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Detachment faulting in continental extension; Perspectives from the Southwestern U.S. Cordillera

Gregory A. Davis
Department of Geological Sciences, University of Southern California, Los Angeles, California 90089-0741
G. S. Lister*
Bureau of Mineral Resources, Geology and Geophysics, Canberra City, 2601 Australia

ABSTRACT

Cordilleran detachment faults, as defined here, are extensional faults of low initial dip, probably less than 30°, and subregional to regional scale. Some detachment faults have large translational displacements, i.e., in excess of several tens of kilometers. First interpreted as Tertiary extensional structures in the eastern Great Basin by Armstrong (1972), they are now known to be widespread throughout those Cordilleran regions that have undergone greatest Cenozoic extension. Detachment faults are commonly, but not necessarily, associated with lower-plate mylonitic gneisses that compose the so-called "metamorphic core complexes." Probably nowhere in the U.S. Cordillera are detachment faults more widely and spectacularly developed than in the region that borders the lower Colorado River in southernmost Nevada, southeastern California, and southwestern Arizona. We believe that our studies and those of numerous other workers in this region, the Colorado River extensional corridor of Howard and John (1987), provide a number of new perspectives on the origin, geometry, and evolution of Cordilleran detachment faults.

Detachment faults are best explained as evolving shallow-dipping shear zones that have accommodated Tertiary crustal extension (Wernicke, 1981). The fault zones are believed to root at midcrustal or lower upper crustal depths into broad zones of intracrustal flow, the tectonic regime in which mylonitic gneisses form. At their upper ends, major detachment faults either reach the surface directly or terminate at shallow depth into pull-apart complexes of closely spaced normal faults. Along these evolving shear zones, lower-plate mylonitic gneisses are drawn upward and out from beneath upper-plate rocks. As footwall gneisses rise structurally upward, they are retrograded, sheared, and shattered at progressively colder and shallower crustal levels to form the chloritic breccias and microbreccias characteristic of many major detachment faults. At advanced stages of detachment fault evolution, lower-plate mylonitic gneisses formed at depths >12 km are tectonically juxtaposed beneath unmetamorphosed supracrustal rocks and exposed at the surface through combinations of crustal upwarping, tectonic denudation, and erosion.

Contrary to popular belief, the master detachment faults exposed today are probably not in their entirety those faults that formed at the start of extensional deformation, but rather are only the youngest in a succession of major detachment faults. Detachment faults undergo warping at high angles to the direction of crustal extension, probably in large part related to isostatically induced distortions of originally more planar faults. Such warping leads to the development of younger, more planar fault splays that either

*Present address: Department of Earth Sciences, Monash University, Clayton, Victoria, 3168 Australia

cut upward into former upper-plate rocks (excisement) or downward into former lower-plate rocks (incisement).

Recognition of such geometric complexities offers fresh insights into deciphering the evolving strain patterns within major detachment terranes. Studies in the Whipple Mountains region of southeastern California indicate that: (1) detachment faults have formed by both excisement and incisement tectonics; (2) northeast-southwest-trending "folds" of major detachment faults, oriented parallel to the direction of extension, are in reality primary corrugations or flutes in the fault surface (a conclusion previously reached by other workers in nearby areas); (3) most normal faults in the upper plates of major detachments originally had listric geometries before losing their flattened lower segments as the consequence of excisement tectonics; and (4) detachment faults can transect upper crustal rocks as primary, low-dipping shear zones without pre-existing, shallow-dipping structural controls (e.g., thrust faults) on their localization; the northeast-southwest-trending curviplanar geometry of the Whipple fault does, however, seem to mimic preexisting fold structure in lower-plate mylonitic gneisses crossed by the fault. Finally, the rate of translation along master faults of some evolving detachment systems apparently can be very rapid (>1 cm/yr), much faster than rates that we (and perhaps other workers) once deemed reasonable. A very good case can be made on the basis of geochronologic and field studies that footwall mylonitic gneisses were transported upward along the Whipple detachment system from lower upper crustal depths to near surface levels in less than 2 m.y. (between 18 and 20 Ma).

INTRODUCTION

With few exceptions, information published prior to 10 years ago on extension of the continental crust focused attention on moderately to steeply dipping normal faults as the fundamental structures of extended terranes. Steeply dipping swarms of igneous dikes were recognized as the consequences of limited crustal extension in some areas (e.g., feeder dikes for Miocene Columbia Plateau basalts in northeastern Oregon and west-central Idaho). Some investigators of the Basin and Range province of the U.S. Cordillera (e.g., Thompson, 1959; Stewart, 1971) realized that the widely spaced (ca. 20–30 km) and relatively steeply dipping (>50°) range-bounding faults of that province could produce only limited crustal extension, and that below some critical depth (ca. 8–15 km) the rheology of the hotter than normal Basin and Range crust should not permit extensional strain by normal faulting. Estimates of total extensional strain in the province by these geologists were geometrically constrained to low values, e.g., 10 to 15 percent, by the wide spacing and steep dips of observed range-front faults.

Other workers, however, were troubled by such low estimates. Hamilton and Myers (1966), impressed by geophysical data indicating that the Basin and Range crust was considerably thinner than that beneath adjacent provinces to the west and east, concluded that Cenozoic extension had probably been on the order of 50 to 100 percent. Subsequent field-based studies, such as those of Anderson (1971), Armstrong (1972), and Davis and Burchfiel (1973), indicated that certain areas within the Basin and Range province had indeed undergone extension compatible with Hamilton and Myers's estimates for province-wide strain. Davis and Burchfiel, for example, analyzed geometric relations within a geologic terrane offset by the major northeast to east-striking Garlock fault of southern California. They concluded that extension within the 200-km-wide Great Basin area north of the fault had exceeded extension in Mojave Desert areas to the south and east by approximately 100 percent, and they interpreted the Garlock fault as an intraplate transform bounding the southern limit of the Basin and Range province in California.

Cordilleran geologists now agree that the Basin and Range province (Fig. 1) has indeed experienced larger (ca. 50–100%) extensional strains than those envisioned by most early workers, although such strains are nonuniformly distributed across the province (e.g., Stewart, 1978; Miller and Gans, 1984). Such agreement has come about through a growing awareness of two major aspects of basin-and-range geology: (1) that crustal extension occurred widely throughout the province prior to the late early to middle Miocene (locally, Pliocene) formation of present basin-and-range topography (e.g., Loring, 1972, 1974; McDonald, 1976); and (2) that low-angle normal faults are widespread and extremely important structural elements of those Cordilleran regions that have experienced the greatest extensional strains.

The second of these geologic aspects is the principal topic of this paper. Shallow-dipping normal faults were until recently the neglected structures of the U.S. Cordillera. In eastern Nevada and western Utah, areally extensive faults of this type were first described by Peter Misch and his students in the late 1950s and 1960s (e.g., Snelson, 1957; Misch, 1960; Nelson, 1966, 1969; Thorman, 1970). These faults, generally developed within pre-Jurassic strata, were interpreted as updomed, hinterland exposures of Mesozoic thrust faults belonging to the foreland fold and thrust belt to the east (Misch, 1960, 1971). Geometrically similar faults, clearly of Cenozoic age because they truncate inclined

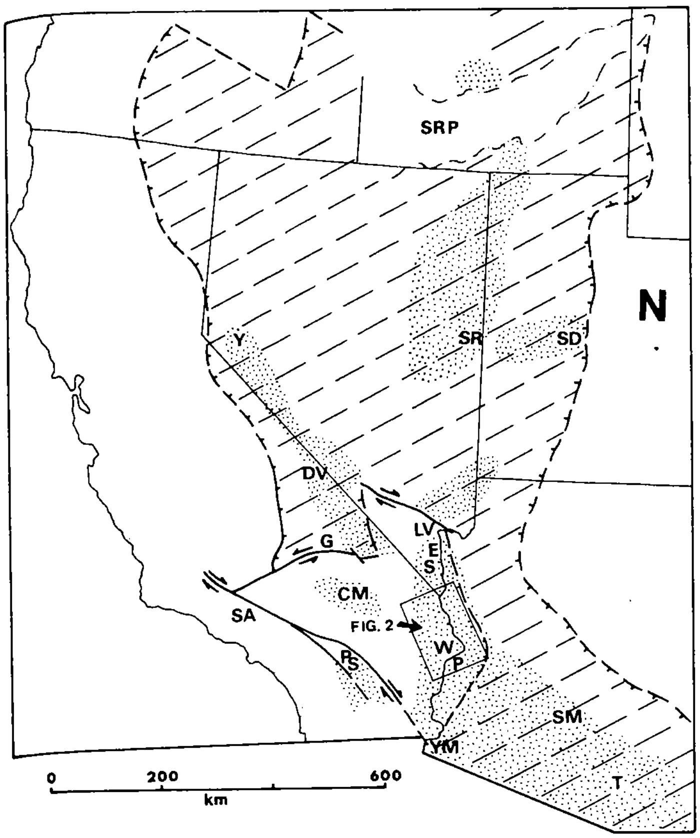


Figure 1. Location map of southern U.S. Cordillera showing area of Basin and Range province (diagonally ruled pattern) and locations of detachment fault complexes (stippled pattern). Geographic localities and geologic features cited in the text (in alphabetical order): CM = central Mojave Desert; DV = Death Valley; E = Eldorado Mountains; G = Garlock fault; LV = Las Vegas; P = Parker; PS = Palm Springs; S = Searchlight; SA = San Andreas; SD = Sevier Desert; SM = South Mountain; SR = Snake Range; SRP = Snake River Plain; T = Tucson; W = Whipple Mountains; Y = Yerington; YM = Yuma.

Tertiary strata in their upper plates, are also widely exposed in the Death Valley–eastern Mohave region and in an arcuate belt extending southward from Las Vegas along the lower Colorado River and curving southeastward into southern Arizona. However, they too, were generally interpreted as thrust faults (e.g., in the Death Valley–eastern Mohave area: Noble, 1941; Hewett, 1956; Hunt and Mabey, 1966; in the lower Colorado River area: Ransome, 1931; Coonrad, 1960; Wilson, 1962; Terry, 1972) or as unconformities (Kemnitzer, 1937; Terry, 1972; and Dibblee, 1971, for the central Mohave region). In retrospect, it is difficult to understand why the low-angle faults in these geologic terranes were so widely mistaken for thrust faults, given that they characteristically juxtapose younger over older rock units, and unmetamorphosed or low-grade metamorphic rocks over higher grade metamorphic and/or mylonitic rocks.

Armstrong (1972) was the first to reinterpret the low-angle faults in areas studied by Misch and his students as the consequence of major regional Tertiary extension, although Lee and others (1970) had earlier recognized Tertiary movement on the Snake Range décollement from Tertiary K-Ar dates in fault zone rocks. Terming such structures "denudation faults," Armstrong believed that they were the product of regional extension, beginning about 40 Ma, of a crust thickened during Mesozoic compression. He recognized the likelihood that some low-angle normal faults might reactivate pre-existing thrust faults, and he suggested that faults formed at depth during early stages of extension might be cut by steeper, younger faults of higher crustal position as tectonic denudation continued.

Some Tertiary low-angle faults described in the Basin and Range area during the 1970s were initially interpreted as extension faults, rather than being misidentified as thrust faults, because of observable offsets of Tertiary strata in their hanging and foot walls. Such faults, with low dips attributable to primary listric geometry or to rotation of once-steep faults, are not as cryptic as the major low-angle "denudation" faults of the eastern Great Basin, or those of the lower Colorado River–southern Arizona area where Tertiary strata are present only in upper-plate positions. Ernest Anderson's recognition (1971) of downward-flattening (listric) late Miocene normal faults in the Eldorado Mountains, southern Nevada, was a major contribution to Cordilleran extensional tectonics for at least two reasons: (1) it documented the existence of closely spaced listric normal faults, which were interpreted as flattening downward into either a subhorizontal basal fault surface or inflating plutonic complex below which extension by normal faulting had not occurred; and (2) it demonstrated that the listric geometry of the imbricate Eldorado Mountains faults and the consequent down-dip rotation of late Tertiary strata along them had produced large extensions of supracrustal rocks. As the Eldorado normal faults become subhorizontal, some steeply dipping strata truncated by them are subhorizontally translated by amounts equal to the cumulative dip-slip on the entire family of listric structures. Hence, Anderson's paper was among the first to demonstrate the efficacy of low-angle normal faults in producing large extensional strain in the crust transected by them.

Proffett's study of multiple phases of normal faulting in the Yerington area of west-central Nevada (1977; Proffett and Dilles' detailed map, 1984) was another landmark contribution. In that area, very shallow dipping normal faults displace steeply dipping Tertiary strata and basement rocks for distances up to several kilometers. Proffett interpreted the shallow-dipping faults as originally much steeper, moderately listric structures that cut subhorizontal Tertiary strata and its underlying basement. His explanation for present geometric relations relied on downward rotation of hanging wall units along early formed listric faults, followed by the repeated rotation of these faults and their wall rocks along younger generations of listric faults. Proffett's thorough documentation of multiple, overprinted episodes of closely spaced normal faulting helped direct geological attention away from the simple, wide-spaced range-bounding fault system of the Great Basin. It also illustrated that rotated low-angle normal faults, however formed, occur far from Anderson's Eldorado Mountains and are extensional phenomena that must be reckoned with in analyses of province-wide strain.

Fault geometries quite comparable to those described by Proffett have been mapped above the Snake Range décollement (detachment) in eastern Nevada (Gans and Miller, 1983; Miller and others, 1983), but there, domino-style rotation of originally moderately dipping (50° to 60°) planar faults of two generations is interpreted as the mechanism for producing very low dipping normal faults and extensions locally as great as 450 to 500 percent. Rotation is restricted, however, to the upper plate of the Snake Range décollement.

Low-angle faults in the Eldorado and Yerington areas owe their low dips to various combinations of original listric geometry and/or rotation of listric or planar normal faults. But such faults are not the detachment faults of principal concern to this chapter, although it is likely that they have developed in the extending upper plates of unexposed detachment systems. It is to such detachment fault systems that we now turn.

DETACHMENT FAULTS

The detachment faults of the Cordillera have known many names: décollements, abscherung zones, denudational faults, dislocation surfaces, and LANFs (low-angle normal faults; Wernicke, 1981; Brun and Choukroune, 1983), but the term detachment fault (Carr and Dickey, 1976; Davis and others, 1979) has gained most common usage. Two characteristics of detachment faults have already been cited—that they juxtapose younger over older or structurally higher over structurally lower rocks, and that they commonly separate upper-plate unmetamorphosed or low-grade metamorphic rocks from lower-plate crystalline rocks of higher metamorphic grade; the latter commonly have mylonitic gneissic fabrics. Four other characteristics can be added: (1) the master faults of detachment complexes are developed on a regional to subregional scale (e.g., the Whipple detachment fault system of the lower Colorado River area, California, Arizona, and southernmost Nevada, appears to have once underlain at least 15,000 km^2; the northern Snake Range décollement extends for at least 50 km in an east-west dimension); (2) the upper plates of detachment faults are typically extended by one or more generations of normal faults that either merge downward into the detachment or end abruptly at it without a shallowing of dip; similar faults are absent or less well-developed in lower plates; (3) several lines of evidence (e.g., telescoping of metamorphic facies, juxtaposition of deep structural levels beneath much shallower levels, and difficulty in matching displaced upper- and lower-plate units) suggest that some detachment faults have very large displacements; (4) detachment faults are commonly underlain by a distinctive sequence of first ductilely, then brittlely deformed, rocks, with each younger superposed member of the sequence developed in progressively narrower zones. The most common sequence, oldest to youngest, is (a) mylonitic gneisses (not always present); (b) sheared, retrograded mylonitic gneisses (or, in their absence, other crystalline basement rocks) widely termed "chloritic breccias"; (c) pseudotachylites and flinty cataclasites or microbreccias, sometimes with injection vein geometries from layers resembling melt generation surfaces (Sibson, 1977); and (d) fault gouge.

We present evidence elsewhere in this chapter that detachment faults develop initially through the upper crust with low angles of dip, probably less than 25° to 30°. In our opinion, low-angle faults that owe their present shallow dip to the rotation of originally steeper faults (e.g., the Singatse and parallel faults of the Yerington area: Proffett, 1977), or that constitute the lower portion of listric faults that steepen upward (e.g., some faults in the Eldorado Mountains: Anderson, 1971) should not be termed detachment faults. *The essential elements of extensional detachment faults,* as the term is used here, *are low angle of initial dip, subregional to regional scale of development, and large translational displacements, certainly up to tens of kilometers in some instances.*

DETACHMENT FAULTING IN THE COLORADO RIVER REGION SOUTH OF LAS VEGAS

Probably nowhere in the U.S. Cordillera are detachment faults more widely and spectacularly developed than in the region that borders the lower Colorado River between Las Vegas, Nevada and Yuma, Arizona, and a wide contiguous terrane that extends eastward from the river across the southern third of Arizona and into Sonora, Mexico (Fig. 1). This Oligo-Miocene detachment terrane is quite likely offset by the San Andreas fault system near Yuma, Arizona (Garner and others, 1982). Portions of the displaced terrane may be present in the eastern Peninsular Ranges west of the Salton Sea and south of Palm Springs (Wallace and English, 1982; Engel and Schultejann, 1984).

This discussion emphasizes detachment faulting in the northern lower Colorado River (NLCR) region, i.e., from the approximate latitude of Parker, Arizona, northward to the southern tip of Nevada (Figs. 1, 2). The terms "Colorado River extensional corridor" (Howard and John, 1987) and "lower Colorado River detachment terrane" refer to the same region. Most of the studies of the past decade in this region have been conducted by three groups of researchers: by the U.S. Geological Survey, largely under the leadership of Keith Howard, but including important independent studies by Will Carr, Ivo Luchitta, Neil Suneson, and Barbara John among others; by J. Lawford Anderson and G. A. Davis and their students at the University of Southern California; and by Eric Frost and his students at San Diego State University. As one measure of the intensity of geologic studies in the NLCR region, the results of 18 separate investigations in that area by more than 30 earth scientists were published in 1982 (Frost and Martin) in a volume commemorating the pioneering Colorado River area studies of Ernest Anderson and Warren Hamilton. The results of this collective

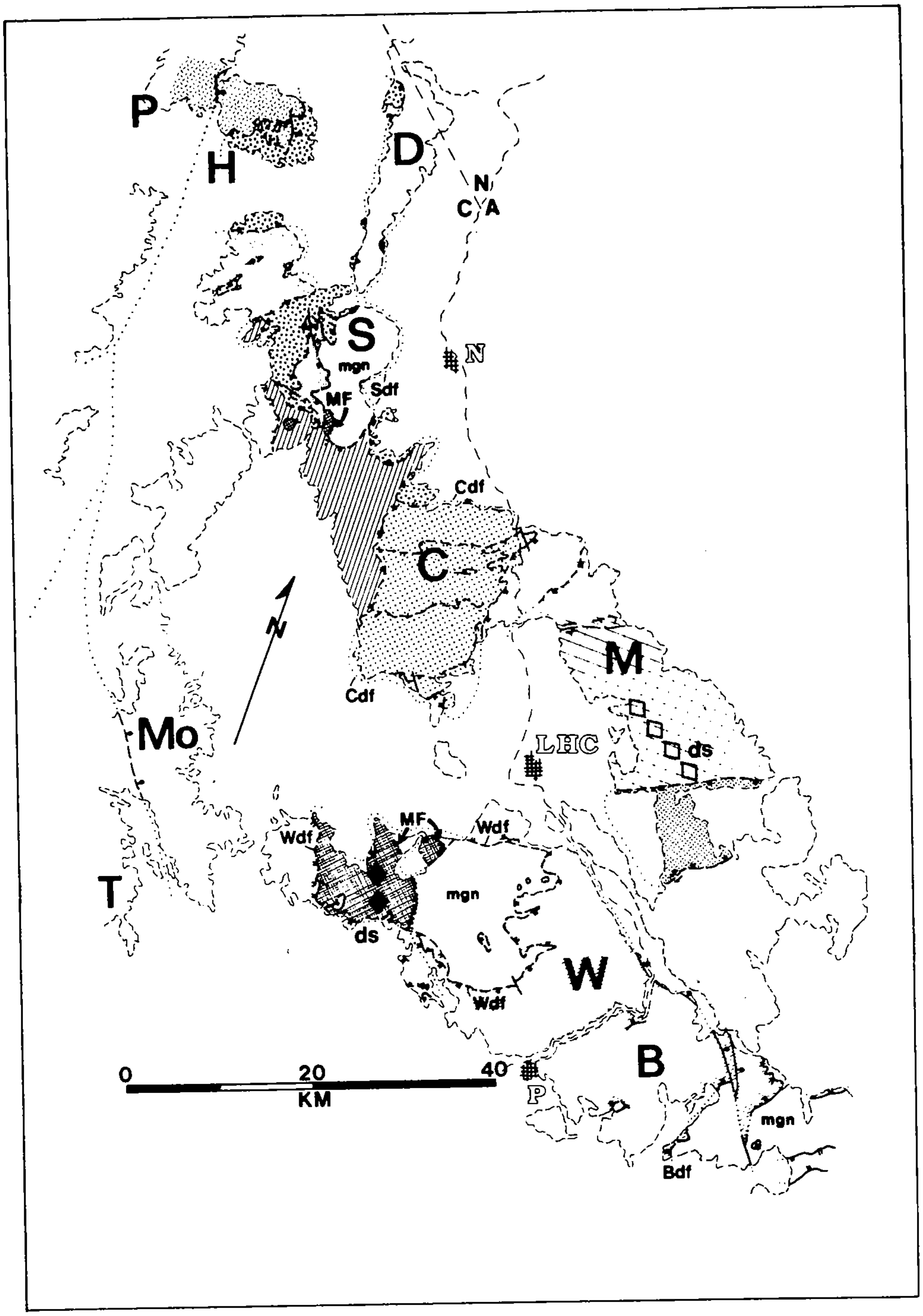


Figure 2. Simplified geologic map of lower Colorado River region of southernmost Nevada (N), southeastern California (C), and western Arizona (A). Map emphasizes location of major faults, primarily low-angle detachment faults. Various detachment-bounded plates are designated by different patterns. Cities from north to south (open letters): N = Needles; LHC = Lake Havasu City; P = Parker. Mountain ranges (in alphabetical order): B = Buckskin; C = Chemehuevi, D = Dead; H = Homer; M = Mohave; Mo = Mopah; P = Piute; S = Sacramento; T = Turtle; W = Whipple. Geologic abbreviations: Wdf = Whipple detachment fault; Sdf = Sacramento detachment fault; MF = mylonitic front; mgn = mylonitic gneiss; ds = dike swarm, offset and displaced by faults of the Whipple detachment system. Data for map (north to south): Spencer and Turner (1982); Spencer (1985); McClelland (1982, 1984); Howard and others (1982b); G. A. Davis and colleagues (unpublished data); and Wilkins and Heidrick (1982).

geological effort in the Colorado River extensional corridor are briefly summarized below in order to define the environment of detachment faulting. The summary includes a number of generalizations with which, we believe, most of the area's workers would agree—a statement not meant to imply that there are no significant differences of opinion about major aspects of the area's evolution.

GEOLOGIC SUMMARY OF THE NORTHERN LOWER COLORADO RIVER REGION

This region is underlain by Precambrian crystalline basement rocks with well-established ties to the North American continent. This basement consists of a heterogeneous assemblage of high-grade orthogneisses, paragneisses, and amphibolites metamorphosed approximately 1.7 Ga; and younger igneous intrusions of diverse age and composition. Among the latter components of the Proterozoic basement assemblage are (1) synkinematic to late kinematic, foliated granitic rocks (ca. 1.7 Ga), (2) postorogenic ("anorogenic") granites (ca. 1.4 to 1.45 Ga; see Anderson, 1983, for description and appropriate references), and (3) diabase dikes and sills (1.1 to 1.2 Ga?).

The Precambrian basement of the region was once covered by a thin (ca. 2 to 3 km) platform sequence of Cambrian through Triassic sedimentary rocks and Jurassic volcanic and clastic rocks related to development of an Andean-type continental margin in the southwestern United States (Burchfiel and Davis, 1972, 1975). Scattered plutons of Jurassic and Cretaceous age are an additional expression of arc development. The Paleozoic and Mesozoic cover was eroded from the NLCR region north of Parker, Arizona, prior to late Oligocene(?) and Miocene deposition of terrestrial sedimentary and volcanic rocks. Pre-Tertiary basement and supracrustal units were strongly folded, thrust-faulted, and metamorphosed during late Jurassic and/or Cretaceous time in areas southwest (e.g., Harquahala Mountains), south (e.g., Big Maria and Riverside Mountains), and west (e.g., Old Woman and Arica Mountains) of the NLCR region shown in Figure 2. The extensional corridor north of Parker (Fig. 2) was in general little affected by this deformation since there is widespread preservation throughout its crystalline basement of complex Precambrian structural and intrusive relations, e.g., in the eastern Whipple Mountains (Davis and others, 1980), the Turtle Mountains (Howard and others, 1982a), and the Buck Mountains east of the Mojave Range, Arizona (Howard and others, 1982b). Precambrian crystalline rocks in at least the southern half of the NLCR region (Fig. 2) did not escape, however, a regional Mesozoic thermal event that reset K-Ar clocks in Precambrian hornblende and biotite to ages typically between 135 and 160 Ma (Anderson and Frost, 1981). Locally, portions of the Precambrian NLCR terrane experienced penetrative strain, possibly during this thermal event. Plutons of the anorogenic 1.4-Ga suite and younger Precambrian diabase sills and dikes exhibit mylonitic shear zones and penetrative foliation development of tectonic origin at some localities in the NLCR region (Howard and others, 1982b; J. L. Anderson, personal communication, 1986). The age or ages, geometry, and extent of post-1.4-Ga deformation are poorly known.

By Oligocene time the lower Colorado River region north of Parker had been denuded of its Paleozoic and Mesozoic cover. During an episode of Oligocene(?) and Miocene arc volcanism, areas both north and south of Parker experienced profound crustal extension. This extension was largely manifested by the development of an east-rooting system of detachment faults in the NLCR region as far north as the town of Searchlight in southern Nevada (Fig. 3). To the north of Searchlight (Fig. 1) lies a separate Miocene extensional terrane, studied by Anderson (1971) and mentioned briefly above. The geometry of fault structures within this northern terrane, in the Eldorado Mountains, the Highland Spring Range to the west, and the McCullough Mountains still farther west, is compatible with the existence of a west-rooting detachment system beneath it (S. Davis, 1984), although neither the inferred detachment nor rocks that would be lower plate to it are known to be exposed.

Multiple low-angle faults have been recognized throughout most of the detachment terrane south of Searchlight, both in upper- and lower-plate positions (Fig. 2). Interpretations vary as to whether all such faults developed with shallow dips or represent, in some cases, initially steep normal faults that have experienced rotational shallowing of dip. Upper-plate rocks above the major detachment faults in the Colorado River extensional corridor have characteristically been distended by numerous normal faults, the great majority of which dip northeastward at angles less than 60° (Fig. 3). Complex relationships exist in the detachment terrane between multiple generations of extension faults that now dip both steeply and gently; an interpretation of these relationships is one of the objectives of this chapter.

Lower-plate rocks for the east-dipping detachment system include Precambrian metamorphic rocks generally similar to the 1.7-Ga high-grade gneisses and amphibolites of the upper plate, Cretaceous plutons, commonly sheetlike or tabular in form, and Tertiary dikes and plutons. A steep, nonmylonitic, northeast-striking foliation is discernible within some of the Cretaceous (89 ± 3 Ma) plutons in the central Whipple Mountains, and closely parallels what appears to be Precambrian compositional layering and foliation in enclosing gneisses. Its tectonic significance is not known.

Lower-plate crystalline rocks in some ranges within the NLCR region experienced penetrative early(?) to mid-Tertiary deformation within a crustal shear zone. This event was characterized by the development of a shallow-dipping mylonitic foliation and a regionally consistent northeast–southwest-trending stretching lineation in the mylonitic rocks. The uppermost structural level of penetrative mylonitic gneiss development is the "mylonitic front" of Davis and others (1980). It is exposed in both the Whipple and southern Sacramento Mountains (McClelland, 1982; Figs. 1, 2) as a west-southwest–dipping gradational zone several meters to several tens of meters wide (Fig. 2). Exposures of the mylonitic fronts in the two ranges are in a north-

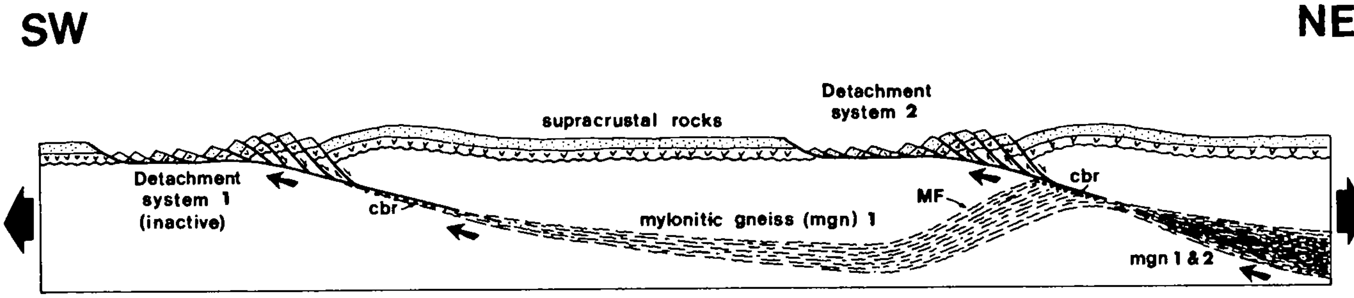


Figure 3. Interpretation of Cordilleran detachment faults as evolving shear zones in an extending continental crust. Width and depth of cross section are approximately 170 and 20 km, respectively. The detachment fault system 1 on the left (southwest) side of the diagram roots at depth into a thick zone of intracrustal flow. Lower levels of mylonitic gneisses formed within that zone are transported upward in the footwall of the detachment fault system where they become retrograded, sheared, and shattered (= chloritic breccias [cbr]) at progressively higher and colder crustal levels. A younger, more easterly splay of the evolving shear zone, detachment system 2, has the geometry of system 1, but captures in its footwall the previously formed mylonitic gneisses of system 1. In this case, the entire kinematically "dead" mylonitic sequence is transported upward; its structural top is mylonitic front (MF) as seen in the lower plates of the Whipple and Sacramento detachment faults, southeastern California (see text). Mylonitic gneisses that form at depth along detachment system 2 presumably overprint mylonitic gneisses formed during earlier phases of deep crustal shear within the evolving zone.

northwest–south-southeast alignment. We interpret the absence of the front and underlying Tertiary mylonitic rocks in the intervening Chemehuevi Mountains (John, 1987) as an indication that levels of exposure within that range are not as deep as they are in the adjacent Whipple and Sacramento mountains. In other words, the mylonitic front lies hidden beneath the Chemehuevis. A corrollary to this conclusion is that the several detachment fault-bounded allochthons of the Chemehuevi Range all lie structurally above the Whipple fault and its presumed Sacramento Mountains equivalent. Mylonitic gneisses now present beneath major detachment faults exhibit shearing and chloritic alteration through thicknesses of as much as 300 m. As discussed below, we interpret both the mylonitization and subsequent shearing and alteration to be phases in a deformational continuum related to Tertiary crustal extension.

Exposures of lower-plate rocks along the extensional corridor from southern Nevada into western Arizona are controlled by two structural elements: (1) a broad (100 ± 30 km) northwest-southeast–trending regional arch; and (2) numerous (ca. 30) northeast-southwest–cross-trending "antiforms" and "synforms" with an average wavelength of approximately 8 km (Frost, 1984) and "limb" dips varying from only a few degrees to as much as 20° to 25°. Lower-plate rocks are exposed in the cores of the doubly plunging "antiformal" structures along the broad hinge of the regional arch. Most recent workers (e.g., Davis and Coney, 1979; Spencer, 1982, 1984, 1985; Howard and others, 1982a) call upon differential isostatic uplift resulting from tectonic denudation of the upper crust as an explanation for the broad northwest-southeast arching at high angles to the direction of northeast-southwest extension.

The most prominent "folds," however, trend parallel to the regional extension direction. An interpretation gaining favor for these northeast-southwest–trending structures is that they are not folds at all, as initially assumed by most workers in the extensional corridor, but gigantic primary flutings or corrugations of the detachment faults oriented parallel to displacements along them (Woodward and Osborne, 1980; Woodward, 1980; Osborne, 1981; Wilkens and Heidrick, 1982; John, 1987). One of us (Davis, *in* Davis and others, 1980, 1982, Fig. 1) has until recently regarded such structures in the Whipple Mountains as late-formed folds because the largest "antiform" in the Whipple detachment fault (Figs. 4, 5) is defined by the geometry of folded, lower-plate sheetlike granitic plutons and mylonitic foliation as well.

However, the limbs of the northernmost Whipple antiform as defined by mylonitic foliation dip more steeply (by 5° to 25°) than do limbs of the antiform defined by dips on the detachment fault (Fig. 5). An early, pre-detachment phase of lower-plate folding is indicated by this relationship, and supported by foliation relationships in the west-central Whipple Mountains. Here, the antiformal geometry of mylonitic foliation is not expressed by the more planar orientation of northwest-striking, southwest-dipping foliation directly beneath the Whipple mylonitic front (Figs. 4, 5). Thus, antiformal arching of lower-plate plutonic sheets and mylonitic foliation seem to have predated a higher level of transpositional foliation development directly beneath the mylonitic front. The front is in turn cut discordantly by the War Eagle detachment fault and the still-younger Whipple fault. It seems appropriate to conclude that the curviplanar geometry of the Whipple fault is not due to folding per se. Instead, the shape

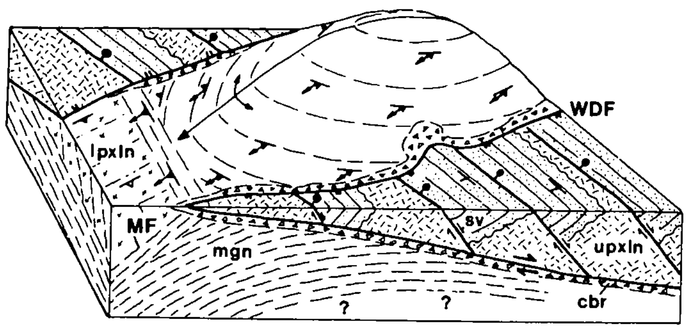


Figure 4. Diagrammatic representation of geologic relations, Whipple Mountains, southeastern California, viewed to north. Width of block diagram is approximately 30 km; vertical topographic relief, ca. 1.1 km, is highly exaggerated in the diagram. Structural features designated by symbols: WDF = Whipple detachment fault; MF = mylonitic front. Rock units: lpxln = lower plate crystalline rocks (predominantly Precambrian gneisses); mgn = undifferentiated mylonitic gneisses (see text); cbr = chloritic breccias; upxln = upper plate crystalline rocks (predominantly Precambrian, but not correlative with "lpxln" within area of figure); sv = Miocene sedimentary and volcanic rocks.

of the fault can be described as corrugated, with the axes of the corrugations oriented parallel to the well-documented direction of fault displacement. This corrugated geometry may have been strongly influenced by preexisting patterns of foliation-defined folds in mylonitic gneisses cross-cut by the fault (i.e., the form of the fault mimicked to some extent the curviplanar structure in the gneisses). John (1987) reached a similar conclusion in her excellent treatise on detachment faults in the Chemeheuvi Mountains, noting that, "There appears to be some influence of pre-existing structures on the overall geometry of the fault system," and the strength of her arguments led us to reevaluate our former position on folded detachment faults.

The tectonic implications of the juxtaposition of upper- and lower-plate units in the lower Colorado River detachment terrane deserve special attention. Mylonitic gneisses of Tertiary age lie in the footwalls of some of the region's major detachment faults, commonly directly below Tertiary strata of broadly equivalent age. Thus, Oligo(?)-Miocene sedimentary and volcanic rocks were being deposited on Precambrian and younger basement rocks while mylonitic gneisses were forming by ductile shear at greater depth in that basement. As explained in greater detail below, J. L. Anderson (1981, 1985, 1987) has undertaken thermobarometric studies of mylonitic gneisses from several areas in the Mojave-Sonora Desert region of California and Arizona. His most detailed investigation has been of samples collected in the Whipple Mountains, where such gneisses have a thickness in excess of 3.5 km. According to his analysis, mylonitic gneisses now in the upper two-thirds of the section, formed at an average depth of approximately 16 km $\pm$ 4 km (4.4 $\pm$1.1 kbar).

There has been relatively little erosion of the upper-plate crust in the Whipple Mountains since Miocene time, ca. 20 Ma, because thin (ca. 1–3 km) Miocene surficial deposits are still widely preserved at the earth's surface. It seems clear that the telescoping of upper midcrustal mylonites against coeval supracrustal rocks has occurred primarily by the upward displacement of the former in the footwalls of master detachment faults (Fig. 3). For most faults, arguments about which wall or plate was active and which was passive are purely relative. This is not the case for the major detachment faults of the Colorado River extensional corridor. Using the earth's surface (Miocene and present) as a crude reference datum, the active plate during crustal extension and detachment faulting in the corridor south of Searchlight, Nevada, was the lower plate. Although some vertical movement of the upper plate cannot be discounted, the present juxtaposition of structural levels once up to 16 $\pm$ 4 km apart (in a vertical sense) has been accomplished primarily by the transport of lower plate rocks upward and southwestward out from beneath North America, the upper plate. This is an important point in understanding the nature of detachment faulting during crustal extension, and one agreed upon by most present workers in the area (cf. Luchitta and Suneson, 1981; Reynolds and Spencer, 1985; Howard and John, 1987). It makes no sense geologically to conclude that Miocene surficial deposits were actively downfaulted 16 km with respect to sea level or some absolute earth reference surface, e.g., the geoid, in order to now lie atop midcrustal rocks (unless one is willing to propose that the Whipples in Miocene time had an average elevation of 16,000 m compared with their present average of 700 to 800 m).

The duration of detachment faulting activity in the Colorado River extension corridor is imperfectly known; typically, it is estimated indirectly by studies of the depositional and structural history of upper-plate Tertiary strata. Throughout an area greater than 15,000 km^2, such strata have been block-faulted and rotated, and now generally dip southwestward at variable angles. Angular unconformities within detached Tertiary sections indicate multiple episodes of block faulting, stratal rotation, and presumably, detachment of upper-plate units. Although dips in rotated Tertiary strata generally increase northeastward across the Colorado River detachment terrane (Howard and others, 1982a; Howard and John, 1987), many exceptions point to significant complexities in temporal and spatial patterns of detachment faulting (Davis, 1986a). The analysis of stratal tilting is complicated by uncertainties in correlation and dating of many Tertiary units, although the regionally widespread Peach Springs Tuff (ca. 18.3 Ma, Glazner and others, 1986) provides a critical datum for such analysis. Otton (1982) concluded that Tertiary strata in west-central Arizona (Date Creek Basin) had undergone southwestward tilting of as much as 70° prior to 21 to 24 Ma. If Otton's age determinations are correct, then detachment faulting can be assumed to have also begun prior to this time. However, in most areas farther west (Fig. 2), the first major episode of tilting occurred either just before or just after deposition of the Peach Springs Tuff (Nielson and Glazner, 1986). Spencer (1985) reported that detachment-related tilting of Tertiary units in the

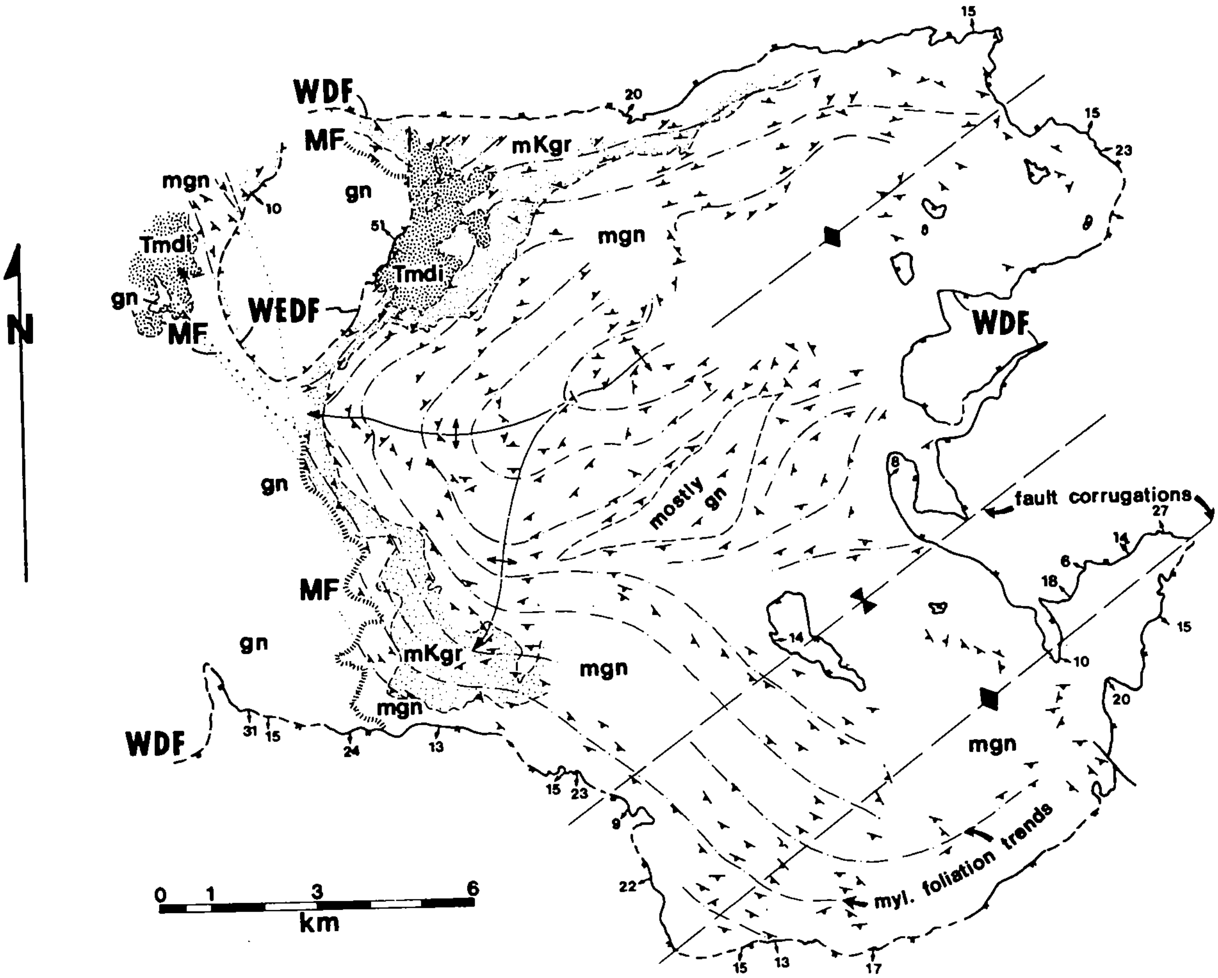


Figure 5. Geologic map of the lower plate of the Whipple detachment fault (WDF), central and eastern Whipple Mountains, California. The map emphasizes the contrasting geometry of the corrugated Whipple fault and folded mylonitic rocks below the Whipple mylonitic front (MF). Foliation trends are shown in mylonitic gneisses (mgn) and a mylonitized composite Cretaceous granitic pluton (mKgr, light stippled pattern), the latter with a sheetlike geometry (now folded). Foliation attitudes depict dip: open barbs, <25°; closed barbs, >25°. Nonmylonitized gneisses (gn) structurally overlie the mylonitic front, which is intruded by a composite Miocene diorite/gabbro pluton (Tmdi, heavy stippled pattern). War Eagle detachment fault (WEDF) offsets the mylonitic front ca. 4.5 km in a N30°E direction. The War Eagle allochthon lies within a "synformal" corrugation with the same trend. Displacement along the still-younger Whipple fault, upper plate with respect to lower plate, was N50°E and parallel to the major wavelike corrugations of the fault surface.

northern Sacramento Mountains ceased between 14 and 15 Ma. Collectively, available data from the Tertiary record indicate that detachment faulting had begun prior to 21 Ma and persisted until shortly after 15 Ma. Geologic and geochronologic relations in the Whipple Mountain region, amplified below, indicate that major detachment fault activity within presently exposed structural levels of the range occurred after 19 to 20 Ma. There is little direct evidence on the age of the detachment fault itself. One reset K-Ar whole-rock age for a cataclasized mylonitic gneiss lying 35 m below the Whipple fault is 15.3 ± 0.5 Ma (Davis and others, 1982).

GEOLOGY OF THE WHIPPLE MOUNTAINS

The geology of the Whipple Mountains is generally similar to that of several other ranges within the NLCR region, but the range, with a relief of one kilometer and many deep canyons that cut across the Whipple fault, provides some of the most spectacular exposures of detachment fault–related tectonics in the U.S. Cordillera (Fig. 6). Figure 4 is a diagrammatic block diagram, viewed to the north, of the central and eastern Whipple Mountains. It illustrates the major structural features of the range, including: (1) the Whipple Mountains antiform, a major foliation

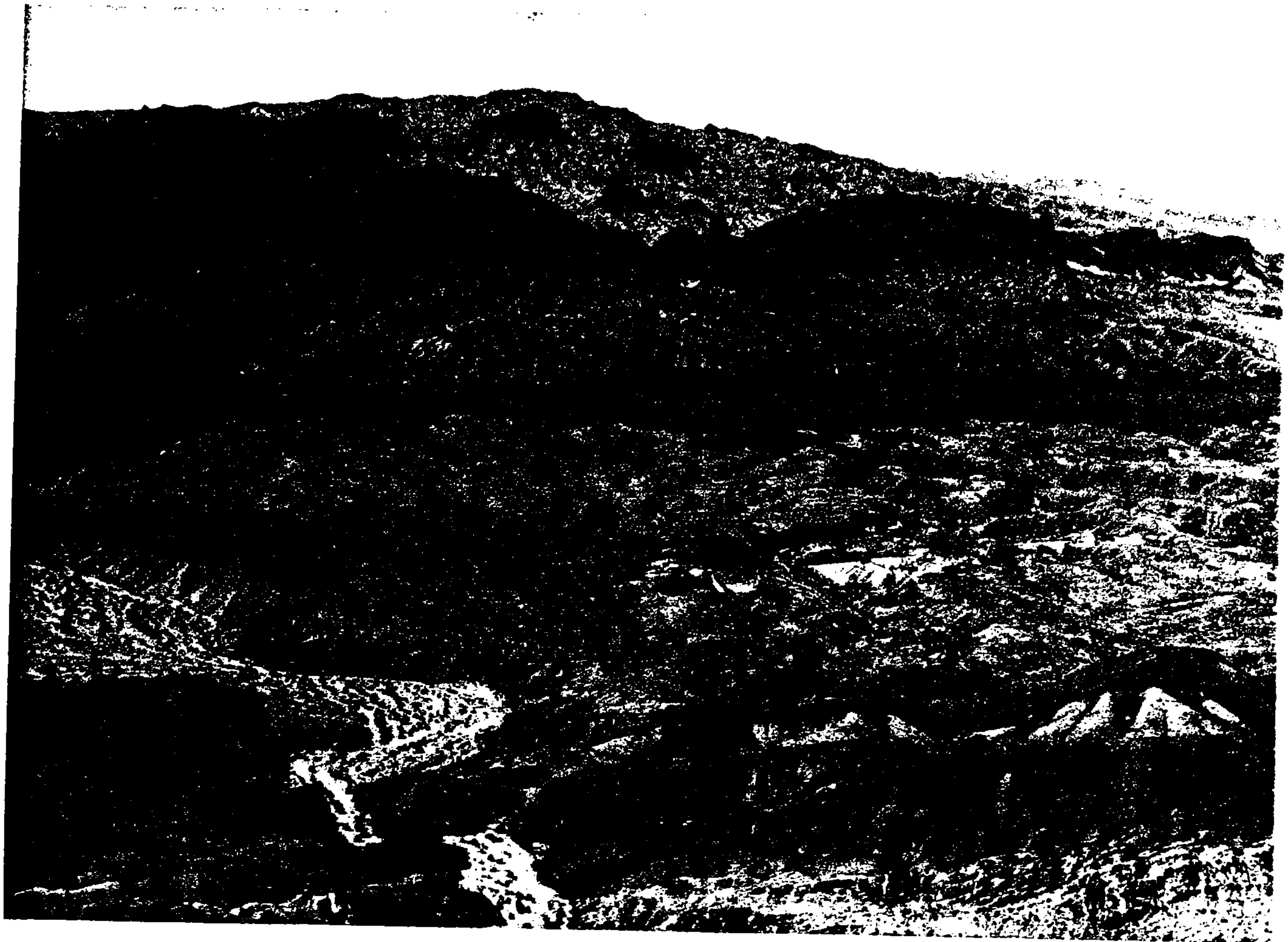
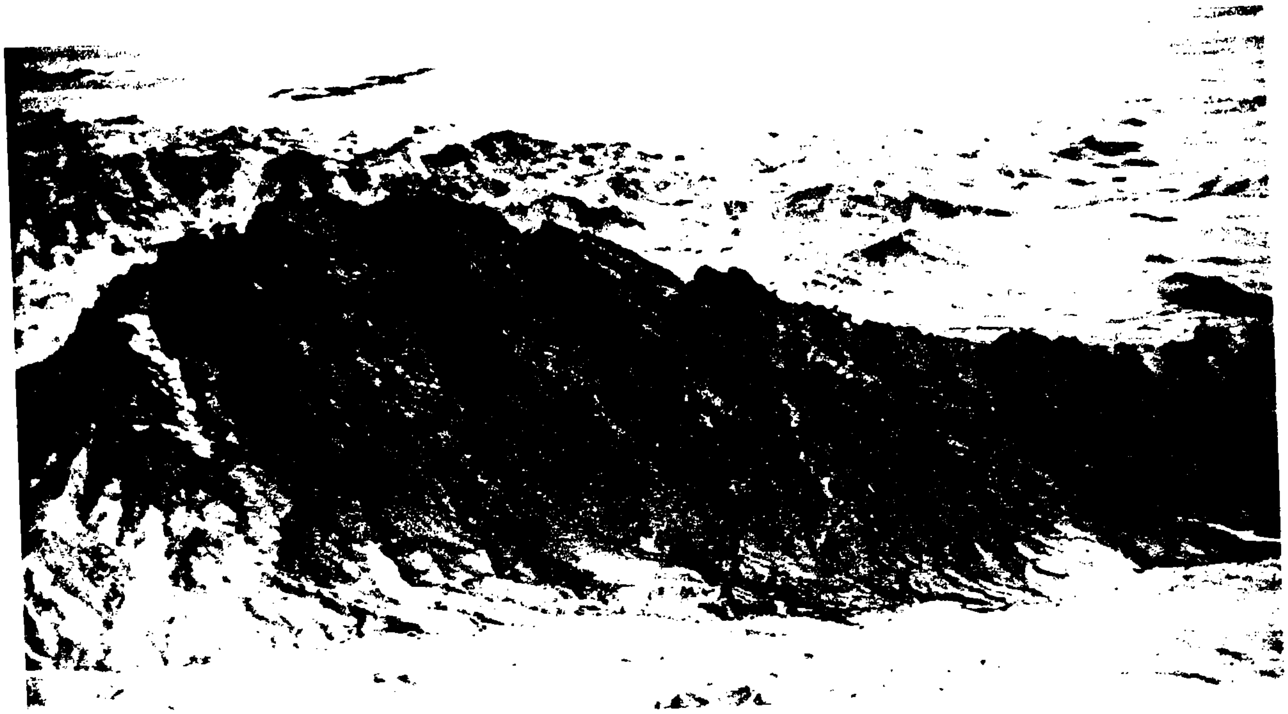


Figure 6. Low-angle aerial view of the Whipple Mountains, California, from westernmost Arizona; the Colorado River (Lake Havasu) along the boundary between the two states is in the middle ground. View is to the west. The crest of the Whipple Mountains rises 1.1 km above the river and its coincident with foliation-defined antiform (Figs. 4, 5) in the lower plate of the Whipple fault. The isolated dark hills above light-colored, lower-plate mylonitic gneisses are klippen above Whipple fault. The prominent contact between light-colored rocks just west of the Colorado River and darker, cliff-forming rocks is a nonconformity separating upper-plate Precambrian crystalline rocks from SW-tilted Oligo(?)-Miocene sedimentary and volcanic rocks.

arch in thick (>3.5 km) lower-plate mylonitic gneisses (mgn); (2) the southwest-dipping mylonitic front (MF), which separates lower-plate crystalline rocks (lpxln) from their structurally deeper, mylonitic counterparts; (3) the Whipple detachment fault (WDF) and the chloritic breccias (cbr) developed beneath it; and (4) the characteristic pattern of upper-plate faulting, a series of closely spaced (1 to 2 km), northeast-dipping normal faults that repeatedly offset and rotate to the southwest upper-plate Tertiary sedimentary and volcanic units (sv) and their largely Precambrian crystalline basement (upxln). The figure illustrates the strong discordance seen in the range between the Whipple fault and both lower-plate and upper-plate structures (Fig. 7).

Protoliths of the mylonitic rocks are not differentiated in Figure 4, but consist primarily of Precambrian gneiss (1.7 Ga), Cretaceous peraluminous (89 ± 3 Ma) and metaluminous (73 ± 3 Ma) plutons—most of them shallow dipping and sheetlike in form, and thin (<3 m), ubiquitous Oligo-Miocene (26 ± 5 Ma) biotite tonalite sills and dikes (Wright and others, 1986). Mylonitization of Whipple Mountains lower-plate rocks occurred under conditions of upper greenschist to lower amphiolite grade. Two feldspar and amphibole-plagioclase thermometries from Cretaceous plutons in the upper two-thirds of the mylonitic section yield temperatures increasing with depth from 458 ± 35 to 535 ± 44°C (Anderson, 1981, 1985, 1987; Anderson and Rowley, 1981; Davis and others, 1982).

Data on the depth of Tertiary mylonitization, and, by infer-



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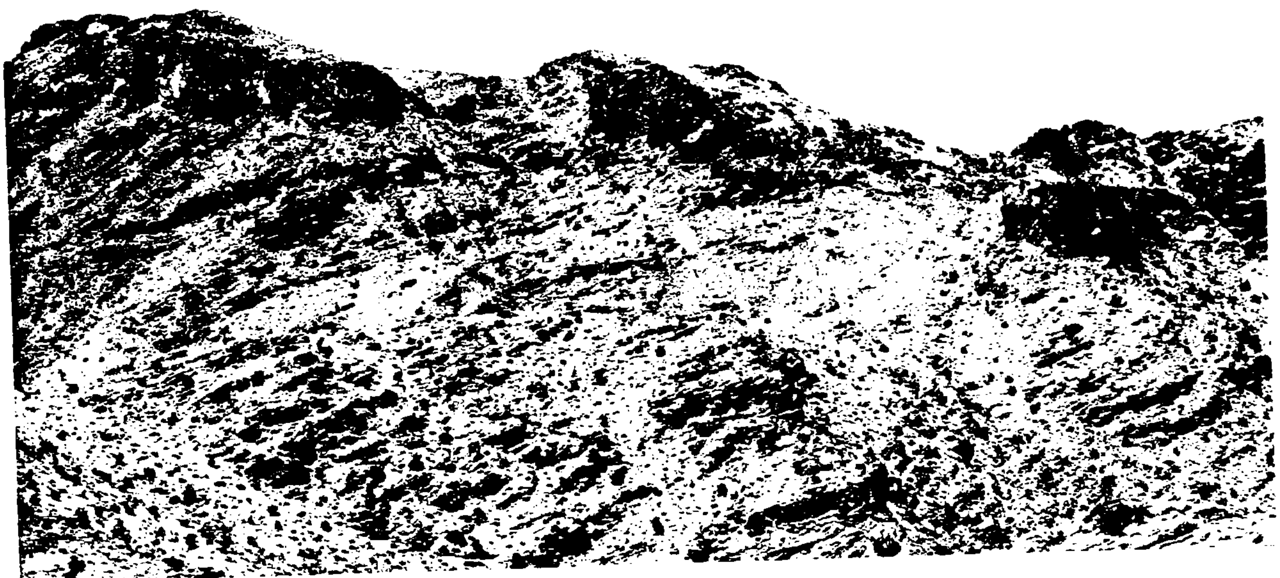

b

Figure 7. a, Structural discordance between steeply southwest-tilted Tertiary strata and the underlying, gently southwest-dipping Whipple detachment fault. This aerial view of Savahia Peak in the southwestern Whipple Mountains (courtesy of John Shelton) is to the southeast. Lower-plate Precambrian crystalline rocks at this locality are above the Whipple mylonitic front. Savahia Peak rises approximately 330 m above the lower terrain in the foreground. b, Structural discordance in the south-central Whipple Mountains between southwest-dipping mylonitic gneisses and a lower-plate detachment fault developed at the base of resistant, cliff-forming chloritic breccias. View is to the north-northwest. The Whipple fault (above the ground surface here) and chloritic breccias below it have a discordant relationship to lower-plate mylonitic foliation throughout most of the range (Figs. 4, 5, 9).

ence, the depth of rooting of detachment fault systems beneath the California-Arizona detachment terranes are still sparse. Anderson (1985, 1987, *in* Wright and others, 1986), using existing calibrations for GAR-PL-BIO-MU, MU-BIO-KSP-QZ, and KSP-PL equilibria, and the P-T limits of alumina solubility in hornblende, has estimated P-T conditions for mylonitization of granitic rocks in the Sacramento and Whipple Mountains, California, and the Santa Catalina Mountains north of Tuscon, Arizona (Fig. 1). Pressure estimates at the time of Tertiary mylonitization in the three areas are, respectively, 2.0, 4.4 ± 1.1, and a poorly constrained 3.8 ± 0.9 to 5.2 ± 1.0 kbar. The pressure range for the Whipple mylonitic rocks corresponds to a depth range of approximately 12 to 20 km; much shallower levels of formation indicated for the Sacramento mylonites appear to us as anomalous, but stem from significant differences in Whipple and Sacramento mylonite mineralogy (Anderson, 1987).

The mylonitic front in west-central portions of the range dips 25° to 50° to the west-southwest. At the front, structurally higher, northeast-striking gneisses, amphibolites, and foliated Cretaceous plutonic rocks (Fig. 4) become abruptly rotated, transposed, and mylonitized (cf. Davis and others, 1980, Figs. 15, 16; Davis, 1987). Prior to mylonitization and at levels below the present mylonitic front, steeply dipping Precambrian gneisses and amphibolites had been intruded by discordant, subhorizontal plutonic sheets of Cretaceous age. The plutons became preferential loci of shear strain during mylonitization, quite likely because of their favorable low-dipping geometry and their quartz-rich mineralogy. Panels or lenses of gneisses with relict steeply dipping foliation are locally preserved between plutonic sheets (cf. Davis and others, 1982, Fig. 2). These relict structural domains are as much as 1 km thick (Fig. 5, see domain labeled "mostly gn" below the mylonitic front). They terminate upward and downward into zones of rotation and transposition commonly associated with the highly foliated contacts of the plutonic sheets. More uniform rotation and transposition of the older gneisses has occurred in lower-plate areas below the front where Cretaceous plutonic sheets were not present and could not, therefore, act as preferential strain guides. Mylonitization did occur locally hundreds of meters above the front (cf. Davis and others, 1980, Fig. 16), especially, but not exclusively, in shallow- to moderate-dipping Tertiary dikes that were intruded during late stages of more pervasive mylonitization at deeper structural levels.

We have collected over 100 oriented samples of mylonitized crystalline rocks from several traverses that in the aggregate cross most of the 3.5+-km-thick mylonitic gneiss section. Fabric and microstructural analyses of these samples (Lister, unpublished manuscript) reveal a variety of kinematic indicators. Sense of shear in most samples could typically be determined by two or three indicators, including S-C fabric relations, oblique foliation in dynamically recrystallized quartz aggregates, asymmetric mica "fish," and asymmetric pressure shadows. These analyses demonstrate that approximately 65 percent of the samples collected formed by northeast-directed shear (higher structural levels relative to lower) parallel to the penetrative mylonitic lineation (N45 ± 10E) and to the sense and direction of transport along the Whipple detachment fault; 18 percent exhibit evidence for southwest-directed shear, and 17 percent were not kinematically definitive. We (Lister and Davis, 1983) have interpreted our data as indicating formation of the Whipple mylonitic rocks in a zone of intracrustal laminar flow. Within the zone, large shear strains took place by dynamic recrystallization (e.g., quartz, micas), with minor diffusional mass transfer (e.g., secondary quartz, K-feldspar), and cataclasis (plagioclase, hornblende, garnet). This conclusion supersedes the earlier interpretation by Davis (Davis and others, 1982) that mylonitization occurred during coaxial strain characterized by the flattening in a subhorizontal plane of preexisting thermally weakened rocks with attendant northeast-southwest extensional flow. However, evidence favoring such pure shear deformation is locally present within at least one of the relict structural domains bounded by well-developed mylonitic shear zones (cf. Davis and others, 1982, Fig. 6D, p. 418–419).

Mylonitization throughout the lower plate of the Whipple Mountains is of Tertiary age, possibly in part Oligocene and clearly of Early Miocene age. Davis and others (1980, 1982) concluded erroneously that the gneisses had formed during Cretaceous time. This age, correctly considered by some of our Arizona colleagues to be unreasonably old, based on their field and geochronologic studies in southwestern and south-central Arizona (cf. G. H. Davis, 1980; Rehrig and Reynolds, 1980; Reynolds, 1982), was based on several lines of evidence, among them: (1) concordant biotite K-Ar (78.5 ± 5.5 Ma) and sphene fission track (82.9 ± 3.0 Ma) ages from a mylonitic plutonic gneiss found as clasts in an upper-plate Tertiary debris flow; (2) geologic relations and isotopic age data indicating Cretaceous mylonitization in areas surrounding the Whipple Mountains, e.g., the Iron and Chemehuevi Mountains, California (Miller and others, 1981; John, 1982, 1987, respectively); and (3) our early interpretation that the Cretaceous plutonic sheets (89 Ma) had been intruded synkinematically during mylonitization. The latter interpretation was subsequently invalidated by finding a steep, premylonitic (and nonmylonitic) foliation in at least two of the 89-Ma plutons.

A Tertiary age for Whipple mylonitization has now been thoroughly documented by recent geochronologic and field studies, although the time of its inception has not been closely constrained. Fine-grained, porphyritic (biotite, plagioclase) tonalite sills and dikes occur throughout the thick section of mylonitic gneiss. Field relations indicate that their intrusion was synkinematic to mylonitization. For example, some biotite tonalite dikes were intruded, mylonitized with their country rocks, folded, and then intruded by identical (in the field) planar dikes along the axial surfaces of the folds. The younger dikes were also mylonitized and have a northeast-southwest–stretching lineation that is parallel to that in the older dikes. Zircons from one of the biotite tonalite dikes have been dated as 26 ± 5 Ma, an Oligo-Miocene age (Wright and others, 1986).

The cessation of lower-plate mylonitization is now rather well constrained. Tertiary dikes of the Chambers Well dike swarm (Davis and others, 1982, Figs. 1, 5) were intruded across the mylonitic front. Below the front, andesite and dacite dikes of the swarm always exhibit mylonitic fabrics, but younger diabase dikes that intrude them are typically not mylonitized or display only schistose fabrics within their chilled marginal zones. One such lower-plate mafic dike, collected above the mylonitic front, yields an undisturbed ^{40}Ar-^{39}Ar hornblende age of 21.5 ± 0.7 Ma (E. DeWitt, written communication, 1984). All Chambers Well dikes and the mylonitic front are intruded by a postkinematic (i.e., nonmylonitic) hornblende quartz-diorite to olivine-clinopyroxene gabbro pluton (Tmdi, Fig. 5). Zircons from the dominant quartz diorite phase yield concordant U-Pb ages of 19 ± 2 Ma (Wright and others, 1986). Actinolite hornblende from the same body has a $^{40}Ar/^{39}Ar$ plateau age of 19.8 ± 0.1 Ma (E. DeWitt, written communication, 1986). Sparse, thin (<0.25 m) aplite dikes that intrude the diorite do have mylonitic fabrics, but the field relations described here indicate that the major episode of lower-plate mylonitization had ceased by 21.5 Ma. All lower-plate rocks described here—the mylonites, all dikes of the Chambers Well swarm, and the diorite/gabbro pluton which intrudes them—are truncated discordantly by the Whipple detachment fault. Its age in the central portion of the range is, therefore, younger than 19 to 20 Ma.

DETACHMENT ZONE ROCKS AND A MODEL FOR DETACHMENT FAULT ORIGIN

Physical descriptions of detachment faults in the southern U.S. Cordillera have been published by many authors, most notably in the Geological Society of America Memoir on *Cordillera Metamorphic Complexes* (Crittenden and others, 1980) and in the volume of papers dealing with the Colorado River region mentioned earlier (Frost and Martin, 1982). Figure 8 illustrates some of the characteristics of such faults. They are typically very planar, but characteristically exhibit domal and basinal or antiformal and synformal geometries on the scale of individual mountain ranges. Fault surfaces, although locally polished, generally exhibit a dull orangish-brown to reddish-brown limonitic(?) patina. This surface is often underlain by dark-colored (black, brown, dark green) layer(s) of flinty cataclasite or microbreccia, usually containing angular, nonoriented porphyroclasts of lower-plate rocks. The layer is generally less than 0.2 m thick and forms a resistant ledge across the topography (cf. Davis and others, 1980, Fig. 24). Phillips (1982) has studied the microbreccia by scanning electron microscopy (SEM) and observed that its matrix is composed of very small (average, 0.004–0.006 mm), equant, interlocking quartz and feldspar grains showing little signs of deformation. Field relationships, especially injection veins into underlying breccias, suggest that these cataclasites flowed during their formation. Phillips suggests that these random fabric fault rocks might have formed initially by cataclastic flow, but that structural superplasticity became the dominant deformational mechanism after grain size had been reduced. Further SEM studies of the cataclasite (H. Green and G. Lister, in progress) reveal evidence of multiple phases of silicification subsequent to repeated episodes of cataclastic flow.

The detachment fault microbreccias and below them sparse, thin (<3 mm) layers resembling pseudotachylite (cf. Fig. 2a of Davis and others, 1986) are overprinted across much thicker (5 to 100 m) "chloritic breccias" (Fig. 4, cbr) which have been derived from lower-plate crystalline rocks. Their development is clearly tied to the detachment fault, since the breccias always lie directly below it and are found both above and below the mylonitic front. These breccias show their thickest development when developed across mylonitic protoliths. Upper-plate rocks have never been observed to be included within the chloritic breccias. Rocks within the breccia zone have variably experienced the effects of locally profound shearing (in extreme development to aphanitic, chert-like cataclasites), in situ shattering, southwest-tilting along northeast-dipping normal faults, and pervasive alteration. The alteration that accompanied shearing and cataclasis is probably lower greenschist in grade, involving as it does the growth of chlorite, epidote, and sericite. The bottom of the chloritic breccia zone is a low-angle fault throughout much of the Whipple Mountains (Fig. 7b), although in some areas the effects of shearing, brecciation, and alteration simply die out downward. Normal faults that cut the breccias are of more than one generation. Some flatten with listric geometries into the fault at the base of the breccias. Other faults cut both the breccias and their underlying low-angle fault, but are themselves truncated upward by the Whipple fault with its associated microbreccias (Fig. 9).

Davis and others (1983, 1986) have interpreted the progressively overprinted sequence of premylonitic crystalline rocks, mylonitic gneisses, chloritic breccias, and flinty cataclasites or microbreccias as indicating that deep-seated rocks have been transported upward from crustal depths of 10 to 15 km or greater along evolving shallow-dipping shear zones (Fig. 3) of the general type envisioned by Wernicke (1981, 1985), Reynolds (1982), and G. H. Davis (1983). The deformational behavior of rocks now exposed in the footwalls of major detachment faults of the NLCR region has changed with time as follows: (1) penetrative deformation (mylonitization) of preexisting crystalline rocks in low-dipping zones of intracrustal laminar flow; (2) upward passage of mylonitic gneisses in the footwalls of detachment fault zones with retrograde formation of chloritic breccias in zones of intense shearing and cataclasis; (3) development of pseudotachylite along narrow (<3 mm) layers cross-cutting the breccias, probably as fault-generated melts formed during intervals of seismic slip; and (4) late-stage formation of discrete detachment faults with associated microbreccias, and, locally, younger fault gouge. These relations suggest progressive deformation of footwall rocks at decreasing crustal depths and temperatures during a Tertiary period of unknown duration. During crustal extension the originally deep lower-plate rock assemblages of the detachment terrane are drawn upward and outward from beneath the brittlely extending upper plate. Lower-plate rocks become exposed at the surface through variable combinations of regional and subregional warping, tectonic denudation, and erosion

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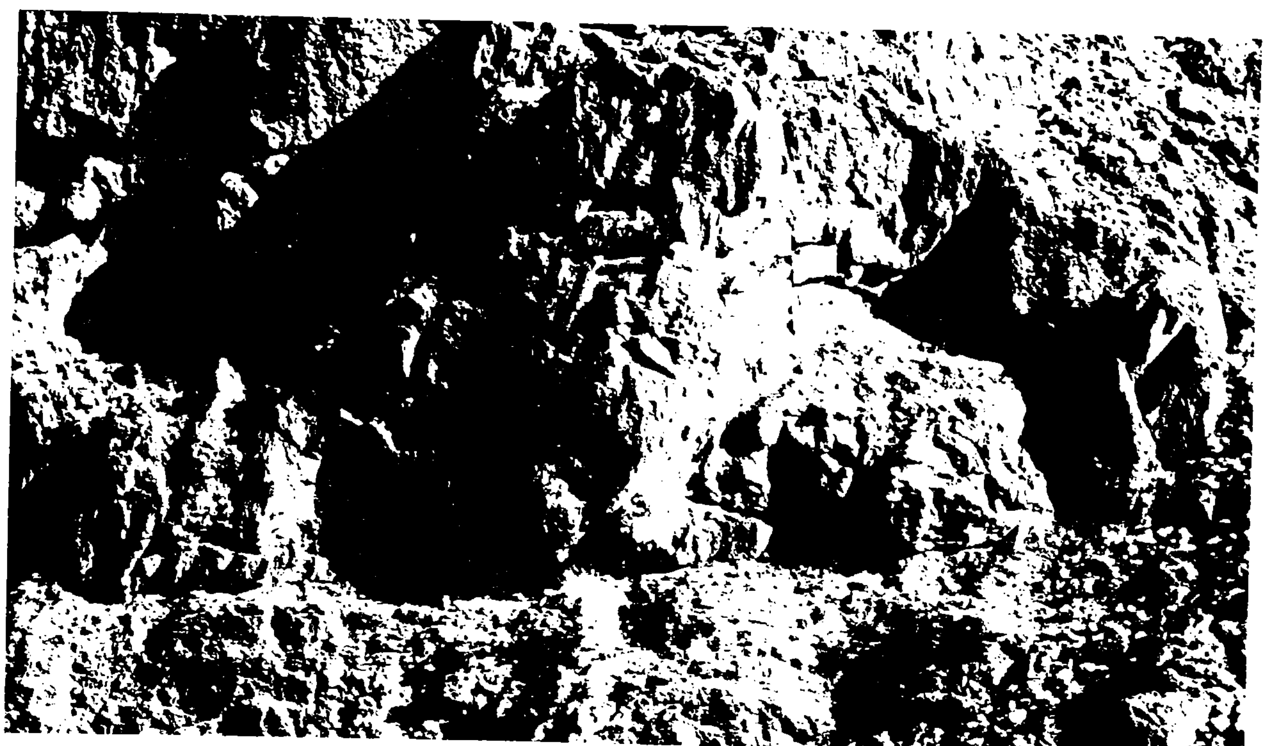

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Figure 8. a, Exhumed Whipple detachment fault surface, south-central Whipple Mountains. Cholla cactus in a pained Gordon Lister's left hand provides scale. The detachment fault surface is underlain by a resistant microbreccia at this locality. b, Exposure of Whipple detachment fault on south side, Whipple Wash. Very planar, gently dipping fault separates lower-plate chloritic breccias from subhorizontal upper-plate Tertiary sedimentary and volcanic section (Tmvs, Fig. 13). Neither the prominent northeast-dipping normal fault (on right) or antithetic southwest-dipping normal fault (to left) flatten into the detachment fault. Height of exposure is approximately 35 m.

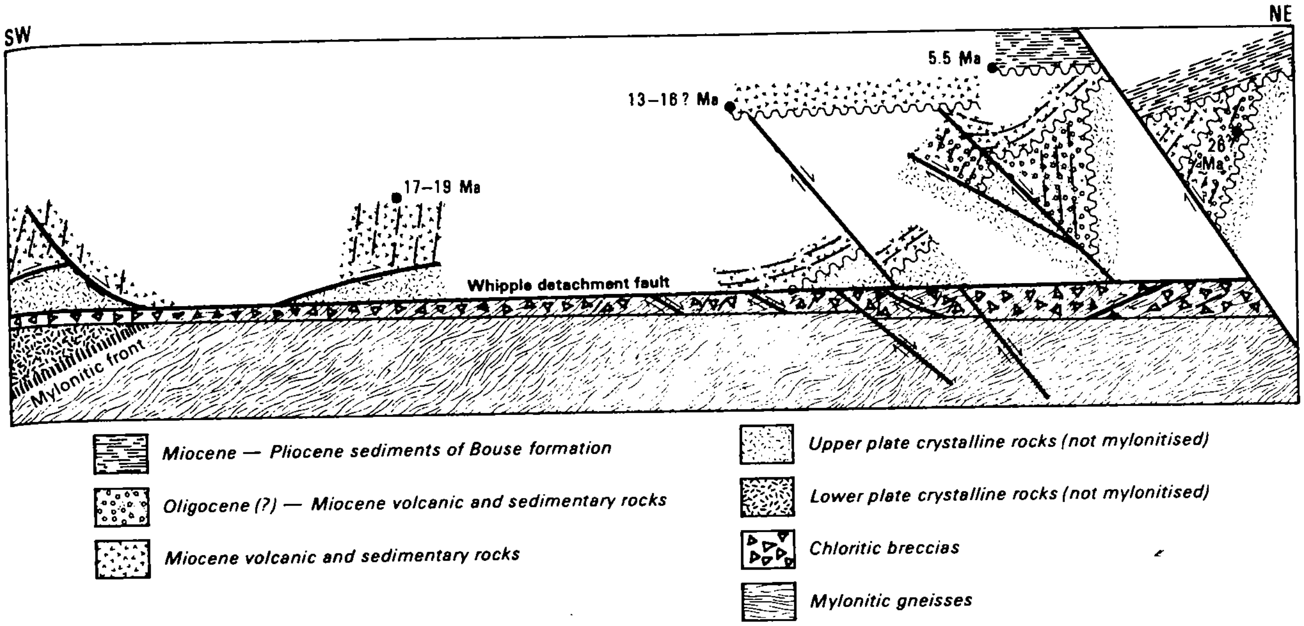

Figure 9. Diagrammatic northeast-southwest cross section through the Whipple Mountains, southeastern California. Section illustrates late age of Whipple detachment fault (WDF) with respect to multiple generations of preexisting upper- and lower-plate structures. Ages are given for most critical stratigraphic units. Listric upper-plate fault (far left) cuts shallow-dipping upper-plate detachment faults and is presumably younger than steeper upper-plate faults (on right). These faults are interpreted as having once been listric with respect to earlier, structurally deeper generations of Whipple detachment faults (see text). The Havasu Springs fault on the far right rotates hanging wall strata of the Mio-Pliocene Bouse Formation and is believed to be a post-detachment normal fault of Basin-Range type. Figure 7b illustrates the angular relationship, shown on this section, between southwest-dipping mylonitic gneisses and the cross-cutting detachment fault at the base of the chloritic breccias.

(Fig. 3). This model accounts for the abrupt contrast in metamorphic grade and deformational history of upper- and lower-plate rocks now juxtaposed across the major detachment faults.

The evolving shear zone model for crustal extension in the NLCR region receives support from geochronologic studies of lower-plate rocks in the Whipple Mountains, and from field studies that indicate the timing and magnitude of displacement along the Whipple fault. Geochronologic studies strongly support our tectonic model by confirming that Miocene mylonitic gneisses experienced very rapid cooling, between roughly 19 to 20 and 18 Ma, presumably during rapid uplift and tectonic denudation. This conclusion was first reached by Dokka and Lingrey (1979) from fission track dating of mylonitic rocks collected from just below the Whipple fault. Five age determinations (three from zircons, one each using apatite and sphene) from three rock samples yielded concordant ages that varied from 17.9 to 20.4 Ma with an overlap of error bars between 18.4 and 19.5 Ma. Because the three analyzed mineral species have widely disparate fission track retention temperatures, their concordant ages led Dokka and Lingrey to conclude that the mylonitic gneisses had experienced a significant temperature drop (>80°, <220°C) between 18 and 20 Ma. $^{40}Ar/^{39}Ar$ age-spectrum dating of Whipple mylonitic rocks by E. DeWitt and J. Sutter (written communication, 1986) offers impressive confirmation of the Dokka and Lingrey study. Tertiary mylonitization (26 ± 5 Ma) of a Cretaceous hornblende quartz diorite in the Whipple Mountains occurred at a temperature of 535° ± 44°C (Anderson, 1987). By 19.2 ± 0.2 Ma, neomineralized hornblende (formed during mylonitization of the diorite) had cooled below about 450°C. Neomineralized muscovite from a closely adjacent wall rock mylonite has a plateau age of 18.0 ± 0.1 Ma, indicating that the mylonitic assemblage "had cooled below about 275°C only 1 m.y. after it was at more than 450°C" (E. DeWitt, written communication, 1986). Orthoclase from a structurally higher mylonitized granitic rock collected elsewhere in the range has a near plateau age of 18.5 Ma. This age indicates cooling of the rock below 150°C, only about 1 m.y. after the hornblende in deeper rocks had cooled below 450°C. Lower plate mylonitic rocks reached the earth's surface during detachment faulting. Tertiary mylonitic clasts are present in some tilted upper-plate Miocene fanglomerates that are older than 16 Ma.

Rapid upward transport of mylonitic gneisses in the lower plate of the Whipple detachment fault system, suggested by the rapid cooling of these rocks, is supported by field studies indicating high collective rates of slip along Whipple system faults after 20 Ma. As mentioned above, the Whipple fault truncates a lower-plate assemblage of rocks near the mylonitic front that includes the Chambers Well dike swarm and a somewhat younger (19.8 Ma), cross-cutting hornblende quartz diorite pluton (Fig. 5). An upper-plate dike swarm, which is almost certainly the offset equivalent of the Chambers Well swarm, is exposed in the Mohave Mountains to the northeast (Fig. 2, Nakata, 1982; Howard and others, 1982a). If this correlation of dike swarms is valid, at least 40 km of offset (horizontal component) has occurred between rocks of the two ranges after approximately 20 Ma (the age of the diorite pluton). If a constant rate of slip is assigned to the Whipple fault system between 20 and 15 Ma, a slip rate (horizontal component) of 0.8 cm/yr (8 km/m.y.) is indicated. However, much higher slip rates are probable for the period 18 to 20 Ma in order to account for the rapid cooling of footwall mylonites during fault-controlled uplift within that time interval. Most, although not all, major tilting of Tertiary strata in the Whipple-Mohave mountains region occurred just before or soon after deposition of the Miocene Peach Springs Tuff (mean sanidine K-Ar age of 18.3 Ma, Glazner and others, 1986; Nielson and Glazner, 1986; Davis, 1986).

FOOTWALL MYLONITIC GNEISSES

The temporal and spatial relations between detachment faults and the mylonitic gneisses exposed below them need amplification. Structural relations throughout the detachment terranes of the southwestern Cordillera indicate kinematic continuity between the Tertiary period of deformation in which the now-exposed mylonitic gneisses formed and the period during which detachment faulting occurred (Davis and others, 1983, 1986). At any given time during their development, major Tertiary detachment faults, such as the Whipple fault, were probably continuous down-dip into progressively wider and deeper zones of brecciation, shearing, and mylonitization (e.g., detachment system 1, Fig. 3; cf. Wernicke, 1981; Reynolds, 1982; G. H. Davis, 1983; Davis and others, 1983, 1986). But lower-plate mylonitic gneisses in the Whipple Mountains are, as one example, somewhat older than the Whipple fault below which they are now exposed. This age relationship has been documented by geochronologic studies reviewed above and is demonstrated by the angular discordance of the Whipple mylonitic front (and most mylonitic foliation below it) with the overlying Whipple fault (Figs. 4, 7b, 9; Davis and others, 1980, 1982, Figs. 1, 2). The Whipple mylonitic front leaves the detachment fault and dips below the ground surface to the southwest (Fig. 4). This is a geometric necessity since only directly beneath and near the capturing detachment fault are footwall rocks, including those at the mylonitic front, elevated to surface or near-surface structural levels (Fig. 3). It appears likely that the front is "seen" in CALCRUST seismic reflection profiles at a depth from 3 to 4 sec, respectively, 16 to 60 km west of the surface trace of the mylonitic front (Davis, 1986b; Okaya and Frost, 1986).

From the relationships described above, it is probable that the Whipple fault did not exist at the time of the penetrative ductile deformation that formed the mylonitic gneisses now exposed in its lower plate. Two extensional tectonic settings are possible for these pre–Whipple fault mylonitic gneisses. One is that they formed at depth along the roots of a more westerly detachment fault system (e.g., detachment system 1, Fig. 3). If so, these early Miocene mylonitic gneisses were presumably crosscut and "captured" by the younger, most easterly Whipple detachment fault and then carried rapidly upward (ca. 18 to 20 Ma) in the footwall of that fault (e.g., detachment system 2, Fig. 3). Dokka and Woodburne (1986) and Dokka and Baksi (1986) have documented the existence of a major detachment fault system in the central Mojave region to the west (Fig. 1). This system was active between about 24 and 20 Ma and may have been kinematically linked with the lower Colorado River detachment system (R. K. Dokka, 1986, personal communication). An alternative tectonic setting is that the mylonitic gneisses formed in an intracrustal zone of noncoaxial laminar flow (a zone of midcrustal delamination) without direct or obvious connection to structural levels now exposed in the southwestern Cordillera (Davis, 1987).

Mylonitic gneisses are not present beneath all major Cordilleran detachment faults, nor should they be. According to the crustal shear zone model described above, major detachment faults lacking lower-plate mylonitic rocks probably either did not have displacements large enough to carry deep-seated mylonitic gneisses to upper-crustal levels, or alternatively (and less likely), the faults were generated at crustal levels above those where mylonitic gneisses could have formed during Tertiary crustal extension (Fig. 3). The Chemehuevi fault in the NLCR mountain range of that name is a case in point of a major detachment fault (displacement >8 km; John, 1987) that lacks Tertiary mylonites in its footwall; Mesozoic mylonitic rocks are present, but are intruded by a late kinematic and postkinematic plutonic suite, a younger phase of which yields a 64-Ma biotite K-Ar cooling age (John, 1987). John has correlated the Chemehuevi fault, which carries southwest-tilted Tertiary strata in its upper plate, with the Whipple and Sacramento faults in areas to the south and north, respectively. Alternatively, we consider the Chemehuevi fault, lacking Tertiary mylonites in its footwall, to lie structurally above the Whipple fault and to have a much smaller displacement. It is possible that the Chemehuevi fault is an upper-plate splay off the Whipple fault and that both northeast-dipping faults merge at depth. Because Tertiary mylonitic gneisses are not exposed in the Chemehuevi Mountains. Howard and John (1987) conclude that in this range "the middle Tertiary crustal extension was not accommodated in a major way by either igneous intrusion or ductile distension at any crustal levels now exposed." We agree with this conclusion for the Chemehuevi Mountains, but believe that it is not applicable to those ranges in the lower Colorado River detachment terrane where deeper levels of Tertiary structural development are exposed (e.g., Sacramento, Whipple, Buckskin mountains with their mylonitic lower plates, Fig. 2).

The geometry at depth of detachment zones and their relationship to spatially associated and broadly coeval mylonitic gneisses is uncertain, and a matter of much disparate opinion among Cordilleran workers. We will not attempt here a discussion of tectonic models that are alternatives to inclined crustal shear zone models, e.g., the coaxial ("pure shear") strain model proposed by Miller and others (1983) for the Snake Range of Nevada; such discussion is undertaken elsewhere (Lister and Davis, 1987). Wernicke (1981, 1985) has postulated that detachment zones may cross the entire lithosphere and root into underlying asthenospheric rocks. We believe, however, that the common occurrence of mylonitic gneisses in the footwalls of major detachment faults argues for rooting of the faults into those lower upper crustal or midcrustal levels (see below) where such gneisses can form. Mesoscopic and microscopic analyses of mylonitic gneisses from several "core complexes" in Arizona, California, and Nevada strongly indicate that they have formed in shallow-dipping ductile shear zones, with a sense of shear the same as the relative sense of displacement inferred for spatially associated detachment faults (e.g., Lister and Davis, 1983). S-C mylonite types I and II (Lister and Snoke, 1984) are common throughout the mylonitic sequences. Kinematic complexities do exist (e.g., subordinate antithetic shear zones, mylonite-bounded relict lenses exhibiting internal coaxial strain effects), but they do not appear to detract from the general conclusion that the movement picture inferred for most mylonitic gneiss terranes is compatible with simple shear tectonic models. This conclusion does not negate the possibility that elements of pure shear were involved in the evolution of some detachment fault/mylonitic gneiss terranes, especially at mid- to lower crustal depths (e.g., Hamilton, 1982; Miller and others, 1983; Lee and others, 1987; Lister and Davis, 1987).

COMPLEXITIES IN THE EVOLVING GEOMETRY OF DETACHMENT FAULTS

As described above in the evolving shear zone model (Fig. 3), the lower plate of detachment zones is drawn upward and out from beneath the upper plate as high levels of the lithosphere are distended. Figure 3 is oversimplified, however, because it suggests that the initial detachment fault that propagated upward from midcrustal levels remains operative throughout the entire history of lithosphere extension. Field relations in the northern lower Colorado River (NLCR) region indicate that this is not the case and that the geometric evolution of detachment fault zones is quite complex. For example, the Whipple detachment fault, the most spectacular of the structural elements in the Colorado River extensional corridor, is unequivocally the youngest major structure of the region in which it is developed (with the exception of postdetachment antiformal warping along the regional northwest-southeast trend). The fault truncates discordantly steeper upper- and lower-plate normal faults that formed during earlier phases of detachment tectonics (Fig. 9), a relationship that

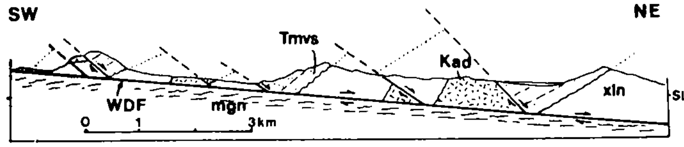

Figure 10. Geologic cross section through upper plate of Whipple detachment fault (WDF), eastern Whipple Mountains (from Frost, *in* Anderson and Frost, 1981). Gene Canyon reservoir lies above easternmost exposures of southwest-dipping Miocene sedimentary and volcanic strata (Tmvs). Other upper-plate units include nonmylonitized, largely Precambrian metamorphic and plutonic rocks (xln) and disrupted portions of a Cretaceous adamellite pluton (Kad). Lower-plate chloritic breccias are present but are not illustrated. Attempts to balance this section, e.g., by realigning the multiply offset Miocene nonconformity below the Tertiary section reveals that large volumes of former upper-plate rocks are now missing. Domino-style rotation above the present Whipple fault of the northeast-tilted fault blocks seen here would produce such major space problems that this mode of rotation appears to be geometrically precluded. No vertical exaggeration.

Davis, his co-workers, and other investigators did not fully appreciate in earlier phases of Whipple studies.

There was an earlier tendency on our parts to regard the internal fault structures of the two juxtaposed plates as somehow developing wholly with the plate in which they are now found. For example, at one stage of our studies it was assumed that upper-plate normal faults were originally planar, that they had originally terminated downward at the Whipple detachment fault and that rotation of tilted Tertiary strata in upper-plate blocks bounded by them had occurred domino-style during displacement on the underlying Whipple fault (cf. Frost, 1980, 1984; Gross and Hillemeyer, 1982). The principal problem with treating each plate as a self-contained, evolving unit is that cross-sections, through the Whipple upper plate (e.g., Figs. 10, 11) cannot be balanced, i.e., palinspastically restored to initial geologic configurations. Restoration of upper-plate geologic relations prior to upper-plate normal faulting clearly indicates that we are now missing rocks that were once contiguous with those still preserved. Such geometries require that most steep faults above the Whipple detachment formed prior to development of the presently observed Whipple fault across them. If it is assumed that these early, steeper normal faults developed above a coeval detachment fault of the Whipple system—an assumption supported by geometric and kinematic data—then the later truncation of these faults by a structurally higher detachment fault (the presently exposed Whipple fault) requires a transfer of some former upper-plate rocks to a lower-plate position. This is the phenomenon of excisement. It and the opposite phenomenon of incisement—the structural transfer of lower-plate rocks to an upper-plate position—have been important processes in the evo-

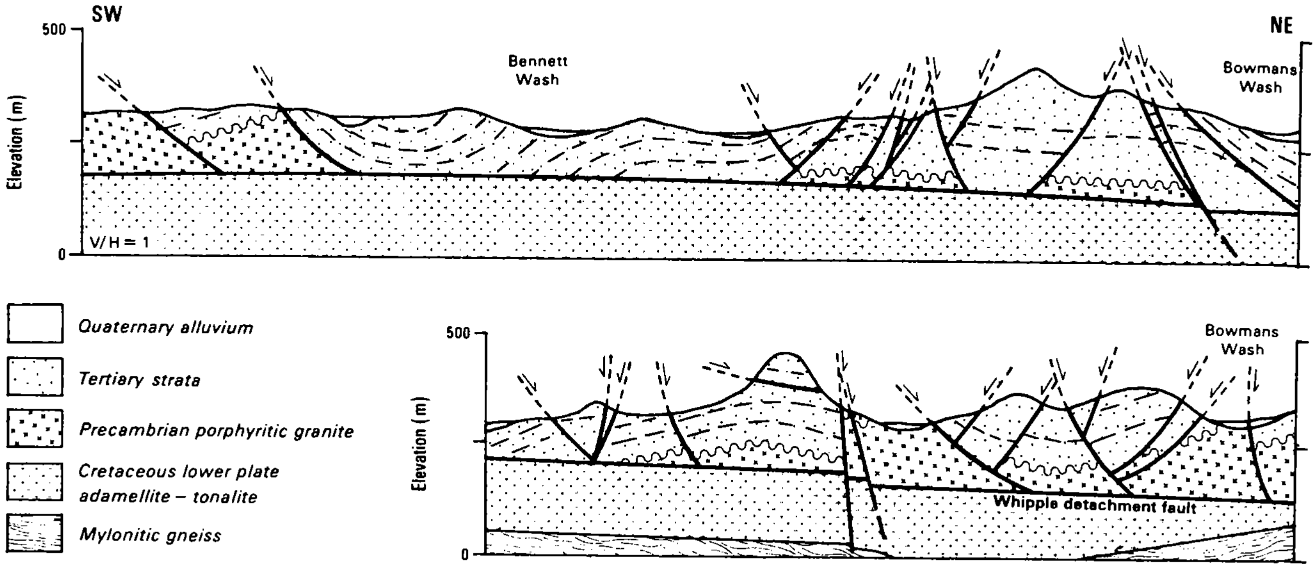

Figure 11. Geologic cross sections of the Whipple detachment fault, southern flank of Whipple Mountains (from Thurn, 1983). These parallel sections lie just southwest of the section of Figure 10. Upper-plate Tertiary strata are generally shallow-dipping to horizontal, broadly folded, and offset by both northeast- and southwest-dipping normal faults. Domino-style faulting did not occur in this area, and Tertiary strata have clearly not been rotated downward into contact with the Whipple fault. Cross sections above the Whipple fault cannot be balanced, suggesting strongly that the Whipple fault postdates upper-plate structure in this area. The Cretaceous lower-plate pluton shares a common mylonitic foliation with the mylonitic gneiss. Lower-plate chloritic breccias developed across the pluton are not illustrated in the sections.

lution of the northern lower Colorado River region (Lister and others, 1984; Dunn and others, 1986).

In Figure 12 we attempt to illustrate how warping of an initially more planar detachment fault during its operation can produce geometric complexities in the development of coeval upper-plate faults. The phenomenon of isostatically induced warping within the Colorado River extensional corridor has been discussed by Howard and others (1982a), and Spencer (1982, 1984, 1985). As an initially rather planar and uniformly dipping detachment fault becomes more and more curved as the lower plate warps upward, a time is reached when mechanical impedence against the continued functioning of the curved fault becomes so great that a new, more planar detachment develops as a higher splay off the preexisting fault (Fig. 12). As displacement (translation) is transferred to the younger fault, excised upper-plate rocks are carried away in the lower plate from their former upper-plate counterparts. Because of this transfer, cross sections drawn through the remaining, lessened portion of the upper plate can no longer be balanced. Figure 12 illustrates our belief that most upper-plate normal faults originally had listric geometries, and flattened into active underlying detachments. Their present more planar configurations are interpreted as the consequence of excisement of their flattened lower portions during development of higher, younger detachment splays.

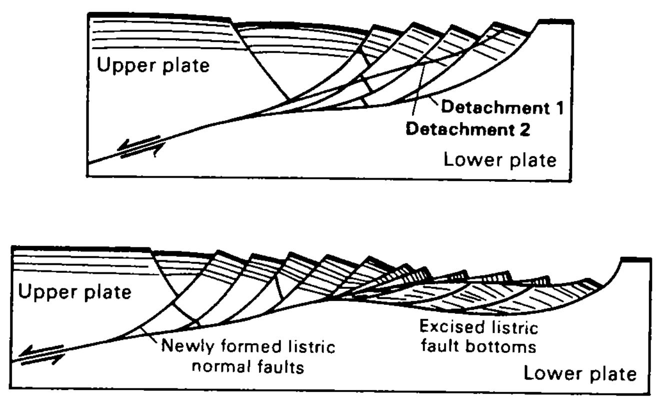

Figure 12. Hypothetic evolution of upper structural levels of an evolving detachment fault zone. Upper diagram shows a major fault splay (detachment 2) that has branched upward from the previously active detachment (1). In the lower diagram, this splay has become the active detachment, and excisement of the lower portion of the former upper plate has occurred. These excised rocks are now transferred to the lower plate of the younger, active fault. Previously active listric faults (phase 1) are abruptly terminated by the younger detachment. Younger listric normal faults now form with respect to the new detachment surface.

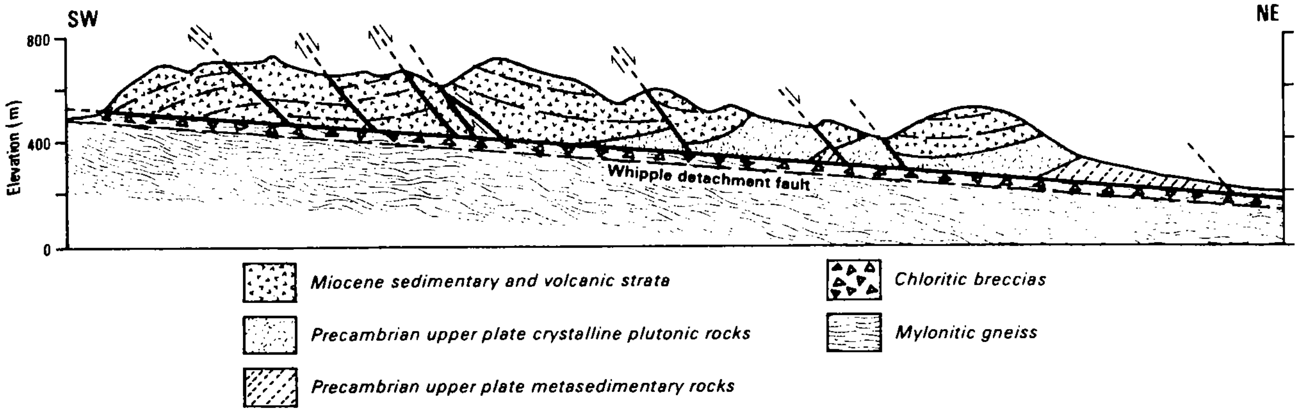

Figure 13. Geologic cross-section from Frost (1980) through the Whipple Wash gorge area, eastern Whipple Mountains, where the geometry of upper-plate faulting is well exposed. Numerous planar normal faults that terminate at, but do not cut the Whipple detachment fault (WDF) are shown (cf. Fig. 8b). No vertical exaggeration.

Domino faulting—the synchronous rotation of planar faults and crustal blocks between them—was described by Thompson (1960) and has been widely considered to be an important mechanism for continental extension (e.g., Wernicke and Burchfiel, 1982; Miller and others, 1983; Jackson and McKenzie, 1983). Two major geometric problems exist with domino mechanisms for upper-plate extension: (1) there is a space problem at the base of the blocks as they tilt, and it becomes progressively aggravated as the width of fault-bounded blocks increases; and (2) domino-faulting may allow fault blocks to rotate, but because abrupt changes in tilt angles between blocks would produce major fit problems, adjoining fault blocks cannot differ very much in their orientation or in the dip of strata contained within them. We do not observe major space problems in the basal part of the Whipple upper plate where it is cut by multiple, more-or-less planar faults, and we do see adjacent fault-bounded panels within which the dip of strata varies markedly from one panel to the other (cf. Fig. 9). Accordingly, we are skeptical of the importance of domino faulting in the lower Colorado River detachment terrane as the primary feature allowing extension of the upper plate, favoring instead the interpretation of primary listric faults offered above and developed further below.

We propose that a basic error made by previous workers in detachment terranes (e.g., in the lower Colorado River region by Gross and Hillemeyer, 1982; Howard and others, 1982a,b; and, until recently, ourselves, cf. Davis and others, 1980) is *the assumption that a given fault, e.g., the Whipple detachment fault, is everywhere the same tectonic surface of the same age.* Geologic observations tell us that this assumption is suspect. For example, portions of the Whipple detachment fault in the Whipple Mountains are underlain by impressively resistant ledges (typically <1 m) of black, flinty cataclasite or "microbreccia" (cf. Phillips, 1982), whereas in other areas the fault is underlain only by a zone several meters thick of poorly consolidated fault gouge. In some portions of the Whipple Mountains, Miocene strata dip vertically into the subhorizontal Whipple fault, but elsewhere only a few kilometers distant, strata of the same age rest subhorizontally (and tectonically) directly on the Whipple fault over large areas (cf. Fig. 11). How can these variable geometric and structural relations be accounted for?

A specific example, drawn from studies of the Whipple detachment fault, can be used to illustrate some of the geometric complexities that occur in the evolution of detachment zones. Figure 13 is a cross section by Frost (1980) of the Whipple detachment fault (WDF) as exposed in the deep canyon of Whipple Wash west of Lake Havasu. This cross section and those of Figures 10 and 11 are typical of cross sections drawn through this and other Cordilleran detachment complexes in that upper-plate rocks (here both Precambrian crystalline and Tertiary) cannot be restored to a balanced geometric configuration by simply reversing normal fault displacements along the upper-plate normal faults. Doing so results in unfilled spaces in the restored sections. These faults, as exposed in the canyon (Fig. 8b), do not flatten downward as they approach the Whipple detachment fault. Gross and Hillemeyer (1982) commented on the geometric difficulties of preparing balanced cross sections of extended upper-plate units and specifically discussed the geometry of faults in the Whipple Wash section (Fig. 13). Essentially they concluded that upper-plate extension is so complicated by successive generations of rotating planar faults, including microfaults, that balanced palinspastic reconstructions cannot be made. They stated, for example (1982, p. 263), that

> Normal faulting within the upper plate is much more extensive than has previously been recognized—to the extent that microfaults are penetratively developed. Such penetrative deformation appears to be essential in producing the rotation and extension in upper-plate rocks throughout the detachment terrane. Ironically, it seems that the abundance of faults is responsible for relatively few of them having been observed. The pres-

ence of penetrative faulting (and thus great numbers of faults) means that offsets along individual faults need be very small, on the order of a few centimeters or less. . . .Very large faults certainly do exist, but the fault blocks which they bound . . . are not rigid blocks, rather they must be penetratively deformed via microfaults.

We do not agree with this analysis, partly because we are skeptical that penetrative microfaulting exists to the extent described, and partly because we do not believe that penetrative faulting in fault-bounded panels can geometrically compensate for the abrupt termination at the Whipple fault of upper-plate faults with dip-slip displacements of tens to hundreds of meters (Fig. 13). As an alternative we propose (Fig. 14) that the surface mapped as the Whipple fault in Whipple Wash is a second- (third? fourth?) generation splay of an earlier Whipple detachment fault, and not the original detachment itself. Figure 14a illustrates a phase of detachment hypothesized earlier, where Tertiary strata lay nonconformably atop upper-plate Precambrian basement rocks. Rotation of strata only in the northeastern part of the upper plate was accomplished along an early listric fault that flattened northeastward into the Whipple fault (WDF 1). Figure 14b illustrates the subsequent formation of an inferred low-angle, upper-plate splay—a low- to moderate(?)-dipping fault that branches upward from WDF 1, climbs across the upper-plate crystalline basement, and into its shallow-dipping Tertiary cover. In this section the hanging wall splay has become the active "Whipple" detachment fault (WDF 2). Basement rocks that were upper plate to WDF 1 have been excised by the development of the higher splay and transferred to the lower plate of WDF 2. The listric fault of phase 1 extension is truncated at depth by WDF 2. It (listric fault 1) now appears to be a planar hanging wall fault with the attendant geometric problems that such faults present at their abrupt intersection with a presumably coeval underlying detachment fault (actually, a younger, truncating structure). Once the upper-plate normal fault is recognized as an older structure, now kinematically dead, the problem of attempting to balance upper-plate strata on either side of it vanishes.

With the transfer of plate interaction from WDF 1 to WDF 2, lower plate chloritic breccias (phase 1), are "switched" (using railroad parlance) to move up beneath the hanging wall of the now-active higher Whipple splay (WDF 2). If the Whipple fault is routinely mapped only as the fault separating mylonitic gneisses and chloritic breccias from "upper-plate" nonmylonitic crystalline or Tertiary rocks, then WDF 2 would not be recognized as a younger phase of the master Whipple detachment updip from where the chloritic breccias depart from it. WDF 1 would still be recognized as the Whipple fault, but WDF 2 (with Tertiary strata in its footwall) would only be considered an "upper-plate" detachment fault offsetting Tertiary strata and its crystalline basement. Without such recognition, the Whipple fault is mistakenly interpreted as a single-generation structure across which sections cannot be balanced and variable dips in upper-plate strata cannot be accounted for. Although microbreccias exist beneath WDF 2 in Whipple Wash, other younger and/or higher

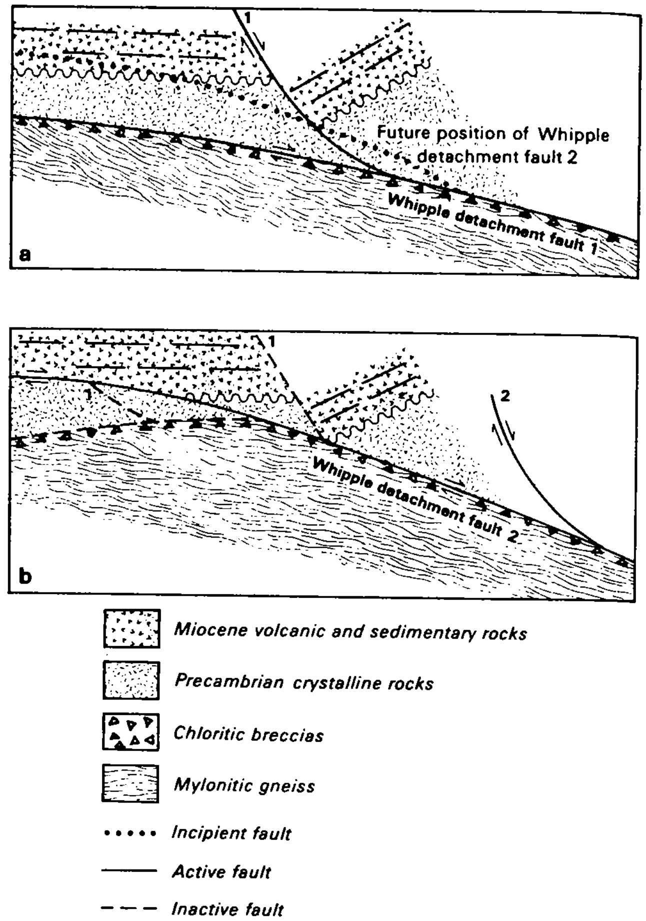

Figure 14. Diagrammatic representation of evolving geometry of Whipple detachment fault to explain geologic relations illustrated in Figure 11. a, Illustration of inferred geometry of Whipple fault during phase 1 of its development. WDF 1 lies at unknown depth beneath Tertiary-Precambrian crystalline nonconformity. Upper-plate listric fault 1 is responsible for rotating hanging wall strata; footwall strata are not rotated. The future trace of a higher detachment fault splay (WDF 2) that will cut upward across upper-plate rocks is shown by the heavy dotted line. b, Illustration of subsequent phase of extensional deformation. WDF 2 is now the active detachment fault. Crystalline rocks formerly in the upper plate of WDF 1 have been excised by the younger, higher WDF 2 splay and transferred to its lower plate. The upper-plate listric fault 1, active during phase 1, has been truncated and offset by WDF 2; it, like WDF 1, is now kinematically "dead." Lower-plate chloritic breccias (cb) are "switched" to follow up along and beneath WDF 2. A second phase upper-plate listric fault, 2, is shown on the right edge of section b. It, or other phase 2 faults, can theoretically rotate previously unrotated strata (e.g., left side of section) or re-rotate already tilted strata (e.g., right side of section).

splays elsewhere in the range may have formed at levels too shallow for microbreccias to develop along them, despite the fact that they, like their earlier phase parents, may be underlain by chloritic breccias and older mylonitic gneisses.

John (1987) saw similar evidence for the evolving geometry of detachment faults. She has suggested that early movement along the Chemehuevi fault (CDF 1 by analogy with Fig. 14) was followed by broad antiformal arching of the fault along a north-northwest trend. After arching, the Chemehuevi fault on the western flank of the arch is believed to have become inactive. However, on the eastern flank, John inferred continued or renewed movement along a postarching, upper-plate fault splay (which we would call CDF 2), which merges northeastward with the original detachment fault (CDF 1). Western exposures of the Chemehuevi fault are characterized by a wide zone of "breccia and cataclasite, lacking any throughgoing planar surface." In contrast, exposures of the presumably younger, shallower fault on the east flank of the arch display thinner breccias, gouge, and a sharp planar fault contact. The evolving geometry of faulting interpreted by John is closely similar to that hypothesized independently in Figure 14 for the Whipple fault.

Howard and others (1982b) and Howard and John (1987) have presented a different view of the Whipple detachment fault than that proposed by us above. They considered the Whipple fault (and the Chemehuevi fault that they correlate with it) to be the sole fault (in a geometric sense) of the detachment terrane at this latitude in the Colorado River extensional corridor. They hypothesized that those portions of the Whipple fault now exposed in easternmost parts of the Whipple Mountains originally lay at a paleodepth of 10 to 15 km, and that Tertiary strata that now dip into the fault originally lay many kilometers above it. In their view, progressive rotation (domino-style) of the upper crust (basement plus Tertiary cover) has occurred along multiple, closely spaced (1–2 km), once steep normal faults that transected the entire crust above the deep-seated Whipple detachment fault. Those upper-plate faults now dip gently, subparallel to the Whipple fault, and are said to bound upended (tilted) sections of the upper crust as thick as 12 or 13 km when measured perpendicular to the Tertiary-Precambrian nonconformity. We cannot agree with this geometric analysis as it applies to the Whipple fault and its upper-plate structures in the eastern Whipple Mountains for two reasons: (1) we consider the present Whipple fault in this area to be a very young detachment, perhaps one of the youngest such faults in the region, and not an early sole fault for the entire terrane above which all extension occurred; and (2) the present Whipple fault was a very shallow structure (<2 to 3 km) at the time it formed in the eastern Whipple Mountains.

Looking at western portions of Figure 13 or the cross sections of Figure 11, there is no basis for postulating that the present Whipple fault once lay 10 or more km below the subhorizontal Tertiary strata that now rest tectonically on it. How is the observed juxtaposition possible? Clearly, domino-block rotation of upper-plate normal faults cannot be called upon to accomplish the lowering of subhorizontal Tertiary strata by 10 or more km to a position of juxtaposition against a "deep-seated" Whipple fault. The Tertiary rocks have not been rotated significantly in some areas of the Whipple Mountains, and the thick section of crystalline rocks once beneath them and above an inferred 10-km-deep detachment fault is missing.

Again, these geometric relations are easily explained by the development of a shallow-dipping splay off an earlier generation Whipple fault. The splay cut upward at shallow angles (<20–25°) across the former upper plate, first through upper-plate basement and then into the thin (<2 to 3 km), faulted Tertiary section deposited atop it. We are convinced that some such splays reached the earth's surface during crustal extension, thus allowing lower plate rocks (e.g., chloritic breccias and mylonitized gneisses) to be eroded and to be preserved as detritus in nearby sedimentary basins. The presence of clasts of both Tertiary mylonitic gneisses, and, less commonly, chloritic breccias in Tertiary sedimentary rocks that are now allochthonous with respect to the present Whipple fault attest to multiple phases of detachment in the evolution of the Whipple core complex. John (1987) and S. Reynolds (personal communication, 1985) have also reported the occurrence of detachment fault-related chloritic cataclasites in allochthonous alluvial fan deposits.

A somewhat similar, but considerably more complicated scenario of detachment-related faulting must be postulated for the Sacramento Mountains, 60 km north-northwest of the Whipple Mountains (Fig. 2). Figure 15a is a diagrammatic southwest-northeast cross section through the southern part of the range, drawing from the study by McClelland (1982, 1984). Figure 15b is a true-scale cross section through McClelland's southern area to better illustrate how thin the detachment fault–bounded allochthons are in this range. As in the Whipple Mountains, mylonitic gneisses, a southwest-dipping mylonitic front, and chloritic breccias (retrograded and cataclasized mylonitic gneisses) underlie the major Sacramento detachment fault. Miocene strata and their nonmylonitized Precambrian basement overlie the fault, also as in the Whipple Mountains. However, significant differences exist between the upper-plate structures of the two ranges. As Figure 15 illustrates, three allochthons lie above the Sacramento Mountains detachment fault: (a) a lower allochthon consisting of Precambrian and Mesozoic(?) crystalline rocks; (2) a middle allochthon containing Precambrian basement and an unconformably overlying section of Miocene sedimentary and volcanic rocks; and (3) an upper allochthon consisting largely of Miocene fanglomerates. Field relations and geochronologic data indicate a consistent younger-over-older stacking of units within the three plates. The highest allochthon, somewhat surprisingly, contains southwest-tilted strata that dip considerably more steeply than do older strata in the underlying middle plate. Multiple generations of steep- to moderately-dipping normal faults cut and/or are truncated by detachment faults above the basal Sacremento detachment. As in the Whipple Mountains, the Sacramento fault appears to be the youngest major fault in the range (Fig. 15) because it cuts late, steep normal faults in its upper plate.

Spencer (1985) presented convincing data that in the north-

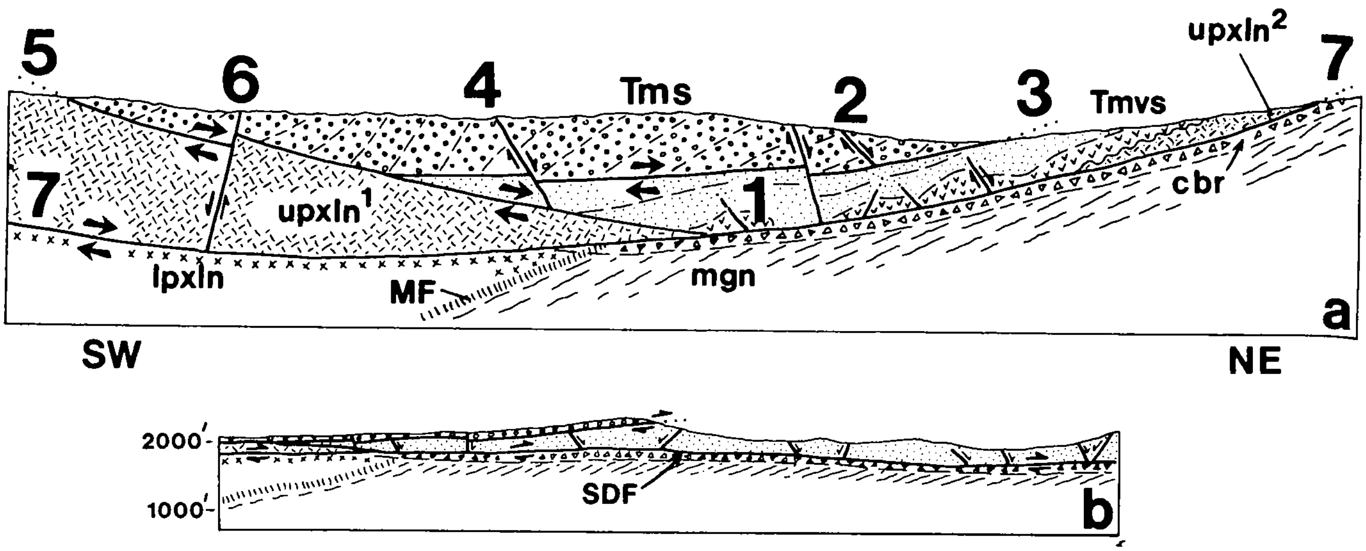

Figure 15. a, Diagrammatic southwest-northeast cross section through the southern Sacramento Mountains (redrawn from McClelland, 1984), illustrating geometric relations in this complexly deformed detachment terrane. Numbers designate sequence of formation (1 = oldest) of high- and low-angle (detachment) faults determined from field studies. Rock unit and structure abbreviations not used in previous figures include Tms = Tertiary Miocene sedimentary rocks (which include basalts in the northern Sacramento Mountains) and SDF = Sacramento detachment fault (section b). b, Southwest-northeast cross section through the southern Sacramento Mountains (McClelland, 1984). Figure illustrates how thin the detachment fault-bounded allochthons shown diagrammatically in section a are when drawn with horizontal: vertical scale = 1.

ern Sacramento Mountains (Fig. 2), as in the Whipple Mountains area (Davis, 1980), the basal detachment fault developed at a very shallow structural level:

> Flat-lying basalt at Flattop Mountains, yielding a K-Ar date of 14.6 ± 0.2 m.y. rests unconformably on tilted conglomerate . . . and is interpreted as post-dating detachment faulting and associated tilting. The contact at the base of the basalt of Flattop Mountain is about 250 to 300 meters above the basal detachment fault. Since the basalt immediately post-dates detachment faulting, the subhorizontal basal detachment fault appears to have been not much more than 300 meters below the ground surface at the time of latest fault movement.

From field relations determined by McClelland, we hypothesize a sequence of faulting events, both high and low angle, that generally explains geometric relations illustrated in Figure 15. Other scenarios can be proposed to explain McClelland's field relations, but we believe that all would require variations on our basic theme. That theme (Fig. 16) requires the formation during crustal extension of both excisement and incisement faults, the latter representing the development of detachment fault splays *below* preexisting active detachments. The consequence of incisement is that rocks formerly belonging to the lower plate of a controlling detachment fault (e.g., fault 5, Fig. 16c) are transferred to an upper-plate position with the development of a younger, structurally deeper splay (cf. Fig. 16c,d). The interpretation of evolving fault geometries in the Sacramento Mountains shown in Figure 16, rests on two assumptions: that moderately steep to steep normal faults of demonstrably different ages all flattened into active lower-angle detachment faults; and that, with the exception of faults developed during deposition of the basal stratigraphic sequence (phase 1 faulting, Fig. 16a), *all* faults in the Sacramento Mountains must have formed in the geologically brief period (<1 m.y., ca. 14 to 15 Ma) after deposition of now tilted basalt flows (14.6 Ma) in the uppermost fanglomerate sequence (Tms, Fig. 15) and before extrusion of the unconformably overlying Flattop basalts in the northern part of the range (14.6 Ma, Spencer, 1985).

Before concluding this discussion of the complexities of detachment faulting in the northern lower Colorado River region, two additional points are noteworthy. The first is that geologists may generally have underestimated both the magnitude and rate of displacement within detachment fault terranes. Reynolds and Spencer (1985) presented evidence that approximately 50 km of relative northeastward displacement of upper-plate rocks occurred on the Bullard (=Whipple?) detachment fault in west-central Arizona during middle to late Tertiary time. To the northwest in the Whipple Mountains, displacement of at least 40 km at a minimum rate of 0.8 cm/yr along the Whipple fault is strongly indicated by its post-20-Ma truncation of the Chambers

Well dike swarm; much faster rates for the period 18 to 20 Ma are indicated by geologic and geochronologic data reviewed in this paper. Furthermore, since Tertiary stratigraphic evidence (Otton, 1982; Davis, 1986) and an Oligo-Miocene age for mylonitization (26 ± 5 Ma, this paper) indicate that crustal extension in the NLCR region predates displacement of the Chambers Well swarm, total lithospheric extension in the region may considerably exceed the 40+ km that occurred after 20 Ma at the latitude of the Whipple Mountains.

Finally, the relationship between Cordilleran thrust faults and younger detachment faults deserves a brief comment. Some workers have suggested that the mechanical problems surrounding the formation of primary, shallow-dipping detachment faults (especially the large angle between steep or vertical σ_1, and shallow-dipping faults) might be resolvable if the development of such faults was controlled by preexisting, shallow-dipping structural anisotropies. Thrust faults of Mesozoic or earliest Cenozoic age are the controlling structures most frequently alluded to (e.g., Allmendinger and others, 1983, for the Sevier Desert detachment, Fig. 1). Although the reactivation of Mesozoic thrust faults as Cenozoic extension faults has been well documented in the eastern Basin and Range area (e.g., Royse and others, 1975; Smith and Bruhn, 1984) we are convinced that at least some low-angle detachment faults must be generated as primary structures, i.e., that they transect crustal rocks without the presence of preexisting shallow-dipping anisotropies. Wernicke and others (1985) have reached a similar conclusion from their studies in the Mormon Mountains of southern Nevada. The Whipple and Sacramento detachment faults described in this chapter both cut upward (southwestward) at shallow angles across basement crystalline rocks and into shallow-dipping, unconformably overlying Miocene strata that clearly never experienced prior thrust faulting.

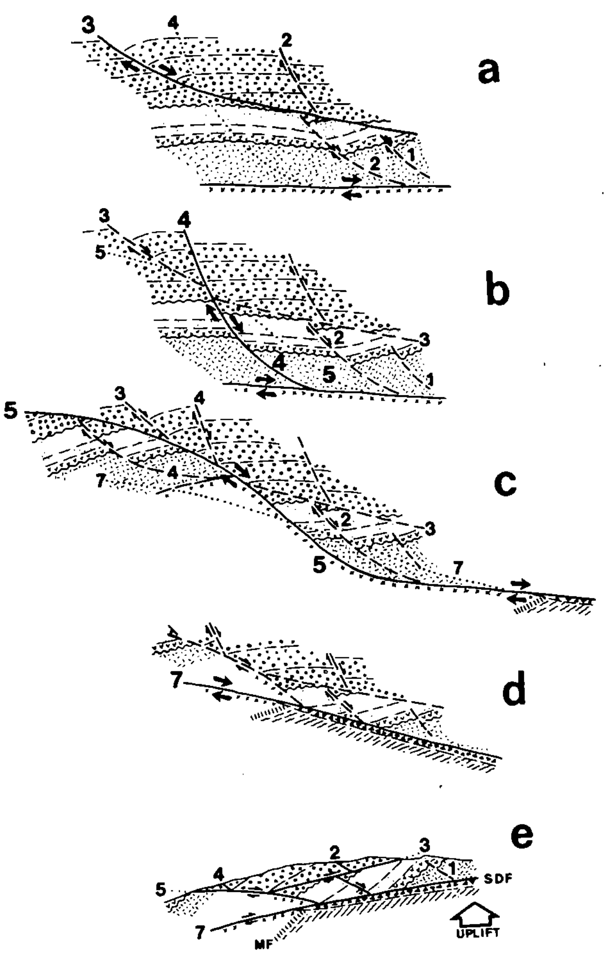

Figure 16. Possible scenario for development of multiple generations of high- and low-angle faults defined by McClelland (1984) and illustrated in Figure 15. Extension faults active in each section (a, b, etc.) are shown by solid lines. Faults previously active are shown by dashed lines. Faults that will become active in the underlying younger section are shown by dotted lines. All faults except generation 1 are interpreted as forming after deposition of uppermost unit (Tms, Fig. 15). All high-angle faults are interpreted as having once had listric geometries with respect to synchronously operative lower detachment faults. Fault 5 is an excisement splay, transferring former upper-plate rocks (b) to a lower-plate position (c). Fault 7, the present Sacramento Mountains detachment fault, has both excisement and incisement geometries (right- and left-hand sides of c, respectively). Compare (e) with present geologic relations of Figure 15.

CONCLUSIONS

In this chapter we have set forth what we believe are several new perspectives on the origin of Cordilleran detachment faults during lithospheric extension. We have drawn heavily upon observations, mapping, and the structural interpretations of numerous workers in the magnificently exposed detachment terrane of the lower Colorado River region of southern Nevada, southeastern California, and western Arizona. Some of our major conclusions are summarized below.

Cordilleran detachment faults are best explained as evolving low-angle shear zones that probably root into lower upper crustal or midcrustal structural levels during continental extension. These detachment zones propagate upward across the overlying crust and either reach the surface directly as low-angle faults or terminate at shallow depths in pull-apart complexes of closely spaced normal faults. Along these evolving shear zones, lower-plate rocks are drawn rapidly upward and out from beneath brittlely extending upper-plate rocks. Footwall rocks, which formed at great depth but now lie in fault contact below broadly synchronous Tertiary supracrustal units, composed the active fault wall. This conclusion is supported by petrologic and geochronologic evidence from the Whipple Mountains that Oligo-Miocene footwall mylonitic gneisses, formed at depths >12 km and at temperatures as high as 535°C, had cooled from temperatures above 450°C to below 150°C between 20 and 18 Ma. Such rapid cooling is best explained by rapid uplift of the mylonitic gneisses in the footwall of the evolving Whipple detachment system.

Although mylonitic gneisses in the footwalls of some Cordilleran detachment faults show kinematic coordination with those faults in having the same direction and sense of shear, they are in some cases (e.g., Whipple Mountains) apparently formed at depth, either along down-dip portions of older detachment zones (Fig. 3) or within intracrustal zones of laminar flow that have no obvious connection to structural levels now exposed. Such mylonitic gneisses are later crosscut and carried upward in the footwalls of the cross-cutting detachment faults, where they become sheared, retrograded, and then brecciated at progressively higher, cooler, and more brittle structural levels. Mylonitic fronts, the tops of intracrustal shear zones characterized by predominantly crystal–plastic flow mechanisms, can also be transported upward in the active footwalls. As such they dip away from the capturing fault and return to considerable crustal depths away from the detachment fault (Fig. 3). This is a geometric necessity since only directly beneath and near the younger, capturing detachment are footwall rocks elevated to surface or near-surface levels. At some distance from the detachment fault the older, lower-plate mylonitic gneisses maintain their lower upper crustal or mid-crustal depths and are only transported laterally during crustal extension. Mylonitic detachment complexes form only when relative displacements along the evolving crustal shear zones are large enough to carry deep-seated mylonitic assemblages to shallow crustal levels. Here they become exposed through combinations of crustal warping, tectonic denudation, and erosion.

The angle now observed in outcrop between the detachment fault and the mylonitic front in its footwall (cf. Fig. 3, Fig. 17 of Davis and others, 1980) should record approximately the angle of initial discordance between deep, shallow-dipping mylonitic gneisses and the somewhat steeper, younger detachment fault zone that transected them and captured them in its rising footwall. For the Whipple Mountains, this angular discordance is approximately 10° to 25°, essentially the same range of angles at which higher levels of the Whipple detachment fault cut through upper-plate crystalline basement and into subhorizontal strata. These angular relationships at upper and lower ends of the evolving Whipple detachment fault system argue strongly for a low angle of dip during the time of its development. Detachment faults thus appear to be capable of transecting upper crustal rocks as primary, low-dipping shear zones uncontrolled by preexisting, shallow-dipping structural anisotropies, although the possible reactivation of older thrust faults elsewhere as detachment faults during crustal extension is not denied by this statement.

Contrary to popular belief, some of the major detachment faults exposed today in the Colorado River extensional corridor are not faults that formed at the start of extensional tectonics, but are rather only the youngest in a succession of major detachment structures. Detachment faults undergo various warpings at high angles to the direction of crustal extension during their development, probably due in large part to isostatically induced distortions of originally more planar faults (and possibly due to reverse-drag flexing above lower faults; Wernicke and others, 1985). Such warpings lead to the development of younger, more planar splays that either cut upward into former upper-plate rocks (excisement) or downward into former lower-plate rocks (incisement). Recognition of such geometric complexities offers fresh insights into deciphering the evolving strain patterns within major detachment zones. Studies in the Whipple Mountain region indicate that most upper-plate normal faults originally had listric geometries before losing their shallow-dipping lower segments as the consequence of excisement tectonics. The northeast-southwest–trending curviplanar geometry of major detachment faults in the Colorado River extensional corridor is probably an expression of primary corrugations (fluting) in the fault surfaces formed parallel to the direction of fault displacement. Some of these corrugations or flutes may have been influenced by preexisting structures now seen in footwall rocks.

Finally, the rate of displacement along some evolving detachment systems may be very rapid, perhaps greater than 1 cm/yr. Tertiary mylonitic gneisses formed within deep-seated intracrustal shear zones beneath the lower Colorado River region were carried upward very rapidly (<2 m.y.) to surface or near-surface levels in the footwalls of somewhat younger detachment faults. Geologists may have underestimated not only how rapidly intraplate extension along evolving detachment fault systems can occur, but the magnitude of extension produced by such systems.

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Processes in Continental Lithospheric Deformation

Edited by

Sydney P. Clark, Jr.
Department of Geology and Geophysics
Yale University
P.O. Box 6666
New Haven, Connecticut 06511

B. Clark Burchfiel
Department of Earth, Atmospheric, and Planetary Sciences
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

and

John Suppe
Department of Geological and Geophysical Sciences
Princeton University
Princeton, New Jersey 08544

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