986*; Gans, 1987). While such an end-member model is consistent with low strength, a layer that maintains constant thickness during unloading would also maintain a substantial lateral-pressure gradient and remain out of fluid equilibrium (Fig. 7c). If the ductile layer were to contain a mafic layer at its base, then the model would also require substantial relief to develop on a mafic-nonmafic interface within the uniformly stretched layer (Fig. 7b).

Thus, while broadly similar, these earlier concepts differ somewhat with the fluid-layer model in: (1) interpretation of the significance of detachments as localizing at the brittle-ductile transition; (2) the implied exclusion of fluid behavior for the deep crust; (3) the solution to the differential thinning problem primarily with a subcrustal magmatic flux beneath extended domains rather than complementary vertical strain of a fluid layer; and (4) the assumption of pure shear of the subdetachment lithosphere rather than complementary flow of a siliceous layer decoupled from relatively viscous mafic and ultramafic layers below. Such flow includes the possibility of major vertical thickening of the fluid layer in precisely those areas where upper crustal extension is greatest, quite in contrast to the kinematics of uniform thinning of deep crust (Fig. 7; Wernicke, 1990).

Role of magmatism. Redistribution of crustal mass via heterogeneous flow of a flotational medium for upper crustal blocks is a simpler mechanism than systematic concentration of the subcrustal magmatic flux in explaining the deep crustal mass budget (Fig. 7), although the two processes may act in concert. The complexities of mantle heat flux, decompression melting (in both crust and mantle), melt extraction, migration, mixing with old crust, and emplacement would all have to be incorporated into a physical model that explained the mass budget by magmatic additions alone, or by magmatic additions superimposed on a uniformly stretched deep crust.

Once derived, the prospects for a purely magmatic model (e.g., of the type proposed by *McKenzie, 1984*, and its derivatives) to keep crustal thickness uniform during heterogeneous upper crustal extension seem remote. Additions of mantle magma to the crust would as likely create relief on the Moho as smooth it, especially given the sensitivity of the amount of melt generated to parameters such as upper mantle temperature (*White and McKenzie, 1989*), which would probably not correlate in any simple way with local variations in upper crustal thinning. Because a fluid layer would eliminate differential loads on the upper mantle beneath it, its existence would preclude the possibility of decompression melting occurring solely beneath the extended domains, i.e., the mantle would have no way of knowing where overlying extended domains are. Magmatic additions are clearly an important part of the crustal mass budget during extension, and they undoubtedly profoundly modifiy the entire lithospheric column (Anderson, 1989), but collusion between the asthenosphere and upper crust to inflate the crust with precisely the amount of magma needed to simulate the simple physics of intracrustal flotation during the early stages of extension seems improbable.

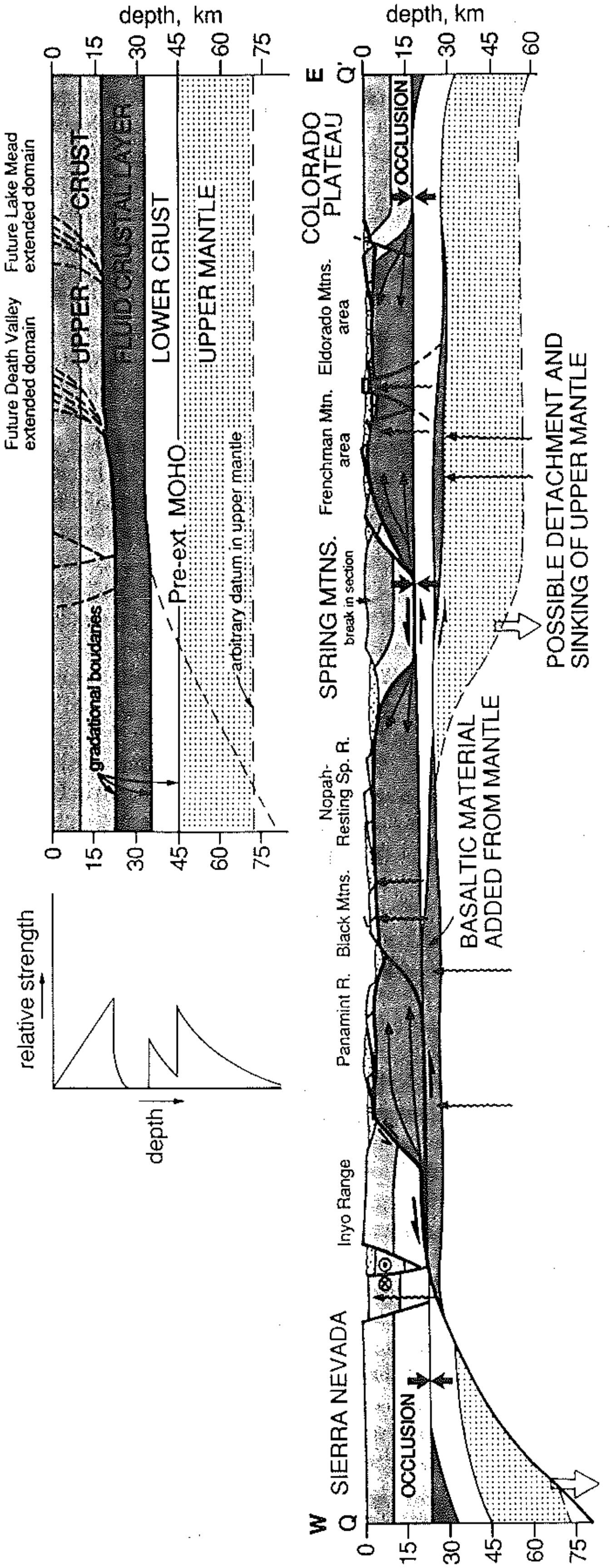
Syn-rift magma centers were emplaced over broad regions of the Cordillera at approximately the same time as extension, indiscriminately across stable blocks and extended domains (Anderson, 1989). This also suggests that the mantle magmatic flux is not concentrated preferentially beneath extended domains, and thus it is unlikely to control their localization, as advocated in some models (e.g., *Gans and others, 1989*). The degree to which mantle-derived magma underlies strongly extended domains in preference to stable blocks may be governed by flotationally driven, lateral transport of magma entrained in the fluid layer, rather than by variations in subcrustal flux (Fig. 7). The fluid layer may be rich in synrift magma, whose rheology and latent heat may substantially contribute to weakening the layer over broad regions. In this sense, magmatism may be a primary control on its development. Determining whether a magmatic flux is required to create a fluid layer, or whether a fluid layer might generally be present in continental lithosphere in the absence of a magmatic flux from the mantle, is a relevant and challenging problem in lithospheric physics.

Application. The fluid-layer concept, and some of its consequences, seem best explored by application to the central Basin and Range because of the large amounts of extension in the strongly extended domains and their abrupt boundaries with stable blocks (section Q–Q′, Pl. 8 and Fig. 8; Wernicke and others, 1988), although it should apply equally well to most of the other domains. The model combines knowledge of upper crustal fault kinematics (Wernicke and others, 1988; *Snow and Wernicke, 1989*) with the fluid-layer concept. In addition, it is assumed that the lowermost crust, if present (mafic, density 2,900 to 3,100 kg/m^3), does not participate to a major extent in the intracrustal flow, but instead accommodates shear between the upper and fluid crustal layers and the mantle. New material is shown to be added to the crust in layers near the Moho, with maximum thickness beneath volcanic regions known from the surface geology. The strain pattern in the upper mantle is shown as being breached along a localized shear zone, in a fashion suggested by the pattern of discrete inclined reflections imaged in the upper mantle beneath the extended terrain around the British Isles (e.g., the Flannan reflections; *Matthews and Cheadle, 1986*; *Reston, 1990*), although such mantle reflections have not been recorded anywhere in the Basin and Range province. The upper crustal blocks are shown sliding westward off of a rigid upper mantle breached beneath the Sierra Nevada, based on geophysical and geochemical studies in the region (Jones, 1987; *Farmer and others, 1989*), although a variety of other kinematic models involving penetrative stretching, detachment, and sinking of mantle lithosphere are tenable.

The presence of a middle crustal layer weak enough to behave as a fluid on geologic timescales is enhanced by a thickened crust and a high geotherm, as may have existed in most of the Cordillera immediately prior to major extension. Based on experimental data on the strengths of rocks, wet quartzose crust would be substantially weaker than a relatively dry, quartz-free, basal crustal layer (Fig. 8a), perhaps inhibiting the participation of the lowermost crust in regional flow relative to the middle part of the crust (summaries of Smith and Bruhn, 1984; *Kirby and Kronenberg, 1987*; and *Kuznir and Park, 1987*).

Core complex tectonites and the fluid layer. The fluid-layer concept raises the question of whether the tectonites of the metamorphic core complexes belong to the layer, and if so, whether they may be used to test the kinematics shown in Figure 8. As shown in Figure 8b, the fluid layer is not exposed along the line of section, always buried beneath the sequentially detached blocks or alluvial fill accumulated between them. Following Zuber and others (1986), the boundary between a fluid layer and overlying crust would probably form below the brittle-ductile transition within a ductile zone where strength decreases downward (Fig. 9; e.g., *Brace and Kohlstedt, 1980*). It would thus neither represent the brittle-ductile transition nor a detachment fault, possibly lying at substantially greater depth than either of these two features. Most of the core complexes may be variably rotated fragments that originally lay above the fluid layer, such that the tectonites would mainly record simple shear related to denudation (Fig. 9). This accounts for the observation that in some core complexes, penetrative ductile strain appears to diminish structurally downward away from the detachment faults (*Compton and others, 1977*; *Reynolds and Rehrig, 1980*; G. H. Davis, 1980; *Malavieille, 1987*). In this case, core-complex footwalls would represent a layer of relatively strong mid-crustal rock, rotated and flexed above unexposed fluid beneath it.

One of the best-studied core complexes both geologically and geophysically, the Whipple Mountains in the Whipple extended domain (Davis and Lister, 1988; Pl. 8) exposes a good candidate for the boundary between the fluid layer and the upper crust. In the footwall of the Whipple Mountains detachment, crystalline rocks largely undeformed during extension give way structurally downward across a sharp boundary (not a detach-



ment) to strongly deformed mylonitic gneisses. The boundary (mylonitic front of Davis and Lister, 1988) is inclined toward the proximal boundary of the extended domain, and is aligned with a band of dipping seismic reflections that flatten out at mid-crustal levels beneath the proximal block, apparently continuing beneath it at regional scale (Frost and others, 1987; Davis and Lister, 1988; *Flueh and Okaya, 1989*). The mylonitic front thus appears to represent a surface exposure of a boundary between regional ductile flow below and relatively rigid crust above (Frost and others, 1987; Gans, 1987), possibly representing the top of the fluid layer where the ductile crust strengthens rapidly upward (Fig. 9). This situation may be geometrically analogous to the west sides of the Spring Mountains and Colorado Plateau blocks (Fig. 8b), except that conditions of erosion and denudation in the Whipples were favorable to yield a glimpse of the boundary.

The majority of kinematic indicators in the Whipple mylonites show the same sense of shear as on the younger detachment that cuts them (Davis and others, 1986). This would appear to be opposite to that predicted by the fluid-layer model (Fig. 8b), in which flow out from beneath a proximal block would impart a sense of shear on the top of the layer opposite to that of the detachment. However, as shown in Figure 9, an upper boundary to the fluid layer initially situated several tens of kilometers away from the breakaway zone (beneath the flat segment of the fault at points A and B, Fig. 9) would exhibit the same sense of shear as the detachment, because the fluid layer initially flows toward the breakaway zone (Fig. 9a). These kinematics should continue until the wave of uplift reaches the to-be-exhumed mylonite front, whereupon the sense of shear might briefly reverse. Given that a small percentage of the shear indicators in the Whipple mylonites are opposite to that on the detachment (Davis and Lister, 1988), a test of the model shown in Figure 9 would be to determine if the oppositely directed indicators are younger than those sympathetic with the detachment.

The pre-mylonitic fabrics present in some of the core complexes mentioned earlier (Ruby Mountains domain, *Wright and Snoke, 1986*; Snake Range domain, Lee and others, 1987) might be an expression of the fluid layer. As discussed below, the kinematics of the fluid layer might in general contrast sharply with those of uplift and exposure via tectonic denudation.

Figure 8. Application of fluid layer concept to the central Basin and Range, with upper crustal kinematics after Wernicke and others (1988). Line of section shown on Plate 8. Datum in upper mantle is not intended to be an isotherm or mechanical boundary. Circle with "X" denotes motion of block away from viewer; circle with dot denotes motion toward the viewer. Complementary flow within fluid layer allows upper crustal blocks to separate with minimal internal fragmentation, keeping thickness of upper crustal plus fluid layers nearly constant. Occlusion of stable blocks with lower crust promotes the development of high-angle faults within both extended domains and stable blocks, and the eruption of mafic lavas previously restricted to zones of mixing in the lower crust and fluid layer. See text for discussion. Lithosphere shown to fail by large-scale simple shear, but may be accompanied by detachment and sinking of the mantle part of the lithosphere.

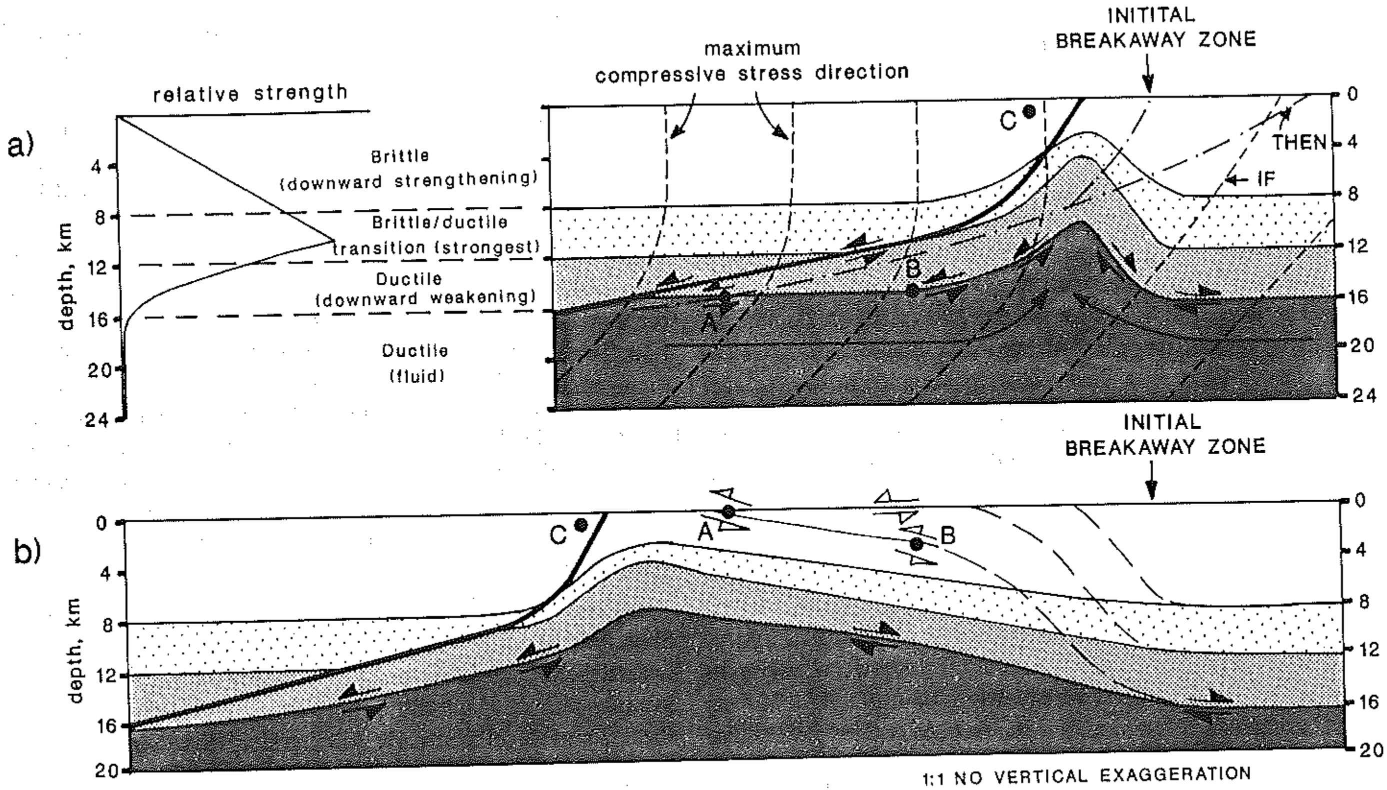


Figure 9. Cross section showing possible relations between detachments and various rheological layers. (a) Initial position of detachment shown as dot-dash line; active position shown as heavy line. Detachment flattens across strongest portion of crust, such that its initial dip through the ductile crust is low, due to rotated stress trajectories (dashed lines; e.g., *Yin, 1989*; Melosh, 1990). In some instances, inclined stress trajectories can be present in the brittle layer (marked IF), resulting in low initiation angles at shallow levels (marked THEN). Initial movement on the detachment forces fluid flow toward the breakaway zone (solid arrows at top of fluid layer), resulting in top-to-the-left shear at points A and B. Detachment is not the brittle-ductile transition, but a zone of shear that cuts downward across it at low angle (e.g., *Wernicke, 1981*). Neither is the top of the fluid layer the brittle-ductile transition, but rather the lower limit of ductile crust strong enough to remain out of fluid equilibrium. Perturbation of fluid layer, which can be magma rich at the onset of extension, might trigger volcanism soon after extension begins. (b) With continued movement, translating hanging wall and reference point C about 40 km to the left, uplifted ductile and fluid layers cool, forming new brittle crust. Long dashed lines show positions of rheologic boundaries at the onset of extension. Model predicts fossil flow direction across ductile/fluid boundary (arrows at points A and B) in the same sense as shear on the exhumed detachment. Drawing intended to schematically reproduce observations in the Whipple domain, such that point A is equivalent to the surface trace of the Whipple mylonite front (see text for discussion).

Scale and direction of flow. An interesting problem posed by the fluid layer concept is the scale at which flow might occur. At what lengthscale would the fluid layer be influenced by the separation of two stable blocks? Figure 8 is drawn to suggest that middle crust under the Sierra Nevada and Colorado Plateau is involved in flow to a limited extent near their edges, but this is unconstrained. The crust beneath the central Colorado Plateau is 10 to 20 km thicker than in the Basin and Range (e.g., *Hauser and Lundy, 1989*), perhaps indicating that the scale is limited to a range of 100 to 200 km.

However, the reflective crust apparently connected to the Whipple mylonite tectonites appears to be developed province-wide in the southern Basin and Range (Frost and others, 1987), with a tendency to rise toward the surface near core complexes (McCarthy and Thompson, 1988). In addition, the western part of the northern Basin and Range does not seem to be greatly extended at the surface, yet has crustal thickness equal to the eastern side of the province, raising the possibility that most of the crustal thinning in the western part of the province occurred by massive flow (100s of km) of the middle crust laterally out of the region. The west side may have functioned as a large stable block, which developed classical Basin and Range structure only once its fluid layer had been hydraulically expelled into the strongly extended domains to the east. Seismic-reflection profiles across the northern Basin and Range show pervasive subhorizontal reflections in the deep crust over a broad region of basin-range blocks on the western side of the province, with a relatively transparent upper crust (Allmendinger and others, 1987). These reflections appear to rise eastward into the extended domains, perhaps supportive of this concept. Alternatively, the crust may have originally been thinner on the west side (Coney and Harms, 1984).

The pattern of fluid flow need not be two-dimensional, especially given the complicated map-view distribution of extension (Pl. 8). Flow in a direction perpendicular to the regional extension direction may result across transverse boundaries between strongly extended domains and stable blocks, or could be directed radially toward extended domains if they are equant. Thus, although Figure 8 roughly balances the area of fluid in cross section, in regions such as the central Basin and Range where extended domains end abruptly along strike, cross-sectional area need not be conserved.

Implications for necking instability models. Physical models have been developed to account for the two wavelengths of variation in topography and gravity (about 30 and 175 km) discussed above. These models relate them to necking instabilities in stretched viscous or plastic layers of the lithosphere (*Fletcher and Hallett, 1983*; *Froidevaux, 1986*; Zuber and others, 1986; *Ricard and Froidevaux, 1986*). Extension of strong, upper crustal and upper mantle layers separated by a weak lower crust may result in two wavelengths of boudinage, a short one corresponding to the upper crust and a long one corresponding to the upper mantle, which in turn may correspond to the scale of individual ranges and tilt domains, respectively (Zuber and others, 1986). The origin of the spacing of basin-range blocks likely corresponds to these physics. However, the model shown in Figure 8 ascribes the long-wavelength anomalies to lateral density contrasts within the upper crust and fluid layer, maintaining a flat lower boundary on the fluid layer. The necking instability models simulate only a small amount of initial extensional strain, and, as such, an initial long-wavelength disruption of the upper mantle would somehow have to influence the development of upper crustal strain.

Occlusion of the fluid layer. The model in Figure 8 predicts that at some point, stable blocks in the upper crust and mafic lower crust will occlude, eliminating the fluid layer. The fluid layer would then occupy only the volume between the separating stable blocks, having been squeezed out from beneath them. The fluid layer probably cools during extension (Block and Royden, 1990), assuming upward heat transport due to magmatism is at some point overcome by the effect of advecting the fluid layer toward the surface (Block and Royden, 1990). Thus, its history as a fluid, particularly in its upper portions near the breakaway zone, may be relatively brief once extension begins (Fig. 9). Occlusion and freezing modify the overall rheology of the crust, and thus might be expected to result in a change in structural style once the fluid layer is eliminated.

The most obvious potential effect of elimination of the layer, alluded to above for the origin of the western part of the northern Basin and Range, is that once a stable block is beached atop the relatively strong lower crust and upper mantle, it may tend to fragment internally into large fault blocks. Indeed, the domainal character of extension and the narrow zone of initial breakage between the stable blocks (Fig. 8a) may be due to the smooth sailing (minimal basal traction) enjoyed by the stable blocks early in extension, relative to the increased basal traction on them once the fluid layer is occluded. This effect provides a simple explanation for the development of late high-angle faults observed in many stable blocks (akin to the downward-stepping brittle-ductile transition concept of Eaton, 1982), and is perhaps a control on the downstepping of major detachments observed in some of the domains (Figs. 4 and 7). In the case of the Death Valley region, this would include the late-stage fragmentation of the Inyo Range area (which lies in a stable block) and the active faulting in Death Valley proper (which lies in an extended domain; Fig. 8b).

Post-occlusion tectonics. Once the stage shown in Figure 8b is reached, and depending on whether sufficient driving force is available to continue extension, the entire lithosphere may fail in a manner analogous to the upper crust, perhaps along localized shear zones developed prior to occlusion that link together in some way to begin shearing the asthenosphere toward the surface. In this mode, isostatic compensation would occur in the upper mantle, such that deep topographic depressions may form with relatively little additional strain. Such a style of extension, accompanied by magmatic additions to the crust from the mantle concentrated along the rift zone, may typify the final stages of rifting prior to sea-floor spreading, as in the Salton trough and Gulf of California across the western extreme of the southern Basin and Range (*Fuis and others, 1984*; Lachenbruch and others, 1985).

The pre-existence of a fluid layer may be a criterion that distinguishes wide rifts that retain a substantial thickness of continental crust despite large separations of their bounding blocks (e.g., the Basin and Range) from narrow rifts that evolve quickly into ocean basins (e.g., the Red Sea). The fluid layer spreads extensional strain over a broad area, thereby inhibiting localized upwelling of asthenosphere to shallow levels.

Composition of magmas. Another observation bearing on the fluid-layer concept is the progression from predominantly intermediate to silicic magmatism characteristic of the early phases of extensional magmatism to the basaltic or bimodal eruptive sequences typical of the later phases (e.g., Anderson, 1971; Wernicke and others, 1987; *Gans and others, 1989*). An overall cooling of the crust during extension should enhance brittle deformation, perhaps increasing the likelihood that mafic magmas ponded in the lower crust may be erupted at the surface (e.g., *de Voogd and others, 1988*; *Gans and others, 1989*). In other words, the existence of the fluid layer may inhibit the ascent of mafic magma in the upper crust, perhaps because it is more difficult to breach with the brittle fissures presumably necessary to allow the ascent of mafic magma.

LARGE-SCALE TEMPORAL EVOLUTION OF CENOZOIC EXTENSION

Patterns in timing of Cenozoic extension are complex within strongly extended domains and areas that are only moderately extended, and are highly diachronous on the scale of the western U.S. Cordillera as a whole. The timing of extension in the strongly extended domains broadly follows the pattern of major intermediate to silicic magmatism that swept through the Cordillera in Cenozoic time (Coney, 1980; Lipman, 1980), but many of the details of the timing of maximum extension within a given area and its relation to volcanic activity are unresolved. In many of the domains, strong extension was accompanied by little or no magmatic activity, while conversely many of the most important magmatic centers lie in areas that are well removed from synvolcanic extension. Nonetheless, a broad correlation between magmatism and extension is apparent.

The overall pattern of Cenozoic extension, including its timing and vergence (direction of movement of detachment hanging walls), has been considerably refined over the last decade (e.g., *Reynolds, 1979*; *Chapin, 1979*; *G. A. Davis, 1980*; Coney, 1980; *Anderson, 1981*; Zoback and others, 1981; Chamberlin, 1983; Glazner and Bartley, 1984; Wust, 1986; Wernicke and others, 1987, 1988; Reynolds and others, 1988; *Baars and others, 1988*; Dokka, 1989; Taylor and others, 1989; *Janecke, 1989*; Anderson, 1989; *Gans and others, 1989*; Keller and others, 1990). The discussion below is derived principally from these studies and references therein.

Extension is strongly diachronous at large scale (Fig. 10), spanning most of Cenozoic time. A number of the extended domains appear to follow a four-stage evolutionary history, despite the diachronism of extension on a Cordillera-wide scale (Wernicke and others, 1987). These stages include: (1) the formation of early intermontane basins, probably signaling the onset of at least limited extension; (2) the eruption of predominantly intermediate to silicic volcanic rocks; (3) large-magnitude crustal extension, occurring during or immediately after second-stage magmatism; and (4) basaltic or bimodal volcanism, accompanied by lesser amounts of extension. The period of major extension within a given strongly extended domain may last as much as 10 to 15 m.y. or more (Fig. 6), although because extension has a tendency to migrate within a given domain, the period of tilting of an individual, steeply tilted block is often brief, on the order of 1 to 3 m.y. or less (Fig. 6).

The earliest extension began in the Pacific Northwest in the Omineca extended belt and along the western margin of the Rocky Mountains Basin and Range about 50 to 55 Ma, and was complete by about 35 to 40 Ma (Fig. 10a). There is currently no discernible migratory pattern in either magmatism or extension. Extension verges away from a central stable block in the Omineca extended belt, apparently transformed southeastward into the Rocky Mountains Basin and Range along the Lewis and Clark line. South of the line, vergence flip-flops along strike of the belt, going from southeastward in the north to northwestward or west-southwestward in the south (Fig. 10a).

South of the Columbia Plateaus, major extension apparently did not begin until after 40 Ma (Fig. 10b). Magmatic belts in both the northern and southern Basin and Range were active at their northern and southern extremes, respectively, by early Oligocene time, and migrated with crude symmetry toward one another along the long axis of the province, arriving at the central Basin and Range in middle Miocene time. Major extension in the northern Basin and Range was restricted to an elongate north-trending belt in the middle of the subprovince during this interval, while magmatism migrated southward in an east-west–trending wave that spanned nearly the entire width of the province. The strongly extended domains, as in the Omineca belt, show vergence directed away from a central block, with the most intense strain apparently alternating systematically from one side of the central stable block to the other (Fig. 10b), perhaps a large-strain analog to flip-flopping rift segments separated by overlapping high-relief accommodation zones documented in East Africa (Pl. 8 and Fig. 2; Rosendahl, 1987). There is no clear pattern of propagation of extension from north to south, although future studies might reveal one, as understanding of the magnitude and timing of extension improves. By comparison, the southern Basin and Range appears to show a pattern of northward propagation of extension, but has a more complicated pattern of vergence than the northern Basin and Range (Pl. 8). The extension directions within each of the two regions are generally parallel during this interval, but are northwesterly in the northern Basin and Range and northeasterly in the southern Basin and Range, forming a radial pattern about the convex-west margin of the Colorado Plateau. Modest extension affected the entire Rio Grande rift in late Oligocene time (ca. 28 Ma), and was locally severe.

Once extension and magmatism began to affect the central Basin and Range in middle Miocene time (ca. 17 Ma), profound changes in extensional architecture took place. Extension directions and least-principal-stress orientations were both southwest-northeast throughout the province by this time. Afterwards, while modest extension continued within the older belts, the locus of strongly extended domains apparently shifted to a position along the frontal portion of the Cordilleran fold and thrust belt (Figs. 1 and 10c). In addition, probable strong extension affected the Rio Grande rift and Yerington domains, with variable vergence. While major extension in the Socorro region of the Rio Grande rift occurred prior to 16 Ma (especially between about 28 and 20 Ma), a substantial amount of extension there and elsewhere in the rift is younger. Despite similarities in their pre-middle Miocene evolution, the northern Basin and Range developed into a broad region of large-scale tilted fault blocks accommodating moderate amounts of extension, while the southern Basin and Range, although somewhat faulted, was far less active. The Rocky Mountains Basin and Range became active (although perhaps in part prior to 16 Ma), accommodating moderate extension.

During late Miocene time in the Basin and Range, the over-

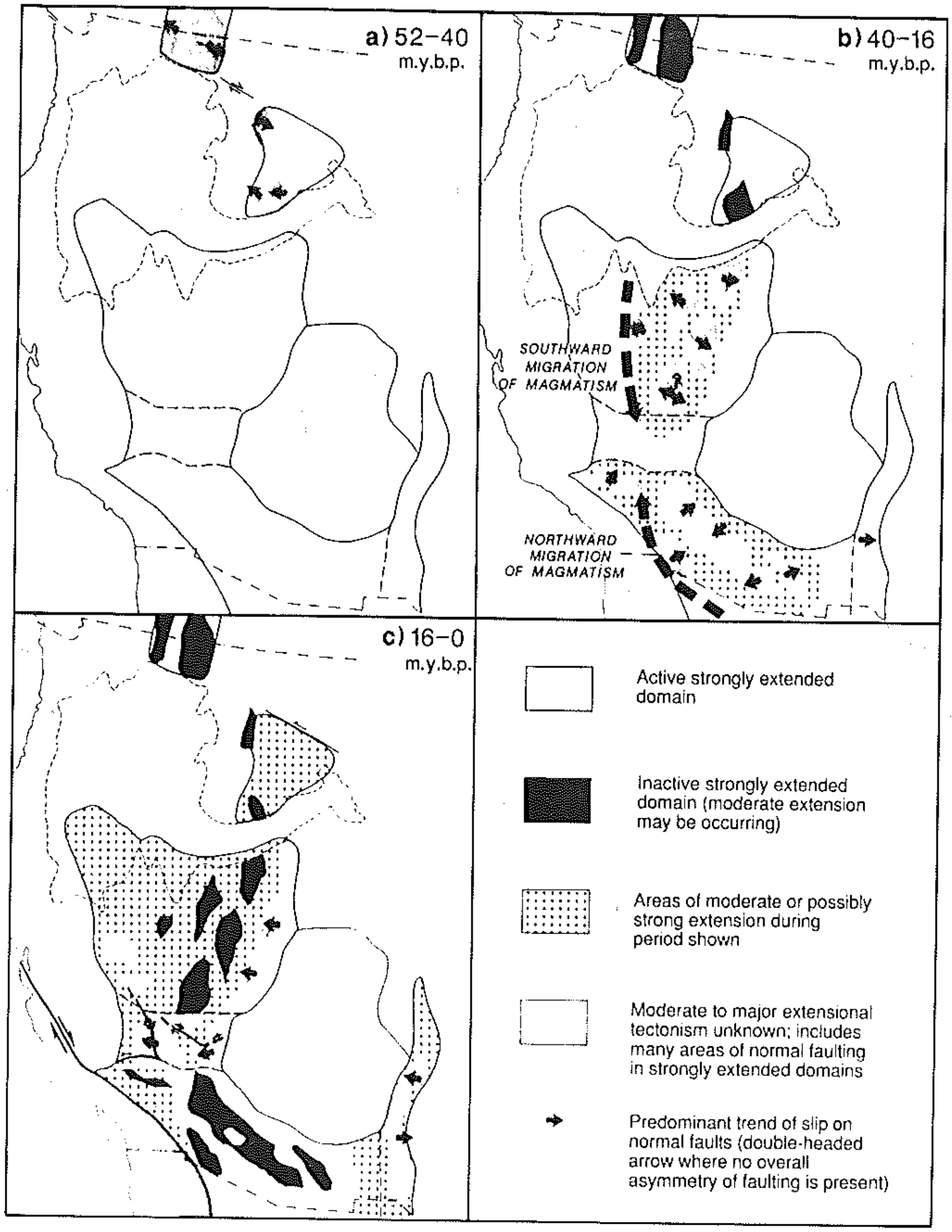


Figure 10. Evolution of Cordilleran extension in time and space. Extensional province boundaries from Figure 1 are included for reference. Heavy arrows in (b) show migration direction of mid-Tertiary magmatism. Note outward movement of major extension between (b) and (c). See text for discussion.

all extension and least-principal-stress directions appear to have rotated clockwise, changing from northeast to northwest. For example, vergence in the older eastern part of the central Basin and Range is predominantly southwestward, while in the younger western part it is mainly northwestward (Fig. 10c). This shift is also apparent in the moderately extended portions of the northern Basin and Range. The shift coincides with, and may be genetically related to the growth of the Pacific–North America transform plate boundary, including the development of the San Andreas fault (Figs. 1 and 10; Pl. 8; Atwater, 1970).

Overall, the locus of extensional strain appears to have migrated from the center of the extended provinces to their periphery as extension proceeded, albeit crudely (Fig. 10c). Most of the extension in the southern Basin and Range was transferred to the opening of the Salton trough–Gulf of California rift in the west, and to the Rio Grande rift in the east. In the north, while broadly distributed, the locus of major extension shifted from the center of the province to the margins, forming a U-shaped belt around the older domains (Fig. 10c). Around the Columbia Plateaus and northward, the development of the Neogene Rocky Mountains Basin and Range and the extended portions of the Columbia Plateaus (which could include areas due west of the Idaho batholith in the central region of the Plateaus) suggests a similar pattern of outward migration of the locus of extension. The eruption of basaltic and bimodal volcanic suites in the Columbia Plateaus and northern Basin and Range appear to have followed a similar overall pattern in the Neogene, tending to concentrate on the edges of the provinces over the last few million years (Christiansen and McKee, 1978). The scarcity of significant post-Eocene normal faulting in the Omineca belt is noteworthy, perhaps a reflection of its position far to the north of the San Andreas fault.

The central Basin and Range is one area where extension did not migrate outward, but migrated westward across the entire width of the Basin and Range in post-middle Miocene time. It seems to have represented a stronghold between westward-migrating batholithic blocks and the interior of the North American plate, failing only once the blocks had already begun to drift at other latitudes. Once failure occurred, however, it developed into the most strongly extended subprovince of the Basin and Range, accommodating about 250 km of extension across a current width of 350 km.

DISCUSSION

Origin of forces causing extension

Forces responsible for spreading may include three, broad, nonexclusive classes: (1) body forces within the lithosphere that cause it to spread under its own weight (*Molnar and Chen, 1983*; Coney and Harms, 1984; *Coney, 1987*; Sonder and others, 1987; Wernicke and others, 1987); (2) tractions applied to the edges of the North American plate, for example as a result of changing relative plate motions (Atwater, 1970; Coney, 1978, *1987*; *Engebretson and others, 1984*); and (3) tractions applied to the base of the lithosphere by flow in the underlying asthenosphere (e.g., *Scholz and others, 1971*; *Elston, 1984*; *Humphreys, 1987*). There is no consensus among Cordilleran geologists regarding the causes of extension, but certain aspects of distribution and direction of spreading with time appear to bear on the importance of each of these classes of force at different times.

Forces originating within the North American plate. The importance of body foces in the early history of extension is suggested by the concentration of extension in central areas where post-compressional crustal thickness would be expected to be maximum (Coney and Harms, 1984) and by the variability of extension directions, which tend to be perpendicular to the boundaries of shortened areas (Pl. 8 and Fig. 10). Further support for the importance of gravitational effects, which also explains the diachronism of the onset of extension at various latitudes, is an apparent correspondence between the temperature of the upper mantle immediately following compression (implied by the degree of plutonism) and the time lag between compression and extension. At lithosphere scale, the geotherm should decrease upon thickening because material advection is faster than thermal reequilibration (*England and Richardson, 1977*). This effect, combined with the increased thickness of relatively weak crustal rocks, may cause the strength of the lithosphere to decrease markedly with time as thermal reequilibration occurs (*Glazner and Bartley, 1985*). Thus, in areas where maximum thickening coincided most closely with the Cretaceous batholiths (Omineca extended belt and the Rocky Mountains Basin and Range), extensional collapse occurred soon after compression because the lithosphere was initially weak. In areas where thickening lay east of the batholith belt and where the crust was only mildly intruded (northern and southern Basin and Range), the delay was longer because more time was needed to warm and weaken the lithosphere, either by conductive relaxation or upwelling of arc magmas. Areas where thickening was accompanied by relatively little plutonism would require the most time to weaken sufficiently (central Basin and Range). Heat carried by upwelling asthenosphere, perhaps related to either subduction of plates offshore or detachment and sinking of the mantle part of the North American plate, might also influence the strength and gravitational potential of the lithosphere, contributing to the effect of body forces (Sonder and others, 1987; Anderson, 1989).

Another group of observations possibly bearing on the importance of body forces is the position of the constrictional strain field in the central Basin and Range with respect to variations in regional topography and width of the province (Wernicke and others, 1988). This region lies where the northern Basin and Range narrows from about 800 km to less than 400 km wide at the latitude of Las Vegas (Fig. 1). The batholithic western margin of the Basin and Range was not rotated about a vertical axis to any significant degree during extension (*Frei, 1986*). Therefore, the percent extensional strain (and therefore net decrease in gravitational potential of the lithosphere; Sonder and others, 1987) strongly increases southward, assuming crustal thickness was approximately the same prior to extension. The abrupt south-facing monoclinal step in topography observed along the northern boundary of the central Basin and Range separates the two subprovinces, whose average elevations differ by about 1,000 m (Eaton and others, 1978; Fig. 1 and Pl. 8). The step is a direct indication of a contrast in potential. The strong component of constrictional strain in the central Basin and Range (and perhaps portions of the Walker Lane belt) may be explained by the tendency of material in the north to flow down the potential gradient, thereby causing north-south shortening during extension.

Forces applied to the edge of the North American plate. The principal plate boundary effects during Paleogene time that are likely to have influenced the development of extension include: (1) a slowing of relative plate convergence, (2) decreasing age of the subducted plate (e.g., *Engebretson and others, 1984*; Stock and Molnar, 1988), and (3) increasing dip of a nearly horizontal Benioff zone inherited from Laramide time (Coney and Reynolds, 1977).

Observations of modern subduction zones show a strong correlation between the stress regime of the overriding plate and the age of the subducted oceanic lithosphere (Molnar and Atwater, 1978), such that extension predominates where old lithosphere is consumed (e.g., western Pacific) and compression where young lithosphere is consumed (e.g., eastern Pacific). The age of subducted lithosphere also roughly correlates with slab dip, which tends to be steeper in regions of old downgoing lithosphere, presumably because of its greater negative buoyancy. Convergence rate, while possibly an important factor, may be comparably fast or slow in either tectonic regime. Thus, the three observations discussed above conflict somewhat. The steepening Benioff zone and perhaps decreased convergence rates would favor the development of extension, but the decreasing age of the subducting plate would hinder it. It is paradoxical that the downgoing slab would steepen as the age of the subducted lithosphere decreased. A slowing of westward drift of North America relative to the Atlantic hotspots at this time may have had an influence in some way (Coney, 1978, *1987*).

The mid-Miocene and younger history of the Cordillera is characterized by a regional clockwise rotation in spreading direction and least-principal-stress direction within extended areas, from east-northeast to west-northwest (*Zoback and Thompson, 1978*; Zoback and others, 1981; *Angelier and others, 1985*), concurrent with the growth of the Pacific–North America transform plate boundary. The change in plate boundary configuration may have changed the orientation of traction exerted on the edge of the North American plate from being approximately normal to the continental margin to being parallel to it (e.g., Anderson, 1989). Thus, as the transform boundary grew, the orientation of diffuse spreading within the continent changed such that spreading could absorb some component of right-lateral shear (Atwater, 1970).

Another effect proposed for the evolution of plate boundary stresses points to the Mendocino triple junction (*Ingersoll, 1982*), which is unstable and requires westward spreading to occur within the North American plate as the triple junction migrates. A related concept is the effect of the disparity in age of the downgoing plate across the Mendocino transform prior to the formation of the triple junction (Glazner and Bartley, 1984). The greater age and lesser buoyancy of the downgoing plate to the north of the transform might tend to induce extensional deformation, as the pattern of extension and volcanism in the southern Basin and Range correlates reasonably well with the calculated positions of the fracture zone through mid-Tertiary time. However, the apparently symmetrical pattern of southward migration of magmatism in the northern Basin and Range at this time is not explained by this effect.

Forces exerted on the base of the North American plate. The magmatism associated with early extension is probably related to subduction (e.g., Lipman, 1980; Coney, 1980), because subduction occurred continuously offshore during Paleogene time, and there is no other expression of arc magmatism toward the trench. This arc was unusual in comparison with most modern arcs, however, in that it consisted of migrating belts of magma about 600 to 800 km long, oriented at high angle to the trench, with localized centers more than 1,000 km inland (Lipman, 1980).

These observations suggest that forces exerted on the crust from below, driven by convective flow related to subduction, may be an important factor in the development of extension. In addition, detachment and sinking of the mantle lithosphere destabilized by compression (a physical concept potentially related to the flat-slab hypothesis for Laramide time) may play an important role in generating a flux of magma from the mantle (*Houseman and others, 1981*). Some workers have stressed the importance of a flux of mantle-derived magma (e.g., *Dickinson and Snyder, 1979*), perhaps localized over the highly extended domains (*Gans and others, 1989*), as a driving mechanism for extension. The width of mid-Tertiary magmatic belts and their widespread occurrence outside areas of strong extension indicate that the thermal disturbance of the lithosphere also occurred over broad regions, seemingly independent of any control of previous tectonic features of the crust (Anderson, 1989). The localized distribution of early extension relative to these magmatic belts casts doubt on a simple relationship between forces that might be associated with mantle magmatic flux and the localization of upper crustal extension.

The observation that magmatism in the extended domains occurs either synchronously with or slightly prior to severe (but not all) extension in a given region (Wernicke and others, 1987) suggests that a partially molten deep crust existed when major extension began (Sonder and others, 1987). Because the generation of these early magmas implies a flux of mafic material from the mantle prior to that time (Anderson, 1989), such a flux is difficult to attribute to decompression melting associated with extension, as in the model of *McKenzie (1984)* and derivatives. Complexities in patterns of sedimentation, volcanism, and extension (e.g., Bartley and others, 1988; Taylor and others, 1989), and the fact that syn-rift sedimentary basins typically predate eruptive activity (Wernicke and others, 1987) indicate caution in generalizing along these lines. A magma-rich fluid layer developed prior to extension raises the possibility that, at the onset of extension, the initial upwelling of the fluid layer near the breakaway zone triggers eruptive activity within the extended domains (Fig. 9a). By this mechanism, volcanism would be expected after a modest degree of extension but before or during the most intense phases, thereby accounting for the common occurrence of pre-eruptive sedimentary basins. Syn-extensional eruptive centers may thus be viewed as a passive response to perturbations of the upper crust. This would explain why moderate-volume eruptive centers localize early in the history of some extensional systems near proximal boundaries (areas summarized in *Gans and others, 1989*), but would allow the most voluminous eruptive centers to lie near zones of major asthenospheric upwelling, independent of the details of the timing and distribution of major extensional strain (e.g., Bartley and others, 1988; Taylor and others, 1989). From a theoretical point of view, assessment of the mechanical influence of arc magmatism on extension, particularly in regard to how mantle flow might exert tractions on the base of the lithosphere, is a difficult problem (*Coney, 1987*).

Variations in the age of the downgoing plate may affect the distribution of forces applied to the base of the North American plate. As the age of the downgoing plate beneath North America progressively decreases through the Tertiary, the time that the slab remains cold after subduction decreases (Severinghaus and Atwater, 1990). Thus, the sinking of an old plate may strongly influence the distribution of flow in the upper mantle (and hence also the forces applied to the base of the lithosphere) because it remains relatively rigid as it sinks, while the subduction of very young oceanic lithosphere might exert less influence on upper mantle flow because the plate is short-lived.

Frontiers

One of the principal problems in studying the extended lithosphere in the Cordillera and elsewhere is the existence of low-angle normal faults indicated from field and seismic-reflection data, and their absence from records of seismicity in extending regions, including the Basin and Range (Jackson, 1987; Smith and others, 1989). The paradox may in part be resolved with the sequential uplift model shown in Figure 5. Here, a regional, shallowly inclined or subhorizontal shear zone may acquire its mylonitic and brittle overprint as it moves transiently up the migrating flexure, such that all activity in the brittle crust occurs along the relatively steep (generally >30°) flexure (*Wernicke and Axen, 1988*; *Buck, 1988*; *Hamilton, 1988*).

However, the evidence for shallow initial dips along the sides of a number of stable blocks cited above is not reconciled by this argument. The observation that some well-studied Basin and Range faults (such as the Borah Peak fault in the Rocky Mountains Basin and Range) project steeply into the mid-crust (e.g., Smith and others, 1989), sometimes cited as evidence against the existence of low-angle normal faults, sheds no light on the problem, as they are simply examples of large steep normal faults, not in conflict with any known concept of Cordilleran extension.

The mechanical unlikelihood of low-angle normal faults may not be a problem when viewed in light of a low-viscosity lower crust. For a variety of assumptions, flow within a viscous layer at the base of the brittle crust is capable of causing rotations of the stress field with depth, such that the initiation of low-angle normal faults, assuming coulombic behavior in the brittle crust, may be expected (e.g., *Yin, 1989*; Melosh, 1990; Fig. 9a). The strong part of the crust may serve as a "stress guide," whose differential motion with respect to lower layers creates subhorizontal shear tractions that rotate the principal stress planes (*Lister and Davis, 1989*). Figure 9a shows that a 45° rotation of the stress field with depth is sufficient to cause initiation of subhorizontal fractures. Faults with low initial dip in the uppermost crust may simply reflect a shallower depth of upward steepening of the maximum compressive stress (Fig. 9a), although special circumstances such as preexisting anisotropies may be necessary to allow the stress field to remain rotated at depths of 0 to 5 km.

Assuming the geological data indicating initiation and active slip on low-angle normal faults is correct, the problem of reconciling slip on these faults with seismicity data on normal faults is still unresolved. The earthquakes analyzed by Jackson (1987), who concluded that seismic slip on normal fault planes inclined at less than 30° is "almost unknown," involves a sampling of 40 earthquakes over the past two decades whose fault plane solutions are well constrained and unambiguously indicate primarily normal slip. Even with the ambiguity of which nodal plane represents the fault (it is resolved for 15 of the events), it is clear that few if any dip less than 30°. Although the data are only certain to within 1 to 15°, the vast majority of the events are clearly steeper, mostly in the 40 to 60° range.

Several possibilities exist to reconcile Jackson's (1987) compilation with the existence of active low-angle normal faults. One is low recurrence intervals on shallow faults, such that the small sample of events did not record them in equal frequency to events on steeper planes. Since they transect the brittle crust at low angle, they have large surface areas and are capable of absorbing proportionately more elastic strain than steep faults. Thus, probable active low-angle normal faults such as the Sevier Desert detachment, which dips at 10° to 15° across the upper 1 to 15 km of crust (Allmendinger and others, 1983), may have to 5 times the surface area of a steep basin-range fault, and have an accordingly lower recurrence interval. Thus, Jackson's (1987) sampling may not have included enough events to document an earthquake on a low-angle fault. Another possibility is that low-angle normal faults are for some reason aseismic (Jackson, 1987). It may be that steep faults are more common in the upper 7 to 1 km than shallow ones, tending to decrease in dip downward across the brittle-ductile transition zone and remain shallow across the ductile crust above the fluid layer (Fig. 9). Such geometry would favor creep along the shallowly dipping portion at 7 to 15 km depth (Jackson, 1987). Whatever the reason, the problem is sure to remain a lively topic of debate in geodynamics.

Progress in assessing the importance of the numerous proposed causes of extension is clearly a frontier area. Improved knowledge of the timing, direction and rate of extension through time, the structure of the upper mantle, and careful continuum modeling, should lead not only to major advances in evaluating the importance of each of the effects described above, but also to the discovery of processes not yet considered.

Promising along these lines are efforts to use isotope geochemistry to constrain the source regions of magma (e.g., Farmer, 1988; *Farmer and others, 1989*) and high-resolution tomographic studies of the structure of the upper mantle (e.g., *Humphreys, 1987*). In combination with knowledge of the upper crustal displacement field determined from geologic studies and shallow-reflection profiling, these techniques may ultimately constrain the three-dimensional strain field of the entire lithosphere during extension, shedding new light on long-debated questions such as the relationship between strain in the upper and lower parts of the lithosphere.

SUMMARY

The Cordillera has experienced a number of episodes of extension throughout its long history, most notably in Cenozoic time. Cenozoic extensional provinces are developed mainly within areas that were shortened during Mesozoic and early Cenozoic time (Fig. 1), although not all shortened areas have been significantly extended. Extension is best expressed east of the great Mesozoic batholiths, which were largely resistant to upper crustal extension (Figs. 1 and 10; Pl. 8), although important Cenozoic extension exists to the west of them as well. The magnitude of extension, variably oriented but roughly east-west (Pl. 8; Fig. 10), lies in the range of 200 to 300 km in the central and northern Basin and Range, possibly somewhat less in the southern Basin and Range and areas in the Pacific Northwest.

The mode of extension appears to vary with time in any given area. The early history of extension tends to be domainal and asymmetrical, characterized by the separation of stable upper crustal blocks along narrow zones of breakage (initially as little as 10 km wide or less) within which faults are inclined in a single direction (Figs. 4 and 8), broadly accompanied by intermediate to silicic magmatism. These blocks were able to separate from tens of kilometers up to as much as 150 km from one another (Death Valley domain), in effect completely removing the upper 7 to 15 km of crust from the strongly extended domains between them (Figs. 4 and 8). The fault zones (detachment systems) were initially inclined at average dips ranging from less than 10° to as much as 60° through the upper crust, probably on average shallowing downward. Hanging walls experienced multiple generations of domino-style normal faults (perhaps mostly at the scale of individual ranges; Fig. 5a), and sequential detachment and flexural rotation of thin fault blocks above a migrating wave of mid-crustal uplift (especially at large scale; Figs. 3c, 5b, and 6). Superimposed on this pattern, major detachments may sequentially step downward, such that higher detachments may be deformed by movement along imbricate splays of deeper ones in the same extended domain (Figs. 4 and 8). In many of the strongly extended domains, denudation has been sufficient to exhume metamorphic tectonites developed during extension along single systems of detachment and ductile simple shear, now exposed across tens of kilometers parallel to their transport directions (metamorphic core complexes). Not all detachments, however, display metamorphic tectonite in their footwalls (Figs. 3a and b).

The stable blocks are typically situated on broad topographic highs and Bouguer anomaly lows. Stable blocks and extended domains alternate with a wavelength of 150 to 200 km, about 3 to 5 times the width of a typical basin-range pair. The topographic difference between the extended areas and stable blocks suggests that the blocks are afloat on a fluid layer in the crust whose average density is probably between 2,800 and 2,900 kg/m^3 (Fig. 7). The existence of a fluid layer is independently supported by an apparent lack of deflection of the Moho across boundaries between strongly extended domains and stable blocks (Fig. 8). Detachment faults do not represent the boundary between the fluid and upper crustal layers, which probably occurs well within the ductile crust below the brittle-ductile transition of quartz (Fig. 9). Rather, they appear to be shear zones developed above the fluid layer that cut gently downward across the brittle-ductile transition and through a substantial thickness of crust below it, perhaps locally exhuming rocks initially within the fluid layer (Fig. 9).

The broad synchronism of early extension and magmatic activity suggests that the development and maintenance of the fluid layer may depend on magmatic heat. The occurrence of magmatic centers outside of strongly extended domains, including some of the most voluminous, suggests that mantle magmatic flux does not occur preferentially beneath strongly extended domains, and thus it is improbable that such a flux either localizes or provides the driving forces for their development in any simple way. Rather, synextensional magmatism within the extended domains may be a passive response to extension, triggered by the perturbation of a magma-rich fluid layer as the stable blocks initially separate.

The later stages of extension are characterized by breakage of both extended domains and stable blocks by broadly distributed, high-angle faults (classical Basin and Range block faulting) that accommodate relatively little extensional strain, and are often accompanied by mafic or bimodal magmatism. This structural style is best developed in the western part of the northern Basin and Range province, where the earlier mode of extension seems to be least well developed (Fig. 10c). Elsewhere in the extended Cordillera, block faulting may be less important than early large-magnitude extension in the development of the topography (e.g., the southern Basin and Range). In the Omineca extended belt, classical Basin and Range topography did not develop. Speculatively, the development of deep-seated block faulting in the Basin and Range may have resulted from the elimination of the fluid layer, either by expulsion from beneath the stable blocks (thereby increasing shear tractions on their undersides) or "freezing" of the layer squeezed upward into the extended domains, as predicted from the kinematics of the fluid-layer model (Figs. 7, 8, and 9). Physical models of lithosphere-scale necking of viscous or plastic layers of variable strength may explain the two wavelengths of observed deformational and geophysical variations.

The timing of initiation of major extension varies from about 50 to 55 Ma in the Pacific Northwest to as recently as about 16 Ma in the central Basin and Range. Following magmatic patterns, initial extension may have propagated along the axis of the Cordillera in two waves moving toward one another through the northern and southern subprovinces of the Basin and Range during Oligocene and early Miocene time, ultimately converging on the central Basin and Range in middle Miocene time. By this time, the extension direction and least-principal-stress direction throughout the Basin and Range province were northeast-southwest, in contrast to Oligo-Miocene and earlier times when extension directions may have been more variable (Fig. 10). The middle Miocene was apparently a time of major reorganization of the extending system. As extension was concentrated along a central axis of the Cordillera in pre-mid-Miocene time, it expanded to include the margins afterwards (Fig. 10). By late Miocene time, the extension direction and least-principal-stress direction rotated clockwise from northeast-southwest to their current northwest-southeast orientations. There is no one time at which a Cordillera-wide shift from domainal large-magnitude extension to classical Basin and Range block faulting occurred, because strong extension occurred at different times in different places throughout the Cenozoic (Fig. 10), including the development of metamorphic core complexes.

Proposals for the causes of extension are varied, complexly interrelated, difficult to model physically, and as a rule not mutually exclusive. An origin of early extension, in part related to buoyancy forces within the lithosphere inherited from Cretaceous and early Tertiary crustal thickening, explains a number of aspects of the development of extension. Similarly, the Miocene clockwise rotation in extension direction and stress field seems best explained as a consequence of the growth of the Pacific–North America transform plate boundary. Changes in relative plate motion in the early Tertiary and the effects of tractions exerted on the base of the lithosphere by mantle flow probably influenced the development of extension, but it remains difficult to assess their importance quantitatively.

The notion that normal faults that initiate or slip at low angles are mechanically unlikely may be erroneous, given the likelihood of strongly variable viscosity of deep-crustal layers. However, observations of normal faults that were active at low dips in the brittle crust apparently conflict with studies indicating that seismic slip on normal faults dipping less than 30° is rare. Resolution of key issues will require improved coordination between geochemistry (particularly isotope tracer work), seismic imaging (especially of the upper mantle), determination of the displacement field of the upper crust, and physical modeling.

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