Crustal extension in the Basin and Range Province, southwestern United States

W. Hamilton

SUMMARY: Cenozoic extension of areally varying ages and amounts has on average doubled the width of the Basin and Range Province. Extensional structures that formed at all depths down to 20 km, and which range in age from Oligocene to Holocene, are widely exposed and are here interpreted in terms of a model of depth-varying deformation. The middle crust is extended by discontinuous ductile shear as internally undeformed lenses slide apart along gently dipping zones of mylonite. The tops of these lenses are undulating detachment faults, the composite area of which increases with time as deep lenses slide out from underneath shallower ones. Brittle blocks of upper-crust bedrock above the detachments respond first by rotating between range-front faults, the same direction of rotation being maintained across a series of lenses, and then by pulling completely apart, while basinal strata fill the gaps and are dragged directly on detachment faults. Some faults rise gently from the main detachment zones and surface as range-front faults. Most tilted-block ranges are isolated atop detachments.

Detachment faults cut out crust. Beneath them are mid-crustal rocks of any age and type and above them are mostly upper-crustal rocks, including extensive syndeformational basin sediments rotated to steep or moderate dips. As attenuation proceeds and components rise, detachment faults evolve from ductile to brittle, develop splays, and are themselves broken by steep brittle structures related to new, deeper detachments. Parts of detachment faults remain active even after exposure at the surface, but slip on them is then limited to the down-dip direction. It is inferred from seismic reflection profiles and rock-mechanic considerations that the unexposed lower crust is extended by more pervasive ductile flattening.

This paper develops a model of depth-varing extensional styles to account for the common denominators in Cenozoic geological structures exposed by erosion and tectonic denudation throughout the Basin and Range Province. The general model was suggested earlier (Hamilton 1982), and resembles in important aspects that of Davis (1980).

Oligocene to Quaternary extension of the style discussed in this paper has affected the Great Basin part of the province (Fig. 1), and also a region N of the Snake River Plain in E-central Idaho and SW Montana. Most of the similar extension in the Mojave and Sonoran Desert sectors of the province was of Oligocene to middle-Miocene age, correlating with early extension in the N. Middle-crust extensional structures are widely exposed in both N and S sub-provinces, whereas shallow structures are seen primarily in the N one. Topographic relief is high throughout the province. Exposures are excellent in the arid SW part, and good in the semi-arid SE and N parts.

Extension has about doubled the pre-Oligocene width of the province, now 350–700 km wide. Palinspastic reconstruction of Mesozoic tectonic and magmatic belts requires such extension (Hamilton 1978). Relationships between strike-slip and extensional faults in the southern Great Basin indicate late-Cenozoic extension to be at least 65% (Wernicke *et al.* 1982). Pre-Tertiary rocks and Cenozoic basin-fill materials comprise sub-equal amounts above exposed detachment faults, requiring 100% extension in many areas. Extension has, however, been erratic in space, time, and amount.

How has the hundreds of kilometres of extension been transmitted through the continental crust? The field data needed to answer this have been greatly expanded by many geologists, particularly those working with detachment structures within the last decade, but there is still much disagreement regarding extensional mechanisms. Although Ransome, Emmons & Garrey (1910) recognized that Basin and Range faulting required crustal extension (they also mapped a detachment fault), more than 50 years passed before extension was broadly accepted over alternatives of vertical jostling or even compression and it was not until almost 1980 that extension of more than 5 or 10% was widely regarded as likely. Among those with more advanced ideas were Hunt & Mabey (1966), probably first to recognize a complete domiform fault as extensional, Hamilton & Myers (1966), who argued for 100% extension and Armstrong (1972), first to see the regional extent of middle-crust extensional structures.

From Coward, M.P., Dewey, J.F. & Hancock, P.L. (eds), 1987, *Continental Extensional Tectonics*, Geological Society Special Publication No. 28, pp. 155–176.

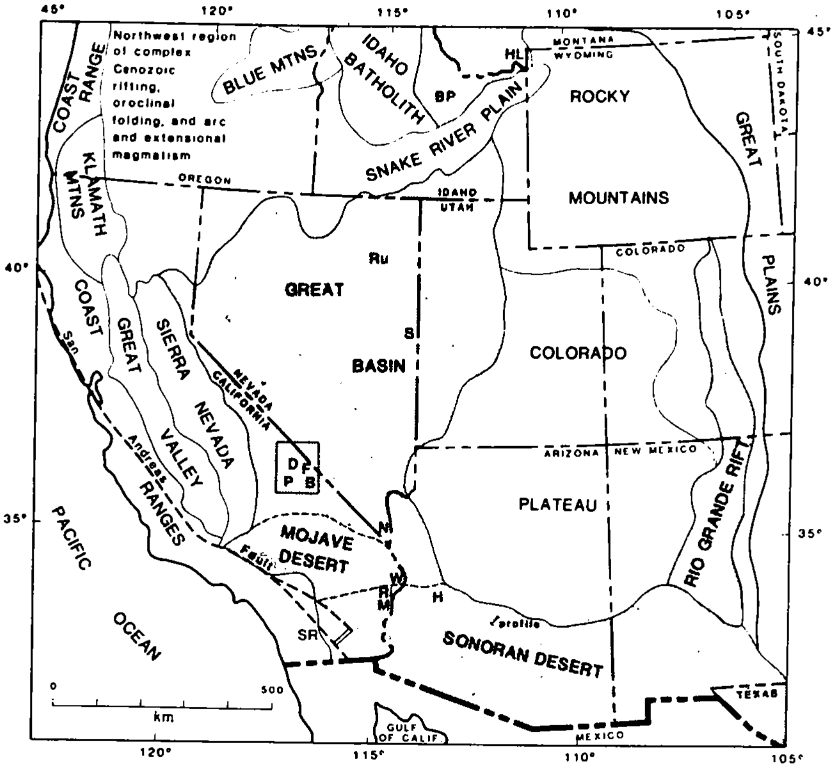
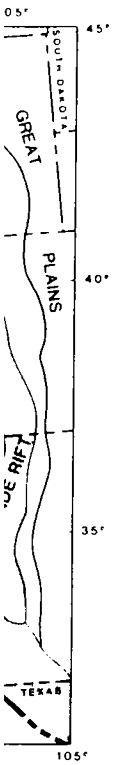


FIG. 1. Index map of part of the western United States. Basin and Range Province of extensional faulting is shaded, as are areas of similar structure in the Rio Grande rift system and in the region N of the Snake River Plain. Extensional faulting also affects much of the 'NW region', and the W and S margins of the Colorado Plateau; most boundaries shown are arbitrary. Death Valley region (Fig. 7) is located by a rectangle; reflection profile (Fig. 24) is located in Arizona. Localities noted in text and figures: B = Black Mts; BP = Borah Peak; D = Death Valley; F = Funeral Mts; H = Harcuvar Mts; HL = Hebgen Lake; M = Big Maria Mts; N = Newberry Mts; P = Panamint Mts; Ri = Riverside Mts; Ru = Ruby Mts; S = Snake Range; SR = Santa Rosa Mts; W = Whipple Mts.

Only a few of the many authors whose reports are integrated here can be cited. I also draw heavily on discussions and field trips with many geologists and on my own detailed and reconnaissance fieldwork. As the structures described here include types unfamiliar to many readers, I illustrate them with a number of photographs.

The model

The model sketched here (Fig. 2) is derived from the systematic variations of observed structures with the inferred depths and temperatures of their formation. Extension is accommodated within the brittle upper crust by rotation about sub-horizontal axes and fragmentation of large blocks that end abruptly downward against undulating regional *detachment faults* that mostly initiated at depths greater than about 10 km. 'Detachment fault' has been used in this context by Davis *et al.* (1980) and many others. 'Denudation fault' (Armstrong 1972) and 'decollement' (Miller *et al.* 1983) are used with approximate synonymity. Upper-crust bedrock blocks are bounded laterally by downward-flattening listric faults, by rotated planar faults and by basin fills deposited in widening gaps between separating blocks. The detachment faults define the tops of great lenses that are separated by gently dipping, anastomosing ductile faults and that retain pre-faulting fabrics in their interiors. As the lenses slide apart the area of their composite top, the detachment faults, is increased and the middle crust is extended by discontinuous ductile flow. Although most faults above the typically mid-crust detachments are probably listric, curving into the detachments, in



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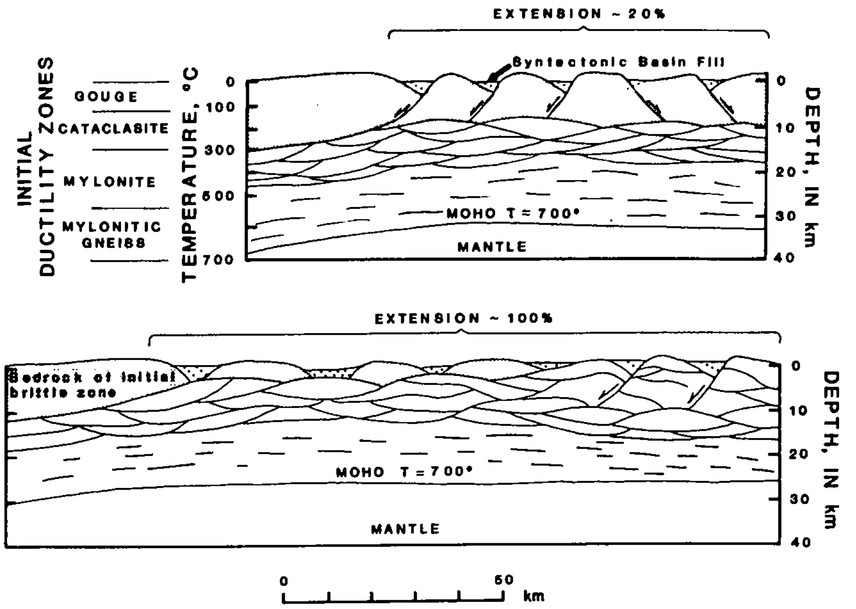

FIG. 2. Cross-sections of extending crust. Brittle upper-crust blocks rotate and separate. Middle-crust lenses slide apart along ductile shear zones; composite upper surface of lenses forms detachment faults that increase in total area with time. Lower crust flattens pervasively. Structural styles are superimposed as components rise to the surface with continuing attenuation. Attenuating crust is partly rebuilt by magmatism, and possibly also by phase change, so crust is thinned by a factor less than extension ratio. Ductility zones after Sibson (1983).

places major gently dipping faults rise for tens of kilometres laterally through upper-crust rocks and curve to steeper dips only at shallow depth.

COCORP reflection profiles (Allmendinger *et al.* 1983) across western Utah show a detachment fault of this upper-crust type, which crops out in the Canyon Range, dipping gently westward for a distance of 100 km or more in the sub-surface, to a depth of at least 13 km. In outcrop and at shallow depth, the fault separates Tertiary strata, rotated to moderate dips, from Palaeozoic strata. The fault most likely steepened into a range-front fault in its upper part, now removed by erosion. The broadly undulating Canyon Range detachment passes westward beneath the Sevier Desert, House Range, Tule Valley, Confusion Range, and Snake Valley, all of which are thus allochthonous above it, and appears to project westward beneath the thick crustal lens whose top is the Snake Range detachment fault. Discontinuous reflections beneath the Canyon Range detachment fault may in part image subdetachment ductile shear zones.

Detachment faults evolve from zones of ductile slip initially 10–20 km deep. Local variations in temperature, strain rate and lithology are reflected in variations in initial depth and configuration of the ductile faults that evolve into detachment faults. As extension progresses and the crust is thinned, most parts of the system rise toward the surface; ductility-facies boundaries rise as extension and possibly also shear heating, outpaces thermal conduction, but in general superimposed structures and fabrics indicate progressions from deformation under deep, hot conditions to deformation under shallow, cold conditions. Ductile faults become semi-ductile, then brittle and splays split away. Still further extension can no longer be accommodated on undulating faults, which then are broken by steep, brittle faults, related to new ductile faults at depth, but limited slip continues on detachment faults even after they are exposed at the surface. Extension in the lower crust is accomplished by more pervasive ductile flattening.

Middle-crust crystalline rocks beneath detachment faults are widely termed core complexes. I do not use the term, for it often incorporates the implication that detachments are necessarily related to local thermal anomalies which is not the case. Some detachments may indeed be related to synextensional magmatism, but most sub-detachment complexes represent nearly random samples of the middle crust, albeit in a region subjected to much synextensional magmatism.

Mathematical and mechanical analysis of depth-varying crustal deformation was presented by Kligfield *et al.* (1984) and Sibson (1983).

Tectonic setting

The middle Tertiary structures of the Basin and Range Province record extension mostly W-southwestward in the S and approximately westward in the N, relative to the continental interior. This extension was perpendicular to the continental margin of the time (Zoback *et al.* 1981) and was synchronous with subduction of Pacific lithosphere beneath most of that margin (Eaton 1984). There was thus great extension of the overriding continental plate in a convergent-plate regime. A common misconception is that overriding lithosphere plates generally undergo shortening and compression; actually, most undergo extension, although not generally as extreme as in the southwestern US. Trenches retreat and roll back into subducting plates. The inclination of a subducting slab marks a transient position, not a trajectory down a fixed slot; subduction occurs at an angle steeper than that inclination and the common regime above a slab is extensional, not compressional. The mid-Tertiary extension occurred after the slowing of the previously rapid westward motion of North America (as viewed in a whole-Earth, zero-sum frame).

Extension of similar style, mostly of mid-Tertiary age, but continuing through the Neogene, affected the Rio Grande rift. Extension there can be visualized as due to the clockwise rotation of the Colorado Plateau about 3° relative to interior North America, about a Euler pole near the Colorado–Wyoming border.

Late Cenozoic extension represents oblique fragmentation of the crust in the NW direction of relative slip between the Pacific and North American Plates, due presumably to partial coupling between those plates (Atwater 1970, Eaton 1984). The structural style of extensional response is similar. There was much inheritance of mid-Tertiary structural grain from Mesozoic structures and from those by late-Cenozoic structures; structures of all of these ages typically have northerly trends in the Great Basin and northwesterly ones in the Mojave and Sonoran Deserts.

Sophisticated climatic analysis of palaeofloras by Axelrod (1985) indicates that surface altitude in the Great Basin, the sector now undergoing broad extension, has increased markedly during the late Cenozoic. This increase has been coincident both with extension and with the rise of much of western North America. Such palaeo-altitude data appear incompatible with speculation by Coney & Harms (1984), that high-standing thick crust spread gravitationally to produce extensional structures.

Magmatism

Mid-Tertiary extension was broadly coextensive in space and time with magmatism both of arc type and of more silicic type intermediate between typical arc and extensional assemblages. Volcanism of both types is now active above subducting lithosphere in the rapidly widening North Island, New Zealand. Thermal softening of the crust by magmatism probably expedited extension. The Colorado Plateau was bypassed by magmatism and left unextended, whereas the Rio Grande rift to the E was the site of extensive magmatism and severe extension. Integration of the high heat flow and widespread late-Cenozoic magmatism with the rates and amounts of extension indicates that these thermal manifestations are largely effects, not causes, of extension (Lachenbruch & Sass 1978) and are enhanced by asthenospheric convection as a by-product of the extension (cf. Steckler 1985). Late-Cenozoic magmatism has been mostly of bimodal basalt-and-rhyolite type that probably reflects the melting of lower-crustal materials by rift-related mantle magmas. Deep-crystallization products of magmas of both arc and rift types must have partly rebuilt attenuating crust.

Structures of the upper crust

The basic structural unit of the Basin and Range Province is commonly but mistakenly perceived to be a tilted bedrock block, which includes a mountain range and the basement of the adjacent basin and which rotates against the next block along a fault that either is flattened with time by rotation, or else flattens downward as a curving, listric surface and that ends as a detachment fault. Such juxtaposed blocks probably formed during an early stage of extension, but most of the province has evolved beyond this stage. The majority of ranges in the

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province represent a mature stage and are either tilted panels of resistant rocks isolated from one another atop detachment faults or else are raised sub-detachment complexes. Wernicke *et al.* (1985) came to similar conclusions. Examples from the Death Valley region are discussed below.

Major upper-plate panels typically are 70–150 km long and 10–25 km wide in the Great Basin. The width of a mature bedrock panel is markedly less than the distance between range-front faults of adjacent mountain blocks (mostly 20–35 km). Most bedrock blocks are separated from one another, the gaps being filled by syntectonic basin strata in direct contact with gently dipping faults that pass beneath the tilted blocks. Dips in basin fills increase markedly downward, both stratigraphically and along the beds, toward bounding and underlying faults (Allmendinger *et al.* 1983; Effimoff & Pinezich 1981; Smith & Bruhn 1984). Details are complex, for basin fill deposited during early extension can become the lithified bedrock of subsequent deformation. Blocks and basin fills are broken by lesser normal faults. Upper plates can be shattered into small rotated-domino subblocks (e.g. Ransome *et al.*. 1910; Miller *et al.* 1983), or they can be tilted blocks of crust 10-km thick and little deformed internally (Howard *et al.* 1982).

Upper-plate panels are tilted randomly on a provincial scale, but tend to be tilted mostly in single directions over domains of 10,000 km^2 or more (Stewart 1980, 1983b). These domain tilts may record initial coherence of blocks rotating in the same sense. Non-tilted horsts and grabens are uncommon and where present they generally separate domains of blocks of opposite tilt directions.

Young normal faults have dips mostly of 45–60° where they cut bedrock at the surface (Fig. 3). Faults refract upward to steeper dips in unconsolidated materials, in which extensional collapse features are developed (Myers & Hamilton 1964). Erosion of fast-rising scarps to slopes recording their angle of repose (Fig. 4) occurs largely by landsliding that results in the formation of extensive megabreccias along scarps (Fig. 5). In settings of rapid tilting and faulting, lacustrine strata also can be deposited almost directly against high scarps.

FIG. 3. Normal fault (arrow) between Proterozoic metadolomite (right) and slightly tilted Quaternary alluvium. N end Tucki Mountain, Panamint Mountains, California. The fault cuts only the upper plate, above the detachment fault (not in view). Exposure is 10 m high.

FIG. 4. Late Quaternary normal-fault scarp cutting middle- or upper-Miocene volcanic and tuffaceous rocks. The late-Holocene scarp, 6 m high, along base of mountain, displays structural dip, but the rest of face has been eroded back, mostly by landsliding. Miocene rocks dip steeply E (left) and are truncated at shallow depth by a detachment fault. Slide megabreccia (right foreground, blocky) is separated by broad collapse moat (low ground) from frontal fault. View SE—Artist's Drive fault, Black Mountains, California.

Rates and amounts of extension have been non-uniform in space and time. Extreme late-Miocene to Quaternary extension characterizes the Death Valley region. Most deformation more than 30 km across strike to the E of Death Valley pre-dates the late-Miocene, whereas the faulting that has defined the modern ranges across strike to the W of Death Valley is of Pliocene and Quaternary age.

Gently dipping extensional faults formed in upper-crust, low-temperature rocks in a number of places in the Great Basin and commonly truncate rotated upper plates against lower plates. The western Utah example plumbed by COCORP was discussed previously. Wernicke *et al.*

FIG. 5. Small, Holocene slide megabreccia (centre), and upper edge of large late-Pleistocene slide megabreccia (far left), derived from middle- or upper-Miocene volcanic rocks of the Artist's Drive fault. View N—upper and lower plates and the fault are all truncated downward against the Badwater detachment fault in exposures to the S.

(1985) described a domiform example from southern Nevada, similar geometrically to the more common detachment faults atop mid-crust crystalline rocks but with almost unmetamorphosed Palaeozoic strata in the footwall. Some such high-level faults appear to have reactivated faults that initially were thrusts. Howard & John (this volume) show a detachment fault that splays to shallow normal faults that bound crystalline-block ranges in headwall fashion. It is not yet clear whether such headwall blocks are anchored to the deep crust or are adrift above detachment faults with which the shallow faults merge.

Earthquakes

Most small earthquakes in the Great Basin occur shallower than 8 km, whereas large ones typically nucleate at 10–15 km. Shear strength apparently increases below 8 km, then decreases rapidly across the brittle–ductile transition, which represents a temperature of 300–350°C and is no deeper than 15 km (Sibson 1983; Smith & Bruhn 1984). A major unresolved problem is the much greater depth extent suggested for steep faults by studies of seismicity than is permitted by the continuity of detachment faults seen on reflection seismic profiles.

Tilted blocks of bedrock commonly end downward against detachment faults which pass beneath the ranges, and generally only basin fill is in contact with frontal faults. Analyses of coseismic strain, that incorporate the assumption that adjacent tilted blocks of elastic bedrock are in direct contact at sub-surface frontal faults (e.g. Stein & Barrientos 1985) are probably incorrect. The best-documented deformation accompanying a normal-fault earthquake, that at Hebgen Lake, Montana, in 1958, $Ms=7.5$, consisted largely of absolute subsidence which represented a net decrease of about 1 km^3 of crustal volume (Myers & Hamilton 1964). Coseismic subsidence also greatly excluded rise during the 1973 Borah Peak earthquake, $Ms=7.3$, in Idaho (Stein & Barrientos 1985). Pre-seismic extensional strain high in the crust apparently is stored primarily as dilatation and released mostly as compaction.

Structures initiated in the middle crust

Detachment faults

Domiform or undulating detachment faults have been recognized in perhaps 50 ranges scattered far beyond the narrow zone depicted in some relatively early reports (e.g. Rehrig & Reynolds 1980). These faults are not thrusts, for they cut out, not duplicate, crust and their upper plates display abundant extensional features. Any middle-crust rocks can lie beneath detachment faults: rocks of various lithologies e.g. mid-crust granites, gneisses and migmatites, or metasedimentary or meta-igneous rocks of any grade from lowest greenschist, even slate, to highest amphibolite, and of ages ranging from Precambrian to Miocene. Non-metamorphosed supracrustal rocks occur primarily in upper plates and commonly are rotated steeply downward against the faults with truncation angles near 50°. Granitic rocks of Phanerozoic, mostly late-Mesozoic, age within upper plates are primarily of upper-crust types.

Upper-plate rocks tend to show the same sense of rotation over large domains. The direction of relative slip along a detachment surface, as defined by extensional structures low in the upper plate, is typically in the same sense as that rotation. In a domain of about 20,000 km^2 in SE California and SW Arizona (e.g. Fig. 6; Davis *et al.* 1980; Howard & John, this volume), upper-plate blocks slipped relatively NE over sub-detachment rocks and have general SW dips. As the unfaulted Colorado Plateau lies NE of this tract and as upper-plate structure requires extreme extension, upper-plate blocks must have moved relatively southwestward away from the plateau by amounts increasing with distance from the plateau. The southwestward offset of lower-plate rocks relative to the plateau must be still greater.

The amount of slip between juxtaposed upper and lower plates is commonly difficult to determine; half the province is covered by Quaternary

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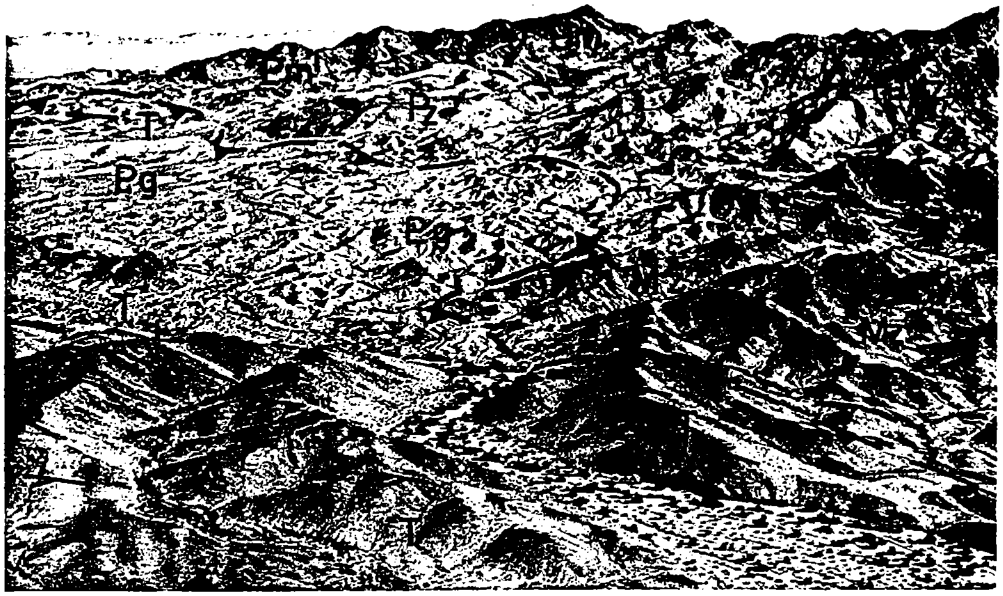

FIG. 6. Undulating Oligocene detachment fault, Riverside Mountains, SE California. Upper plate: light-coloured middle-Proterozoic granite (Ᵽg) and dark Oligocene slide megabreccias and redbeds (T; direction of steep dip indicated); broken by spoon-shaped faults. Lower plate: (Pz), Palaeozoic metasediments; (Mz), Mesozoic metasedimentary and metavolcanic rock; and (Ᵽm), retrograded and variably mylonitized middle-Proterozoic gneiss. View N—slip of upper plate relatively NE. Range mapped by author.

basin materials and too little of the other half displays optimal mixes of upper and lower plates, necessary for comparison. The outcrop overlap of 6 km of quite different crystalline assemblages in the Riverside Mountains (Fig. 6) requires an offset substantially greater than that distance; the geology of lower-plate ranges to the SW, from which relative direction the upper plate came, best fits an offset of 20–30 km. Howard *et al.* (1982) inferred 30–40 km of offset of a dyke swarm between upper and lower plates in the same tilt domain, 70 km NE across strike from the Riverside Mountains. The extreme extension within the intervening upper plate is apparently not matched by variations in slip on the detachment, so the area of the detachment surface probably increased with time. Bartley & Wernicke (1984) made a permissive argument for 60 km of slip on the Snake Range detachment, eastern Nevada. Note that in arguing for little slip on the same structure, Miller *et al.* (1983) invoked such improbabilities as regional metamorphism that ended abruptly at the level of the initial fault, and the formation of kyanite at a depth of only 6 km.

Equivalence of deep and shallow structures

Basin and Range detachment faults are regarded by many investigators (e.g. Davis 1980; Miller *et al.* 1983; Rehrig & Reynolds 1980; Reynolds 1985; Snoke *et al.* 1984) as having evolved from deep ductile structures to shallow brittle ones. Explanations diverge widely beyond this limited consensus. Some models infer that shallow deformation above the detachment faults was quite unlike that of presently active parts of the Great Basin and was typified, perhaps, by close-spaced normal faults of small individual displacements and by depositional basins larger than those of the modern system (e.g. Stewart 1983b). I infer, as do many other geologists (e.g. Wernike *et al.* 1985) that on the contrary, tilted Great Basin blocks are forming above detachments like the older ones now widely exposed and that those older structures had basin–range pairs above them.

Dimensions of basement and basin-fill assemblages above detachment faults (Fig. 7) resemble those of basin-range pairs. There is broad temporal and geographical overlap of structures

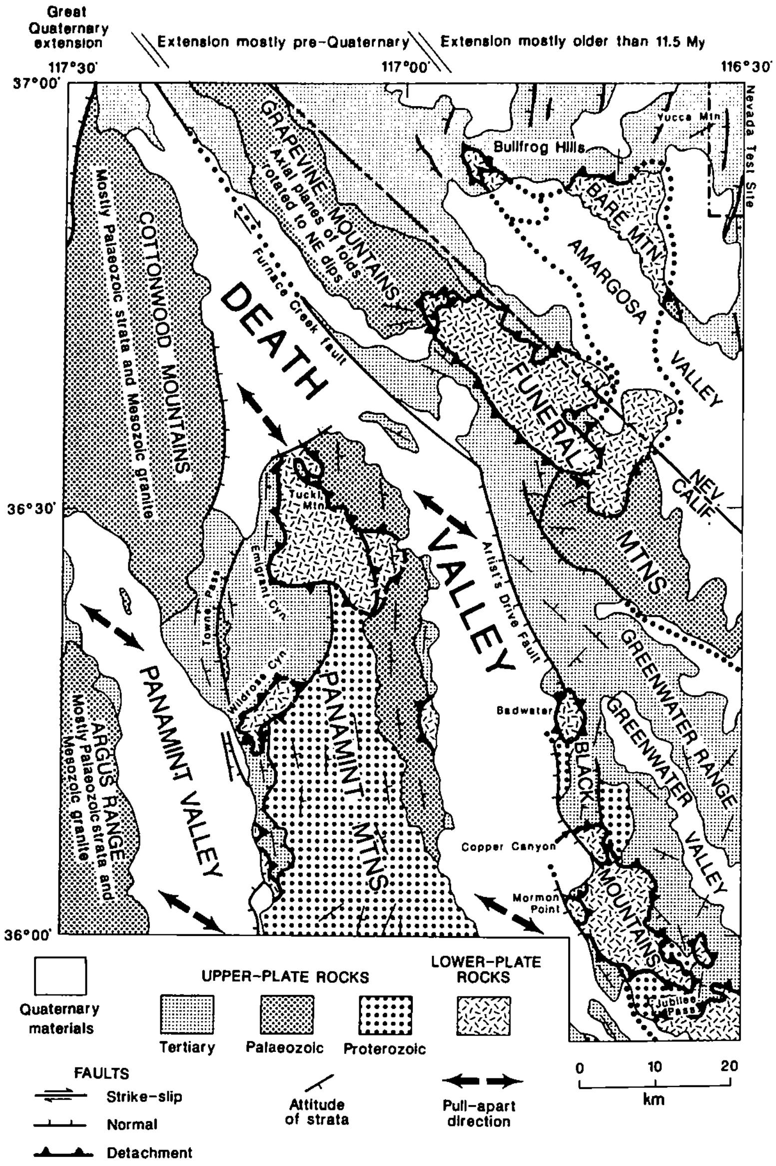

FIG. 7. Tectonic map of Death Valley region, California and Nevada. Most normal faults are omitted. Adapted from published maps, including those cited in text.

exposed as tilted blocks and as detachments. Continuing extension superimposed block structures on detachments at transition times that varied from early-Miocene to Quaternary. There are exposed detachment faults as young as Holocene, and block-bounding faults as old as early-Miocene, nearby within the southern Great Basin. Early-Miocene floras of Nevada indicate that basins and ranges then existing had a topographic relief comparable to modern basin-range pairs (Axelrod 1985). Domains of same-direction tilting of upper-plate rocks above detachment faults are of sizes, typically 10,000–20,000 km^2, similar to those of Great Basin tilted-block domains (Stewart 1980; Fig. 7; Hamilton, unpubl. work). Scarp-facies sediments are widely present in basin strata truncated downard against detachment faults. Abundant, more or less monolithological, megabreccias (Figs 8, 9 & 10) must have slid

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FIG. 8 Rotated normal fault. Stripped fault surface (smooth slope, right and foreground) on Cretaceous muscovite granite dips 30°W; light-coloured landslide megabreccias of granite and Proterozoic metasediments dip 20° E into fault. Pliocene or early Pleistocene range-front fault with 3 km of slip, rotated 25° with upper-plate block above detachment fault (not in view). Jail Canyon, Panamint Mountains; mapped by Albee *et al.* (1981).

FIG. 10. Mid-Tertiary detachment fault, SE Whipple Mountains, California. Oligocene volcanic rocks and slide megabreccias dip left (SW) to truncation against horizontal fault at top of Proterozoic (?) plutonic rocks. Mylonitic carapace shows as layering beneath fault.

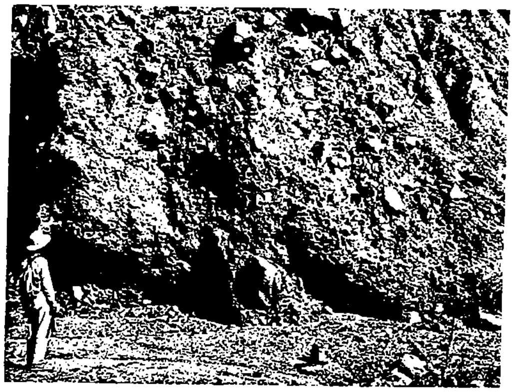

FIG. 9. Oligocene landslide megabreccia, dip 60° right (SW), 100 m above Riverside Mountains detachment fault.

FIG. 11. View N across Sacramento Pass to subhorizontal mid-Tertiary detachment fault (marked), Snake Range, Nevada. Upper plate consists of rotated blocks of unmetamorphosed Palaeozoic strata that dip to the left. Lower plate (including foreground) is of Cambrian and upper-Proterozoic strata metamorphosed to amphibolite facies (kyanite present) and intruded by muscovite granite. Light-coloured low ground is underlain by upper-plate Oligocene redbeds, which also dip moderately left above detachment fault. Area described by Miller *et al.* (1983).

from high, fast-rising scarps like those of Figs 4 & 5. Truncation angles of Cenozoic strata against faults are typically within the range 40–60°, regardless of the dips of strata and faults (Figs 3, 6, 8, 10, 11, 12, 13, 14); rocks now truncated by gently dipping faults have been rotated down from truncations against normal faults.

Rise with time

Most components in an extending system must rise toward the surface with time. Many investigators have cited evidence for the great rise of detachment faults from middle to upper crustal levels as slip on them progressed (e.g. Davis 1980, 1983; Rehrig & Reynolds 1980; Reynolds 1985; Snoke *et al.* 1984). Final major slip has always been at low temperatures and

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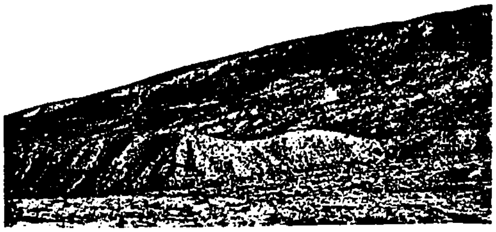

FIG. 12. Late-Neogene Tucki Mountain detachment fault. Smooth bedrock slope is stripped fault, dipping gently W, on upper Proterozoic metasedimentary rocks and Cretaceous muscovite granite. Unconsolidated Pliocene alluvium dips (lines) E to truncation against detachment. Emigrant Wash, Panamint Mountains.

shallow depths. The faults are marked by brittle gouge and incoherent breccia, typically a few decimetres (Fig. 15) to a few metres thick, but ranging from a centimetre (Fig. 16) to tens of metres. Many gouge zones grade upward into decreasingly brecciated upper-plate rocks and small-scale extensional structures are widespread low down in upper plates (Fig. 17). Severe brecciation can extend 100 m into the upper plate.

The depth and temperature of slip decrease with time (cf. Fig. 2). Progressively older and more ductile-strain rocks, in the sequence gouge–cataclasite–mylonite–mylonitic gneiss, are displayed downward from many detachment faults. A detachment-fault gouge zone is commonly bounded sharply against a shear-polished pavement (Fig. 15) on a hard, chloritic microbreccia that displays partial reconstitution at low greenschist facies. This cataclasite is typically a few metres thick and it may grade downward into a zone of sheared chloritic, epidotic rocks, tens of metres thick, Beneath these, or directly beneath the microbreccia in other cases, is rock typically recording more ductile shear. Such rock is laminated mylonite where quartzofeldspathic rocks are involved (Fig. 10) and foliated greenschist-facies rock where metasediments are present (Figs 13, 18, 19). These ductile zones typically fade out downward through thicknesses of tens or even hundreds of metres and define carapaces semi-concordant beneath the detachment faults. Mylonitic gneisses are widespread in the most-uplifted lower plates. The mylonitic carapaces and progressively lower-temperature cata-

FIG. 13. Neogene detachment fault, W side Funeral Mountains, California. Sheared carapace (layering, right) was developed concordantly below detachment in middle-grade Proterozoic metasedimentary rocks. Arrow indicates dip of upper-plate mid-Tertiary redbeds beneath late Quaternary pediment. View N from Keane Wonder Mine.

FIG. 14. Truncation of upper-plate structures against active detachment fault, Black Mountains, California. Interbedded Pliocene clastic strata, slide breccias, and volcanic rocks (lower left) dip obliquely back and right towards the Quaternary normal-fault scarp along high-standing plutonic rocks (left and centre rear). Hanging wall, footwall, and fault are all truncated downward against the active Copper Canyon turtleback (detachment) fault, which forms the smooth face dipping obliquely left through the centre of the view. Light-coloured rocks on the turtleback are a discontinuous carapace of mylonitized dolomite. An active range-front normal fault truncates both upper and lower plates at the top of the alluvial fan.

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FIG. 15. Close up of fault of Fig. 10. Hammer is on zone of breccia 0.4 m thick, above which (in shade) is a thin zone of gouge. Upper-plate Oligocene slide breccia dips steeply left. Lower plate displays several levels of shear-polished pavement in coherent cataclastic microbreccia.

FIG. 16. Mid-Miocene detachment fault, Newberry Mountains, Nevada. Proterozoic monzogranite both above and below. Upper-plate rock is altered red (in part by Precambrian (?) retrogression from granulite facies) and variably brecciated; lower-plate rock is grey and is stripped along its mylonitic carapace. Gouge zone is 1 cm thick.

FIG. 17. Small normal faults in variably brecciated Tertiary redbeds, 50 m above Riverside Mountains detachment fault.

FIG. 18. Calcite mylonite derived from marble, in carapace beneath Ruby Mountains detachment fault, Nevada. Quartzite and amphibolite form boudins and rotated blocks. Area studied by Snoke *et al.* (1984).

clasites are developed in some cases on Tertiary mid-crust granites, barely older than the detachment faults that cut them (e.g. Reynolds 1985; Snoke *et al.* 1984). Much debate centres on whether the ductile carapaces represent old zones of any origin affected latterly by extensional slip, or early slip on mid-crust zones that evolved into the shallow detachment faults. My conclusion, shared with many other investigators, is that the mylonites generally belong to the detachment systems, whereas the more regional and higher-grade mylonitic gneisses are in many cases older than extensional deformation.

Crustal level

There is typically great crustal omission across detachment faults, although large upper-plate blocks rotated down against detachment faults show much omission at their toes but little or none at their heels (Howard *et al.* 1982) Rocks exposed beneath detachment faults mostly dis-

FIG. 19. Retrograde carapace in right-central part of Fig. 13. Z folds (arrows to hinges) and small fault (offsets contact ×2 m) indicate shear of top-to-W (right) sense, same as slip of upper plate. b = boudin.

play petrological evidence for crystallization in the middle crust, whereas unmetamorphosed supracrustal rocks of any age are largely confined to upper plates of detachments younger than those rocks. Mesozoic and Tertiary granitic rocks of upper plates are mostly of upper-crustal type in cross-cutting plutons with contact-metamorphosed wall rocks, whereas correlating rocks beneath detachments include abundant mid-crust migmatites and muscovite granite. I have discussed elsewhere depth variations in granitic and migmatitic systems (Hamilton 1981). Kyanite occurs beneath detachment faults in many ranges, whereas andalusite is uncommon. Sedimentary rocks beneath detachment faults may be metamorphosed only to slate, but even this requires a temperature of 300°C and hence generally a mid-crustal depth. Quartzose mylonites, also requiring a minimum temperature of about 300°C, are widespread beneath detachment faults and uncommon in the young rocks above them. These and other indicators of temperature and pressure require a common minimum depth of erosion of 10 km for rocks beneath detachment faults and of 15 km for many of them, even where those rocks are as young as Miocene.

Rapid uplift generally accompanied extension. K-Ar and fission-track ages of rocks beneath detachments commonly approximate the age of tectonic denudation and record rapid uplift and cooling from ambient temperatures of the middle crust. The dates do not generally record mysterious 'thermal events' despite the over-use of that term in the literature. Slip on detachment faults, and hence local rates of extension, must correspondingly have reached at least 10 km 1,000,000 yr^{-1}, or 1 cm yr^{-1}. At any one time, extension was rapid in parts of the province, as in the modern Death Valley region, and much slower or inactive elsewhere. High strain rates may have resulted in the shear heating of ductile shear zones, expediting further strain.

Precambrian basement rocks were widely eroded to mid-crustal depths before the deposition of upper-Proterozoic and younger strata and so occur in both upper and lower plates. Precambrian basement rocks of upper plates, however, tend to be variably altered, brecciated, and oxidized (Fig. 16) whereas those of lower plates generally lack such low-grade alteration and show instead variable carapace retrogression to mineral assemblages of lower greenschist facies and of higher grades at greater depths.

Wernicke (1985) postulated that detachment faults are segments of ramps that cut through the entire crust. The prediction that depths of origin of terrains, juxtaposed by correlating detachments, vary systematically across-strike on a regional scale is not fulfilled, although local sectors of several ranges each, may show such progressions (e.g. Howard & John, this volume). The superb imaging by Klemperer (Klemperer *et al.* 1986) of the Mohorovicic discontinuity across northern Nevada shows the seismic base of the crust to be continuous and gently undulating, unbroken by ramps as postulated by Wernicke (1985). Upper-crust ramps do, however, apparently connect between mid-crust lenses and the surface, as in the Utah COCORP example.

Davis (1983) inferred that gently dipping detachment faults develop by domino rotation of blocks between faults of initially steep dips. The lack of unmetamorphosed basin fills and other upper-crust rocks beneath most detachment faults is evidence against this view.

Late-Cenozoic detachments of the Death Valley region

Most of the detachment-fault literature has described mid-Tertiary complexes and this has led to the erroneous impression that structures of that age are different from younger ones. Accordingly, I emphasize here the detachment faults of late-Neogene age in the Death Valley region (Fig. 7), which is within the most actively extending part of the Great Basin. Here are the highest steep scarps, the greatest deformation of Quaternary materials, the deepest topographic closures of structural depressions and the youngest exposed detachment faults of late-

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Miocene to Holocene age. Unresistant Neogene upper-plate materials are widely preserved above the detachments and late-Quaternary valley fill hides much less structure than in less active regions. Mica K–Ar ages of 10–15 Ma for sub-detachment rocks in several ranges presumably date the early stage of vigorous tectonic denudation here as of late-middle-Miocene and early-late-Miocene age.

The domiform Tucki Mountain detachment fault in the Panamint Mountains has flank dips of 10–30° and a broad crest (Figs 20 & 12; Hunt & Mabey 1966). The fault is sub-parallel to the layering in lower-plate rocks, which are upper-Proterozoic metasedimentary rocks and Cretaceous muscovite granites. A retrograde, mylonitic carapace on these materials parallels the domiform fault, which is marked by gouge and incoherent breccia. The upper plate consists of unmetamorphosed Palaeozoic strata in the E, Pliocene (?) slide megabreccias on the crest and Pliocene fluvial strata, slide megabreccias, and basalt to the W, all steeply to moderately E-dipping to truncations downward against the detachment fault and broken by many W-dipping normal faults that end against the detachment.

I found in my own reconnaissance another small detachment dome in the Wildrose Canyon area, where E-dipping Pliocene strata and slide megabreccias are rotated down-to-the-SE against it. Detachment faults, with megabreccias in the upper plate, are exposed in several places to the S along the W base of the Panamint Range, where Albee *et al.* (1981) in part mapped them as 'slide surfaces'. Detachment faults are exposed also at the E foot of the range (the 'Amargosa thrust fault' of Hunt & Mabey 1966), where the lower plate is of mid-Proterozoic augen gneiss intruded by abundant Tertiary dykes. K–Ar biotite ages of 11–14 Ma in the gneiss (Stern *et al.* 1966) presumably reflect rapid Miocene uplift by tectonic denudation. I infer from the alteration of the Tertiary dykes, that they were injected while the wall rocks were still at ambient mid-crust temperatures.

The central and southern Panamint Mountains form a great tipped crustal section, 10–12 km thick and dipping about 20° E, entirely allochthonous above these detachment faults that are exposed to the W, N, and E, as I interpret the reports of Albee *et al.* (1981), Hunt & Mabey (1966), and Labotka *et al.* (1980). The down-section progression of the overlying crustal block begins with upper-Miocene volcanics (Cemen *et al.* 1985) and unmetamorphosed Palaeozoic strata and continues W through increasingly metamorphosed Palaeozoic and Proterozoic rocks, first andalusitic and then sillimanitic (Labotka *et al.* 1980). A shallow, cross-cutting Miocene granite occurs in the middle of the section and a Cretaceous muscovite granite occurs at the same low crustal level as that of sub-detachment rocks. Normal faults dipping gently W through the E-dipping crustal section (Labotka *et al.* 1980), presumably merge with the underlying detachment (cf. Fig. 20). The range-front fault, that bounded this upper-plate has been tilted with the block and with the flanking hanging wall slide breccias (Fig. 5). The palaeo-range-front fault is exposed all along the W slope of the Panamint Mountains (Albee *et al.* 1981), has a slip of about 3 km (Labotka *et al.* 1980) and projects obliquely up the Tucki dome to its crest (Fig. 20). The hanging wall breccias are truncated downward against an exposed detachment fault. The whole assemblage has been raised along young normal faults at the W and N edges of the range, although even the modern faults appear to offset upper-plate

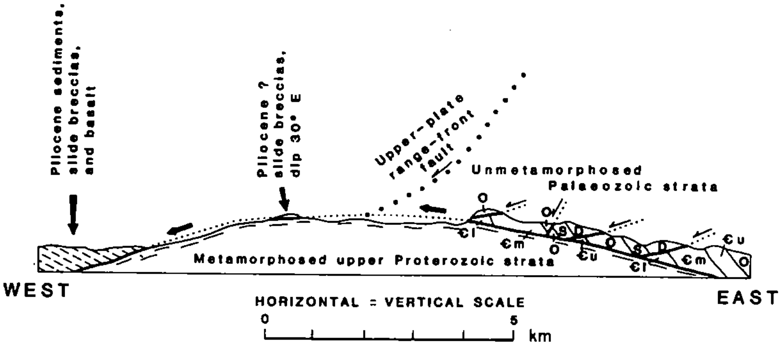

FIG. 20. Section across Tucki Mountain detachment fault, Panamint Mountains, California. Sense of slip of upper plate is the same in both flanks of dome. Palaeozoic strata: Є1, Єm and Єu = lower, middle and upper Cambrian; O = Ordovician; S = Silurian; D = Devonian. Adapted from Hunt & Mabey (1966, their fig. 86).

materials more than lower-plate ones and hence may until recently have ended at the same detachment fault.

What I regard as a domiform detachment fault in the Funeral Mountains was mapped as many faults of diverse types by Hunt & Mabey (1966) McAllister (1971), B.W. Troxel and L.A. Wright (in Labotka 1980), and Giaramita (1984, who recognized its detachment character). Rocks inside the continuous elliptical trace of the detachment, so inferred, are upper-Proterozoic strata and possibly also older-Proterozoic basement rocks, metamorphosed during the Mesozoic to greenschist to uppermost amphibolite facies and intruded by muscovite granite. Kyanite occurs in rocks of all grades, and metamorphism occurred at a uniform depth of about 20 km (Labotka 1980, pers. comm. 1986). Upper-plate rocks are unmetamorphosed uppermost Proterozoic to Miocene strata, the latter including slide megabreccias rotated to steep to moderate E and SE dips and truncated against the detachment (Fig. 13). The Neogene stratigraphy is summarized by Cemen *et al.* (1985), who do not make a detachment interpretation. A greenschist-facies carapace on the lower plate displays isoclinal folding in the same shear sense as the slip direction of upper-plate rocks (Fig. 19) and opposite to vergence in unretrograded deeper lower-plate rocks (Fig. 21). Giaramita (1984) similarly recognized that W-verging folding and shearing (progressing with time from ductile and mylonitic to brittle) affected high lower-plate rocks at the N end of the dome synchronous with Tertiary detachment and denudation.

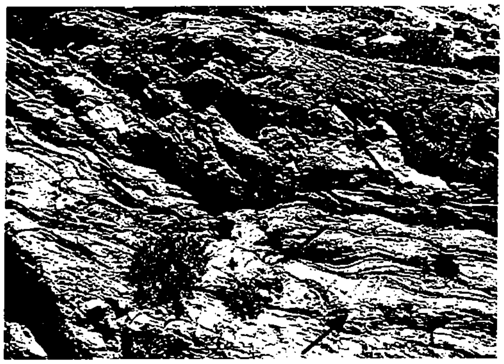

FIG. 21. Lower amphibolite facies Proterozoic metacarbonates, below carapace near area of Fig. 19. S folds (arrows to hinges) indicate shear of top-up-to-E (left) sense: Cretaceous synmetamorphic shearing had sense opposite to that in Neogene carapace. Exposure is 4m high.

Bare Mountain displays a detachment fault (The Tertiary thrust fault of Cornwall & Kleinhampl 1961; M.D. Carr & S.A. Monsen, unpubl. map; P.P. Orkild pers. comm., 1985) representing either another culmination on an undulating Funeral Mountains fault, or possibly the top of a separate mid-crustal lens. Miocene volcanic and clastic strata as young as about 12 Ma dip moderately to steeply E to their truncation at the gently dipping Bare Mountain fault atop Palaeozoic and upper-Proterozoic strata. These are metamorphosed to greenschist to lower amphibolite facies and are cut by gently dipping Tertiary faults. Biotite K–Ar ages of crystalline rocks beneath the Funeral Mountains and Bare Mountain detachments scatter from 10 to 50 Ma (M.D. Carr, pers. comm. 1985), so variations in synextensional and pre-extensional uplift and erosion are indicated. A composite detachment fault system exposed nearby in the Bullfrog Hills consists of an upper sheet of unmetamorphosed Palaeozoic strata intervening between downward-truncated Miocene rocks above and variably mylonitized gneisses beneath (Florian Maldonado, pers. comm. 1985).

The Black Mountains expose undulating detachment faults, of late-Neogene age, at two structural levels, as I interpret mapping and descriptions by: Drewes (1963—he recognized little extension); Otton (1977—summarized by Wright *et al.* 1974—who recognized great extension); and Wright & Troxel (1984—they inferred mostly imbricate listric normal faults where I identify undulating detachments). My inferences have been shaped also by my own reconnaissance and by discussions with B.W. Troxel. The structurally lower part consists of three small domiform masses, the NW-trending Badwater, Copper Canyon (Fig. 14), and Mormon Point turtlebacks, that are disconnected at present erosion levels and are separated by detachment faults from overlying Cenozoic materials in the W and plutonic rocks in the E. These masses consist of variably mylonitized mid-Proterozoic plutonic rocks capped by discontinuous carapaces of mylonitized and extremely attenuated upper-Proterozoic metasedimentary rocks (Otton 1977). The domiform turtlebacks might represent either giant mullions on a continuous detachment surface, or separate lenses above deeper faults. Drewes (1963) presented a lens-structure interpretation of a different sort. Above the turtleback detachments, to the E is a thick sheet forming a structurally higher lens of diorite and monzogranite of Mesozoic and (?) Tertiary age. Bounding this sheet on the top is the upper detachment fault, marked by breccias and lenses of upper- and lower-plate rocks in a

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zone as thick as 100 m. Above this zone, or elsewhere truncated sharply against the detachment, are variously mid-Proterozoic plutonic rocks, unmetamorphosed upper-Proterozoic and Cambrian sedimentary rocks and, mostly, mid-Miocene to Quaternary volcanic and sedimentary materials. The Cenozoic sections include extensive intercalated slide megabreccias (most 'chaos' and much 'Palaeozoic sedimentary rock' of Drewes 1963, and most 'megabreccia' and many 'fault slices' of Wright & Troxel 1984), from which I infer that high scarps were present, and hence that extension was very active, while they were being deposited. Around the southern Black Mountains, upper-Miocene rocks generally dip more steeply than do Pliocene rocks, although both are truncated by the same undulating faults, and mid-Quaternary strata are rotated to gentler dips in the same direction (Wright & Troxel 1984), whereas in the N, steep-dipping Pliocene strata are semi-concordant to upper-Miocene strata (Cemen *et al.* 1985; Hunt & Mabey 1966). Rotation synchronous with persistent faulting is indicated.

Each of the three exposed turtleback detachment faults is now active as a low-angle normal fault. A large, active normal fault, limited to upper-plate rocks, ends truncated downward against each turtleback (Fig. 7). Footwall Neogene volcanic rocks, hanging wall Quaternary slide breccias and the late-Quaternary Artist's Drive fault (Figs 4 & 5) are all truncated downward against the gently dipping Badwater turtleback. The Copper Canyon turtleback detachment truncates, from beneath, an upper-plate complex of footwall plutonic rocks, hanging wall tilted Pliocene strata and the intervening large normal fault (Fig. 14). Quaternary strata are rotated to steep truncations against the Mormon Point turtleback fault. In each case, the upper-plate rocks are slipping in the downhill direction on the turtleback detachments; gravitational sliding toward fast-deepening Death Valley is indicated. Faults that originated as zones of ductile slip in the middle crust have risen to the surface and have evolved until they now bound what are in effect giant landslides. Turtleback and upper-plate complexes alike are broken by normal faults that define parts of the present range front (Fig. 14).

Slip on the Bare Mountain detachment was largely completed before about 11.5 Ma, whereas detachments in the Bullfrog Hills continued to be active until perhaps 7 Ma; the Funeral Mountains were probably affected by considerable Pliocene slip and activity has continued throughout the Quaternary in the Black and Panamint Mountains. Late-Quaternary normal faults, with the usual 50° dips, break the Panamint and Black Mountains (Fig. 14) detachments. Major detachment faults must now be active at shallow depth beneath Death Valley, Panamint Valley, and areas W and NW of them.

Above all of these late-Neogene detachment faults, upper-plate strata, Proterozoic to lower-Pliocene, mostly display rotation to general steep to moderate E dips and hence a consistent slip relatively westward across both flanks of each detachment dome, up E flanks and down W flanks. Fault segments of opposite dips belong to continuous domiform or undulating structures, not, as they have often been interpreted, to kinematically unrelated structures sliding down the sides of domes, although the turtleback faults have evolved recently to the condition where such sliding is now occurring. Except for this final phase, which may be exemplified also by late slip on the Tucki Mountain (Fig. 12) and Wildrose Canyon detachments, the domes did not extrude upward, shedding slide masses off their flanks, for regional slip of upper plates requires continuous cover.

Tertiary strata comprise about half of the aggregate width of the upper plates truncated downward against detachment faults exposed in the Death Valley region (Fig. 7), and this requires at least 100% extension of the pre-Tertiary crust. Oligocene and Miocene strata may represent basins of markedly different configuration to that of the Pliocene ones, which in turn were quite different from modern ones (Cemen *et al.* 1985). Basin fills were dragged directly against detachment faults; pre-Tertiary upper-plate rocks did not underlie the basins. B.C. Burchfiel (pers. comm. 1985) regards his mapping as requiring similarly that the Quaternary fill of the actively widening northern Panamint Valley be dragging directly on a shallow detachment fault.

Throughout the region of Fig. 7, high, young fault scarps of the type commonly thought of as typifying Basin and Range structure are developed mostly in the upper-plate assemblages of detachment faults. The lower plates outcrop primarily as domiform masses and are broken into tilted fault blocks to only a minor extent. Relationships between modern and ancient basins and ranges and domiform detachment faults can be seen in Fig. 7. Most of the modern relief correlates with a position relative to the domes rather than with block faulting younger than the exposed detachment faults. The domiform sub-detachment masses and the resistant parts of the tilted upper-plate rocks above them form the modern ranges. The palaeobasins and the palaeoranges represented by, respectively, Tertiary and pre-Tertiary upper-plate materials are distributed randomly with regard to trough or crest positions across the domes. Perhaps initial basins and ranges were positioned systematically above lows and highs in underlying detachment faults (cf. Davis 1980) but were dragged across the expanding composite surface of sub-detachment lenses to their present positions. Where resistant rocks stand high atop a dome, normal faulting toward the adjacent lowlands is facilitated.

Active right-slip faults trend NW and serve as transforms between the modern obliquely separating, N-trending basin-range blocks. I infer that the sub-detachment crustal lenses, now active at depth, are sliding apart in a NW–SE direction beneath the modern blocks, that the northerly trends of those blocks are inherited from older extensional systems and that the strike-slip faults are restricted to upper plates allochthonous above detachments.

The detachment faults disappear beneath upper-plate materials at the E and W sides of the region discussed here. Rock assemblages of upper-plate type are broken by small to very large late-Cenozoic normal faults for about 75 km in both directions. As the age range of normal faulting differs across the region, the age range of major subjacent detachment faulting probably varies correspondingly.

Stewart (1983a) inferred that the part of the Panamint Mountains here regarded as the tilted upper-plate crustal block slipped about 45 km relatively westward from above the Black Mountains turtleback complexes. I concur with this conclusion; Cemen *et al.* (1985) do not.

Increasing area of detachment faults

If the analysis of detachment geometry summarized here is basically correct, then the widening of a normal-faulted terrain must be a response to an increase in area of the underlying detachment faults. The ages and intensities of detachment faulting over broad terrains vary complexly. Upper and lower plates do not display either systematic variations in initial crustal depth nor slips that increase undirectionally, both of which would be required were detachments planar faults cutting through the crust.

Sub-detachment structure

I infer that the middle crust is extended by the sliding apart, along ductile shear zones, of lenses that are little deformed internally. The area of the composite top of these lenses thus increases as deeper lenses are pulled out from beneath shallower ones.

Beneath the cataclastic carapaces beneath detachment faults, lower-plate rocks largely retain pre-detachment fabrics: synextensional ductile shear occurred in discrete zones and was not pervasive. Still deeper, additional gently dipping zones of brittle or ductile shear, discordant metamorphically and structurally with the older rocks, have been recognized (Figs 22 & 23). Such zones divide sub-detachment complexes into lenses. Major faults beneath the upper

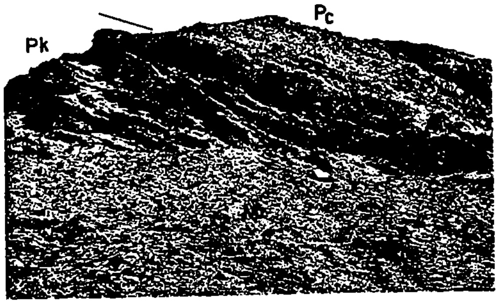

FIG. 22. Horizontal sub-detachment Tertiary fault, Big Maria Mountains, California. Permian Kaibab Marble (Pk) and Coconino Quartzite (Pc) are truncated downward against sub-horizontal Mesozoic metasedimentary rocks (Mz), all metamorphosed at upper greenschist facies. Overturning is a product of Cretaceous recumbent folding; slip on the fault is a few km.

FIG. 23. Internally isoclinal mylonite of low greenschist facies, in fault zone near and like that of Fig. 22. Layers and lenses are rich alternately in mylonitic marble (m) and mylonitic greenschist (g; note contortions in lowest mass) but the two types are finely mixed. Kaibab Marble (Pk) at top is boudinaged but not mylonitized.

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detachment faults in the Death Valley region were noted previously. John (1982) outlined a large sub-detachment lens with her Mohave Wash fault, and Spencer & Turner (1982) defined a smaller one with their 'lowest detachment fault'. Howard *et al.* (1982) found major sub-horizontal faults just above a detachment. I suspect that two very large lenses are in exposed contact in the Harcuvar Mountains of SW Arizona (cf. Rehrig & Reynolds 1980); a large eastern mass of deep-seated plutonic rocks has a domiform mylonitic carapace, which is overlain in the E by an exposed detachment fault and upper-plate assemblages, but in the W, at Cunningham Pass, by another sub-detachment mass, domiform in its topography, composed of granitic rocks typical of the upper part of the middle crust.

I infer that such lens-defining faults are common although relatively few have yet been mapped. I consider that junctions between detachment domiforms are mostly dihedral angles where deeper lenses emerge from beneath shallower ones; but such junctions in general are hidden beneath upper-plate rocks or beneath the alluvial fill of basins. Such lenses are shown in Fig. 24. In SE California, an undulating detachment fault dips S from the Riverside Mountains (Fig. 6) and another dips N from the Big Maria Mountains, the next range to the S; the distinctive lowerplate rock assemblages of the two ranges are very similar, except that markedly higher-grade Cretaceous metamorphism has affected the southern range, and it is possible that the Riverside lower plate moved about 25 km, relatively NE, from an initial position atop the Big Maria lower plate.

Domal form

Detachment faults can be irregular, undulating surfaces, but many outcrop as broad domes (e.g. Davis *et al.* 1980; Davis 1980; Rehrig & Reynolds 1980; Figs 7 & 20). Flank dips vary from a few to 25 degrees, locally more. Domes can be symmetrical or irregular, and typically have exposed dimensions of 10–20 by 20–40 km. Some are elongate in the direction of slip of overlying plates, as defined by the perpendicular to the axis of rotation of Tertiary rocks and by ductile-slip fabrics and, like parallel second-order corrugations of the surfaces, may be 'porpoising mullion structures' (John 1984; cf. Wright *et al.* 1974). Other domes are elongate at high angles to slip directions, or are irregular and I consider that these shapes record the interactions between diversely shaped lenses as they slid apart.

Other explanations that have been proposed for the domes do not appear to me to be broadly viable. Conjecture that the domes are products of the diapiric rise of hot cores that shrugged off their covers (e.g. Drewes 1981) is disproved because major offsets on bounding faults show no preference for down-dip directions, consistent senses of rotation of upper-plate rocks instead being maintained across series of domes (Fig. 7). Explanations invoking local heat sources (e.g. Miller *et al.* 1983) cannot generally be valid because many sub-detachment complexes were at only ambient middle-crust temperatures during extension. Plutonic rocks are present in many domiforms, as they are throughout the middle crust, but are of all ages from Precambrian to Miocene, whereas the domes were outlined in Cenozoic time and many domes expose little magmatic rock of any age. The common denominator of rocks beneath detachment faults is the depth, not the temperature or age, of their formation. Compressive folding could not have produced the domes, for it is incompatible with the continuing formation of extensional structures shown by many domes. Doming by isostatic uplift following differential tectonic denudation (Spencer 1984) would require, in view of the relatively steep flanks and small dimensions of many of the domes, an improbably low viscosity of the lithosphere.

Dykes

Sub-vertical mafic to silicic dykes of mid-Tertiary age, perpendicular to the direction of extension, are present in many ranges, mostly beneath detachment faults, but are generally sparse. Dyke swarms are present only locally (e.g. John 1982; Spencer & Turner 1982). The intrusion of dykes was therefore not a major contributor to regional extension at exposed crustal levels. Many lower-plate dykes exhibit static alteration to greenschist-facies minerals, which is thought to indicate emplacement at mid-crust depths and retention at greenschist temperatures.

Superimposed structures

Many detachment faults are broken by steep normal and strike-slip faults. Presumably the detachment complexes had risen to high crustal levels where further low-angle slip was not possible, and subsequent extension was accommodated

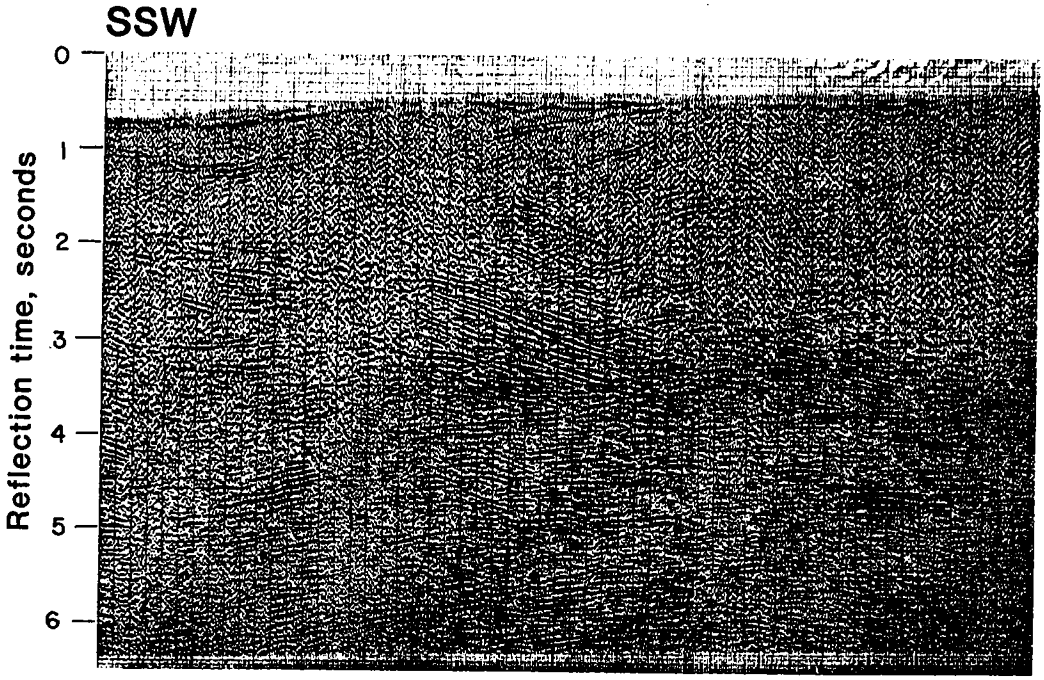


FIG. 24 (Above, and facing page). Deep-crustal seismic reflection profile in S-central Arizona (Fig. 1). The northern two-thirds of the line lies on or near outcrop of lower-plate middle-crust plutonic rocks, structurally beneath a mid-Tertiary detachment fault that outcrops with a northerly dip at the N end of the profile. The southern one-third lies on alluvial basin fill, and underlying bedrock may contain a detachment fault dipping SW. A well a few km W of the S end of the line penetrated 1 km of gravel and arkose, 2 km of upper-plate granitic rocks, a detachment fault at 3 km, and 2.5 km of sub-detachment plutonic rocks, including Tertiary muscovite granite, to bottom (Reif & Robinson 1981). Five biotite and hornblende K–Ar ages between the detachment fault and the bottom of the hole range only from 25 to 31 Ma (Reif & Robinson 1981), indicating rapid uplift from mid-crustal depths at the time of detachment faulting. The depth scale is based on regional refraction velocities. The profile is 40 km long; effective horizontal scale is 1.6 × vertical scale. The N end of line is near 33°05′ N, 111°00′ W; the S-end is near 32°50′ N, 111°10′ W. Profile provided by Pacific West Exploration and the Anschutz Corporation.

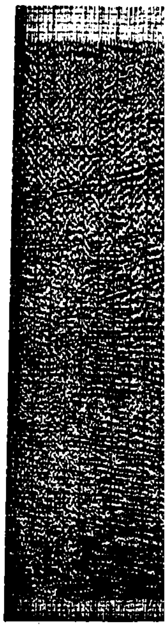
Interpretation: Reflectors represent variable mylonitization and shear transposition during Tertiary extension. The top 10 km (present upper crust, but middle crust before extension) is broken into great lenses which have slid apart; few reflections come from inside the lenses. The undulating apparent fabric of the lower 10 km (pre-extension lower crust) may, if real, record pervasive ductile flattening.

in the rotating-block, upper-crustal mode, new detachment faults being developed at depth (Figs 2 & 14).

Structures of the lower crust

The mode of extension deep in the crust cannot yet be inferred from field observations. Lower continental crust is exposed along the SW edge of the province, but deep extensional structures have not been recognized there.

Reflection profiles in Arizona yield information that may prove relevant to the behaviour of the deep crust. These profiles mostly remain proprietary, but a public one is reproduced as Fig. 24. The ground surface along most of this profile is eroded beneath a detachment fault, hence represents a crustal depth of at least 10 km before mid-Tertiary extension. The top 10 km of the profile displays anastomosing zones of apparent reflectors which I consider to be shear zones and to represent, in part, pre-existing lithologies transposed in the shear direction and, in part, velocity anisotropy in variably mylonitized rocks (cf. Fountain *et al.* 1984; Jones & Nur 1984). I conclude that the acoustically transparent rock between reflective zones consists of large lenses that retain pre-shearing massive or heterogeneous fabric. The lower half



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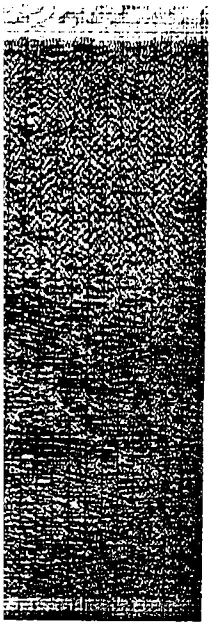

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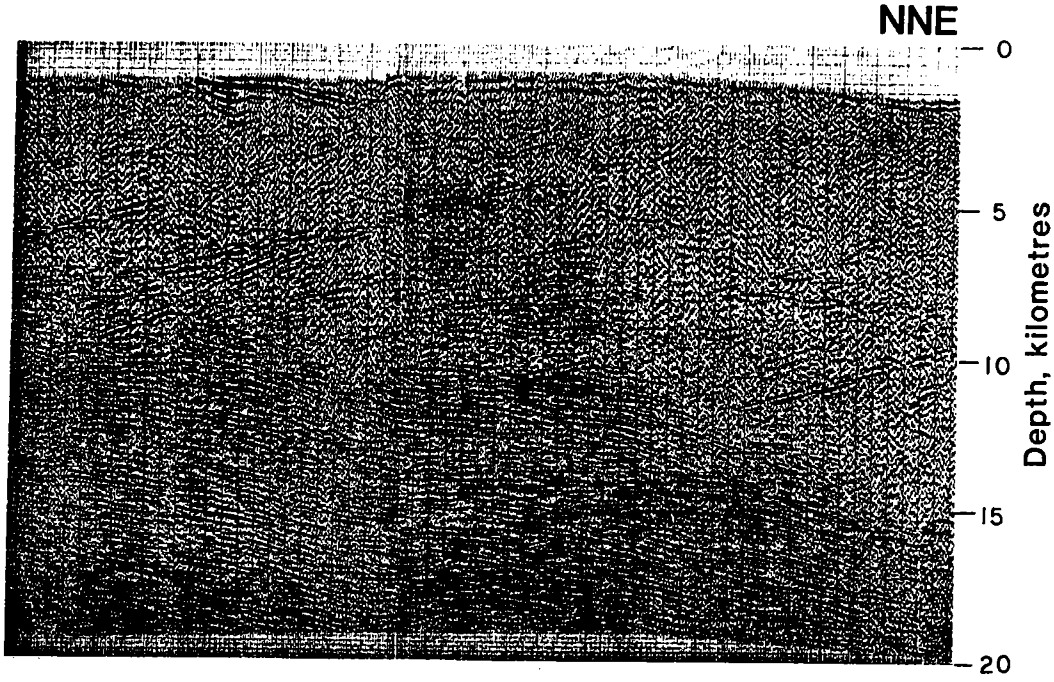

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of this profile, like many of the other Arizona profiles, shows a strong fabric of gently undulating apparent reflectors. If such fabrics truly record varying velocity layering in the deep crust, and are not artefacts due to over-zealous processing or to such acoustic effects as reverberation, then pervasive ductile flattening may be indicated. Such pervasive deformation can be inferred also on theoretical grounds (Sibson 1983).

Other regions

The model advocated here may have a general application to extended continental crust. Domiform detachment faults separating steeply rotated upper-crustal rocks from mid-crustal complexes may be widespread around the world. Most have been interpreted in terms of thrust faulting because of their gentle dips, although this is disproved where thick crustal sections are omitted across them. A few examples are noted here.

The numerous 'Shuswap' domes of mid-crust rocks in NE Washington and interior British Columbia are bounded by detachment faults strikingly like those of the Basin and Range Province, except that those northern faults are of Eocene age. The structures display the same progressions of ductile to brittle structures in carapaces on the lower plates (Lane 1984, and references therein). Lower plates consist of mid-crust crystalline rocks, varying from kyanite greenschist to migmatite and muscovite granite and, locally, granulite, of Proterozoic to Eocene age. Upper plates consist of little-metamorphosed Proterozoic to Eocene strata, often truncated steeply against the gently dipping detachment faults, and upper-crust cross-cutting granitic rocks and their contact-metamorphosed wall rocks. Eocene K–Ar ages in many lower-plate rocks indicate great uplift at that time. Most investigators have interpreted these structures as being due to Mesozoic thrusting or rising of plutons, with minor Tertiary complications. Fox & Beck (1985) are among the few who have explained the structures, largely in terms of Eocene extension, and with this I concur.

The island of Naxos in the Aegean Sea is a detachment complex of Basin and Range type in an extensional setting (Lister *et al.* 1984). Beneath a domiform detachment fault is a lower plate of melange, metamorphosed at blueschist facies in the Palaeogene and then metamorphosed and migmatized under mid-crustal conditions in the Miocene. The upper plate includes unmetamorphosed Palaeogene melange and upper-crustal

Miocene granitic rocks. Both Palaeogene and Miocene indicators thus require great crustal elision across the detachment.

The Betic Cordillera of SE Spain contain high-grade metamorphic rocks separated by gently domiform faults from overlying unmetamorphosed Mesozoic to mid-Tertiary strata, which have steep to moderate dips (Platt 1982; Volk 1967). The faults have generally been assumed to be thrusts despite their elision of thick crustal sections. Platt recognized their extensional nature, although he invoked a rising-dome, gravity-sliding model.

Similarity to thrust deformation

The middle continental crust may deform in the same general lenticular mode under compressive overthrusting as under extension. Late Cretaceous mylonite zones each hundreds of metres thick outline huge, overlapping lenses in a gently dipping stack in the Santa Rosa Mountains of SW California. Reflection profiling shows that the upper-crust basement Palaeogene thrust that bounds the Wind River Mountains of Wyoming flattens downward into a series of lens-defining splays at depths of 17–26 km (Sharry *et al.* 1986) acoustically appropriate for mylonitic zones (Jones & Nur 1984). Such thrust lenses are defined by ductile shear zones in the Proterozoic Grenville crustal-thrust terrain of Ontario, where the pervasiveness of deformation increases with depth of exposure within the lower crust (Davidson 1984).

A great contrast between compressive and extensional regimes is that depth and temperature of mid-crust components increase with time in the former, but decrease in the latter.

Acknowledgments: I started mapping what are now referred to as detachment faults, in SE California 25 years ago, when I termed them 'gravity thrust faults'; but it was not until I attended the 1977 Geological Society of America Penrose Conference on 'metamorphic core complexes', in Tucson, that I began to appreciate the common characteristics and regional prevalence of these remarkable structures. That fine conference was organized by M.D. Crittenden, Jr, P.J. Coney, and G.H. Davis. Discussions and field trips with, and papers written by, many geologists since then have profoundly influenced my views. My recent debt is greatest to G.A. Davis and K.A. Howard, whose interpretations in many ways overlap my own. This manuscript was much improved as a result of criticism by M.D. Carr, A.M.C. Sengor, and J.H. Stewart.

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