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Strike-slip faults

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ABSTRACT

The importance of strike-slip faulting was recognized near the turn of the century, chiefly from investigations of surficial offsets associated with major earthquakes in New Zealand, Japan, and California. Extrapolation from observed horizontal displacements during single earthquakes to more abstract concepts of long-term, slow accumulation of hundreds of kilometers of horizontal translation over geologic time, however, came almost simultaneously from several parts of the world, but only after much regional geologic mapping and synthesis.

Strike-slip faults are classified either as transform faults which cut the lithosphere as plate boundaries, or as transcurrent faults which are confined to the crust. Each class of faults may be subdivided further according to their plate or intraplate tectonic function. A mechanical understanding of strike-slip faults has grown out of laboratory model studies which give a theoretical basis to relate faulting to concepts of pure shear or simple shear. Conjugate sets of strike-slip faults form in pure shear, typically across the strike of a convergent orogenic belt. Fault lengths are generally less than 100 km, and displacements along them are measurable in a few to tens of kilometers. Major strike-slip faults form in regional belts of simple shear, typically parallel to orogenic belts; indeed, recognition of the role strike-slip faults play in ancient orogenic belts is becoming increasingly commonplace as regional mapping becomes more detailed and complete. The lengths and displacements of the great strike-slip faults range in the hundreds of kilometers.

The position and orientation of associated folds, local domains of extension and shortening, and related fractures and faults depend on the bending or stepping geometry of the strike-slip fault or fault zone, and thus the degree of convergent or divergent strike slip. Elongate basins, ranging from sag ponds to rhombochasms, form as result of extension in domains of divergent strike slip such as releasing bends; pull-apart basins evolve between overstepping strike-slip faults. The arrangement of strike-slip faults which bound basins is tulip-shaped in profiles normal to strike. Elongate uplifts, ranging from pressure ridges to long, low hills or small mountain ranges, form as a result of crustal shortening in zones of convergent strike slip; they are bounded by an arrangement of strike-slip faults having the profile of a palm tree.

Paleoseismic investigations imply that earthquakes occur more frequently on strike-slip faults than on intraplate normal and reverse faults. Active strike-slip faults also differ from other types of faults in that they evince fault creep, which is largely a surficial phenomenon driven by elastic loading of the crust at seismogenic depths. Creep may be steady state or episodic, pre-seismic, co-seismic, or post-seismic, depending on the constitutive properties of the fault zone and the nature of the static strain field, among a number of other factors which are incompletely understood. Recent studies have identified relations between strike-slip faults and crustal delamination at or near the seismogenic zone, giving a mechanism for regional rotation and translation of crustal slabs and flakes, but how general and widespread are these phenomena, and how the mechanisms operate that drive these detachment tectonics are questions that require additional observations, data, and modeling.

Several fundamental problems remain poorly understood, including the nature of formation of en echelon folds and their relation to strike-slip faulting; the effect of mechanical stratigraphy on strike-slip–fault structural styles; the thermal and stress states along transform plate boundaries; and the discrepancy between recent geological and historical fault-slip rates relative to more rapid rates of slip determined from analyses of sea-floor magnetic anomalies. Many of the concepts and problems concerning strike-slip faults are derived from nearly a century of study of the San Andreas fault and have added much information, but solutions to several remaining and new fundamental problems will come when more attention is focused on other, less well studied strike-slip faults.

INTRODUCTION

The earliest scientific record of strike slip was of a New Zealand earthquake in 1888, the same year the Geological Society of America was founded. Many other strike-slip faults have been discovered since that time, also as a result of observed surface ruptures. The great San Francisco earthquake of 1906 on the San Andreas fault is particularly noteworthy in this regard. Strike-slip faults, however, are regional structures and require regional studies to document their existence and history. Several decades elapsed after that New Zealand earthquake before the geological community had sufficient data to extrapolate from empirical, instantaneous-strike-slip in earthquakes to more abstract interpretations that permitted tens or even thousands of kilometers of crustal translation by strike slip over geologic time. Then plate tectonics, with its startling concepts of crustal mobility, allowed geologists to overcome the limitations of fixist tectonics which prevailed before the 1960s and to understand the mechanical complexities and tectonics of strike-slip.

Extensive field investigations, innovative experimental studies, images of the third dimension by seismic reflection and drilling, refined dating techniques, painstaking interpretations of paleoseismicity, and analyses of modern earthquakes show that crustal slabs and plates have indeed slipped horizontally great distances during geologic time, as Wegener's hypotheses presupposed. Now we have new ideas of how elongate uplifts and basins rise and fall along active strike-slip faults; we see increasingly com-

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TABLE 1. CLASSIFICATION OF STRIKE-SLIP FAULTS

INTERPLATE (deep-seated)	INTRAPLATE (thin-skinned)
TRANSFORM faults (delimit plates, cut lithosphere, fully accommodate motion between plates)	TRANSCURRENT faults (confined to the crust)
Ridge transform faults* • Displace segments of oceanic crust having similar spreading vectors • Present examples: Owen, Romanche, and Charlie Gibbs fracture zones	Indent-linked strike-slip faults* • Separate continent-continent blocks which move with respect to one another because of plate convergence • Present examples: North Anatolian fault (Turkey); Karakorum, Altyn Tagh, and Kunlun fault (Tibet)
Boundary transform faults* • Join unlike plates which move parallel to the boundary between the plates • Present examples: San Andreas fault (California), Chaman fault (Pakistan), Alpine fault (New Zealand)	Tear faults • Accommodate differential displacement within a given allochthon, or between the allochthon and adjacent structural units (Biddle and Christie-Blick, 1985) • Present examples: northwest- and northeast-striking faults in Asiak fold-thrust belt (Canada)
Trench-linked strike-slip faults* • Accommodate horizontal component of oblique subduction; cut and may localize arc intrusions and volcanic rocks; located about 100 km inboard of trench • Present examples: Semanko fault (Burma), Atacama fault (Chile), Median Tectonic Line (Japan)	Transfer faults • Transfer horizontal slip from one segment of a major strike-slip fault to its overstepping or en echelon neighbor • Present examples: Lower Hope Valley and Upper Hurunui Valley faults between the Hope and Kakapo faults (New Zealand), Southern and Northern Diagonal faults (eastern Sinai)
	Intracontinental transform faults • Separate allochthons of different tectonic styles • Present example: Garlock fault (California)

*See Woodcock (1986, p. 20) for additional examples, both ancient and modern, and for their geometric and kinematic characteristics.

pelling evidence that crustal "miniplates" have rotated about vertical axes in broad zones of simple shear; we have good theoretical and experimental models supported with well-documented geological examples to explain the once thorny question of how major strike-slip faults terminate. Our understanding of how they behave mechanically has improved, but we still know little about these faults in the third dimension or about their thermo-mechanical behavior. Seismology and paleoseismology have been especially important in providing mechanical understanding, because in continental areas such as southern California, New Zealand, and Asia, active strike-slip faults can be studied to give ideas about how they have behaved during geologic time.

Perhaps because it is so well exposed along much of its length, because it is an active fault so close to a highly populated part of the United States having several major academic and governmental scientific groups living almost directly upon it, and because of the presence of petroleum in structural traps along it, the San Andreas fault is the most extensively studied strike-slip fault in the world (see summary papers by Crowell, 1979; Hill, 1981; Allen, 1981). For that reason, the San Andreas fault has been the source of many ideas pertaining to strike-slip tectonics (Hill, 1981); I therefore give the San Andreas fault what may seem to some readers an undue provincial emphasis in this review.

The purposes of this paper are to collect and review a number of principles, concepts, and notions about strike-slip faults; to summarize what is known about their geometric, kinematic, and tectonic implications; to discuss some of the outstanding problems; and to call attention to some challenging problems that must receive attention over the next decade or so to gain better understanding of these structures. A plethora of literature citations is given to show the sources of those principles, concepts, and notions, and to guide the reader toward more specific information. Many references to the older literature demonstrate that many of our beliefs about strike-slip faults were perceived, even if only dimly, by remarkably insightful geologists well before the plate-tectonic revolution.

TERMINOLOGY AND CLASSIFICATION

Strike-slip faults and dip-slip faults are the end members of the spectrum in a kinematic classification of faults (Reid and others, 1913; Perry, 1935). A strike-slip fault is "a fault on which most of the movement is parallel to the fault's strike" (Bates and Jackson, 1987).

The term "wrench fault" was popularized by Moody and Hill (1956), who borrowed the term from Kennedy (1946). Kennedy, in turn, was influenced by E. M. Anderson, who used the term in 1905, because "wrench planes" was used for a long time by the Scottish Geological Survey in the Highlands (Anderson, written discussion in Kennedy, 1946). All of these writers used the term for a deep-seated, regional, nearly vertical strike-slip fault which involves igneous and metamorphic basement rocks as well as supracrustal sedimentary rocks (Moody and Hill, 1956; Wilcox and others, 1973; Biddle and Christie-Blick, 1985). Other writers referred to such faults as "transcurrent faults" (Geikie, 1905); many still do, and it is a good term for any major strike-slip fault whose relation to a genetic classification (for example, Woodcock, 1986) is not clear.

Many writers have used "wrench fault" with increasing frequency for any and all strike-slip faults, whether or not they are regional, vertical, or involve crystalline basement. Many major strike-slip faults are not vertical, however, nor can they be shown to involve basement. Moreover, "wrench fault" carries the kinematic implication of torsion, which adds to the confusion. For these reasons, I recant my previous infatuation with "wrench faults" (Sylvester, 1984) and recommend that *strike-slip fault* be used for a fault of any scale along which movement is parallel to the strike of the fault, that *transform fault* be retained for a plate-bounding

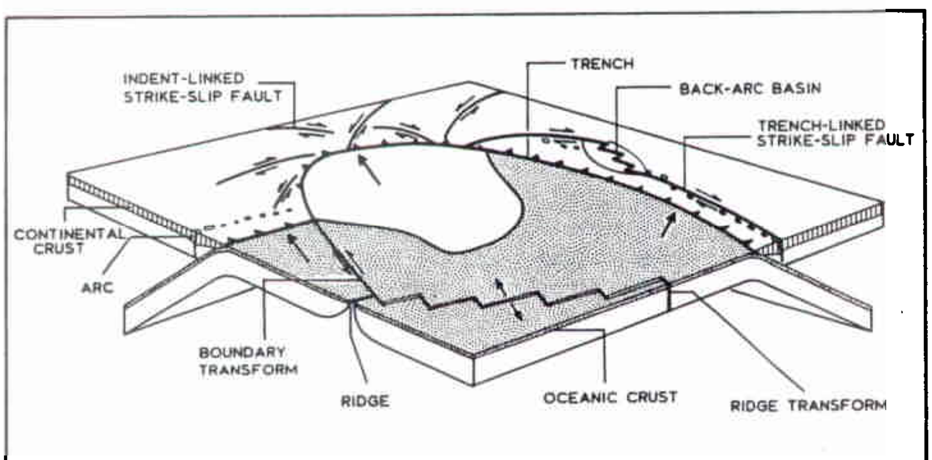


Figure 1. Plate-tectonic settings of major classes of transform faults (redrawn from Woodcock, 1986). Reproduced with permission of The Royal Society (London).

strike-slip fault, and that *transcurrent fault* be revived as the general term for the variety of ike-slip faults which do not cut the lithosphere (Table 1).

A transform fault is a kind of plate-bounding strike-slip fault, regional in scale, that cuts through the lithosphere and fully accommodates the motion between plates. Woodcock (1986) has presented a useful, although genetic, classification of transform faults according to their plate-tectonic setting (Fig. 1). The primary attributes of ridge transforms, the first of Woodcock's four types, are that they link spreading oceanic ridges and cut oceanic lithosphere (DeLong and others, 1979; Woodcock, 1986). They have a short active life as a strike-slip strand, but a long later history as an oceanic fracture zone with dip-slip displacements, and a sense of strike slip that is opposite to the sense of ridge offset (Wilson, 1965). Moreover, the apparent offset of ridge crests may be very much greater than the actual displacement along the ridge transform itself. Ancient ridge transforms are presently restricted to ophiolite complexes, posing difficulties for their recognition, but fairly compelling examples have been described in ophiolite complexes in California (Saleeby, 1977; Cannat, 1985), Washington (Miller, 1985), Newfoundland (Karson and Dewey, 1978), New Caledonia (Prinzhofer and Nicolas, 1980), the Apennines (Abbate and others, 1972), and Turkey (Gianelli and others, 1972). They are recognized by the presence of one or more of the following features: a zone of intensely deformed peridotite having a vertical foliation and horizontal lineation, syntectonic dikes parallel to the shear zone, hydrothermal alteration, and less than normal crustal thickness owing to reduced partial melting. In addition, the penetrative shear strain decreases away from the ancient transform zone, and the sense of shear through the zone may be determined by the bending of the foliation or from the mineral fabrics.

Trench-linked strike-slip faults (Fig. 1) are parallel to the trench and located within, or immediately bordering, the magmatic arc. Woodcock (1986) does not regard them as true transforms, but they are, because they cut through the lithosphere and delimit a fore-arc plate against a hinterland plate. Although displacements measurable in hundreds of kilometers are typical of trench-linked strike-slip faults, these faults accommodate only part of the total displacement behind the trench at a convergent plate boundary (Woodcock, 1986, p. 22). The Semangko fault zone (Fitch, 1972; Page and others, 1979; Karig, 1980; Hla Muang, 1987) may be regarded as the prototype of this kind of fault. The Atacama fault inboard of the Chile trench in South America (Allen, 1965) may be another example. Ancient trench-linked strike-slip faults may be partially or completely obliterated or buried by the very plutons and volcanic ejecta that they channel to the surface along the volcanic arc. Thus, major structural and lithologic discontinuities in the country rocks between batholiths in the Peruvian Andes may be vestiges of a trench-linked strike-slip fault which localized a long narrow line of subarc intrusions in an "Andinotype" orogenic environment (Pitcher and Bussell, 1977; Pitcher, 1979); similarly, a Mesozoic trench-linked strike-slip fault may have been responsible for localization of plutons in the Sierra Nevada batholith of California to the extent that they eventually obliterated the fault. Tobisch and others (1986) give evidence that the batholith formed in a long-lived extensile tectonic regime related to tumescence, arising from subduction-zone heating and magmatism, and they implied that the extension may have been related to orogen-parallel strike slip. The Foothills fault system, active in Late Jurassic time in the western Sierra Nevada, may have been part of such a system of faults (Clark, 1960; Cebull, 1972); so also may have been the proto–Kern Canyon fault which was active in the southern Sierra Nevada when the batholith was emplaced in Mesozoic time (Saleeby and Busby-Spera, 1986). In addition, an increasing body of field evidence shows that intrusions of granitic plutons are related spatially, temporally, and genetically to strike-slip faults. Shear-heating at depth along strike-slip faults is an added factor in enhancing crustal anatexis (Michard-Vitrac and others, 1980), and several workers have demonstrated the synkinematic intrusion of granitic plutons into pull-apart structures along major transform fault zones which, of course, provide a domain of extension for the intrusions (Davies, 1982; Hutton, 1982; Castro, 1985; Gapais and Barbarin, 1986; Guineberteau and others, 1987).

The San Andreas (California), El Pilar (Venezuela), Chaman (Pakistan), and Chugach-Fairweather–Queen Charlotte (Alaska and Canada) faults are examples of active and dormant boundary transform faults in Woodcock's classification (Fig. 1). They accommodate the horizontal displacement between continental or, rarely, oceanic plates which move horizontally with respect to each other. Boundary transforms have long lives and large displacements, comparable to those of trench-linked strike-slip faults (Woodcock, 1986, p. 23). They may evolve naturally from indent-linked strike-slip faults, or they may reactivate old, steep faults having a variety of orientations and mechanisms, including ancient subduction zones, as Hill (1971) proposed for the Newport-Inglewood zone in southern California, and as Freeland and Dietz (1972) postulated for the El Pilar fault in northern Venezuela. Conversely, arc and trench structures may owe their position to the former existence of regional strike-slip faults (Sarewitz and Karig, 1986).

Indent-linked strike-slip faults (Fig. 1) are not true transform faults, because they do not cut the lithosphere. They juxtapose pieces of continental lithosphere, especially in zones of plate convergence and tectonic escape. Displacements on single faults may range from tens to hundreds of kilometers, and they may reactivate any kind of available pre-existing, steep type of fault. Present examples include several strike-slip faults in Tibet and southern China that formed in response to collision with India, in the central part of southern Japan where the Pacific plate converges with the Asian plate (Sugimura and Matsuda, 1965), and in Iran where the northeast edge of the Arabian plate converges with Eurasia (Berberian, 1981; Tirrul and others, 1983). Indent-linked faults may be the main cause of the pervasive lineament networks in Precambrian continental crust (Watterson, 1978; Burtman, 1980; Woodcock, 1986, p. 23).

Intraplate, or intracontinental, transform faults are regional strike-slip faults which are similar to indent-linked strike-slip faults in that they are restricted to the crust (Lemiszki and Brown, 1988), but they need not be genetically related to "indentor tectonics," although they typically separate regional domains of extension, shortening, or shear. The Garlock fault, southern California, separates the southern end of the extended Basin and Range province from the Mojave Desert characterized by dextral shear and regional rotation about a vertical axis (Davis and Burchfiel, 1973). A series of strike-slip faults similarly terminates the northern end of the Basin and Range province from a domain of plate convergence and associated arc tectonics in Oregon (Lawrence, 1976).

A *tear fault* accommodates the differential displacement within a given allochthon, or between the allochthon and adjacent structural units (Biddle and Christie-Blick, 1985). Tear faults generally strike transverse to the strike of the deformed rocks and are sometimes called transverse faults or even transcurrent faults for that reason. Tear faults have been known for a long time in many fold-thrust belts (for example, in the Jura Mountains, Switzerland; Heim, 1919, p. 613–623; Lloyd, 1964), and early model studies reproduced them experimentally (Cloos, 1933; Lee, 1929).

Transfer fault has been used informally but increasingly for strike-slip faults that connect overstepping segments of parallel or en echelon strike-slip faults. Commonly located at the ends of pull-aparts, they "transfer" the displacement across the stepover from one parallel fault segment to the other. They have also been called "oblique faults" (Mann and others, 1983).

DEVELOPMENT OF THE CONCEPT OF STRIKE-SLIP FAULTING

The Book of Zechariah, written in 347 B.C., contains what may be the first reference to strike-slip faulting (Freund, 1971); in fact, one may regard it as a prediction, although the sense of slip is not specified: "and the Mount of Olives shall cleave in the midst thereof toward the east and toward the west, and there shall be a very great valley; and half the mountain shall remove to the north, and half of it to the south" (Zechariah 14:4).

The Swiss geologist, Arnold Escher von der Linth, may have been the first geologist to discover and correctly interpret the geology of a strike-slip fault (Şengör, 1987, written commun.; Şengör and others, 1985). Escher noted horizontal slickensides and the surprising linearity of the 8-km-long trace of the what is now called the "Sax Schwendi fault" which cuts the Säntis folds south of Wildkirchli in the canton of Appenzell, and he showed them to Suess (1885, p. 153, 154) in the 1850s. Escher's mapping clearly showed that the displacement is sinistral and ranges from 500 to 800 m.

The earliest report of strike slip during an earthquake may be the anecdote that a sheep corral was transected by the surface rupture of the great 1857 earthquake on California's San Andreas fault and was thus deformed into a structure shaped like the letter "S" (Wood, 1955). Freund (1971), however, gave credit for the first published record of strike slip to McKay (1890, 1892), one of New Zealand's most distinguished field workers, who documented surface strike slip associated with the earthquake of September 1, 1888, on what is now called the "Hope fault" on the South Island of New Zealand. Kotó (1893) described left-lateral displacement of up to 2 m and vertical displacement up to 6 m at Midori, Japan, associated with the Mino-Owari earthquake of 1891. The Chaman fault in Pakistan was discovered in 1892 when the Quetta-Chaman railroad was sinistrally offset 75 cm during an earthquake on that fault (Griesbach, 1893). Those earthquakes were in fairly isolated places, however, and their descriptions were published in journals with limited circulation.

The phenomenon of strike-slip faulting was roundly evinced to the scientific world when a maximum of 4.7 m of right-lateral slip abruptly occurred on the San Andreas fault in the great San Francisco earthquake of 1906. "Had the San Andreas been classified before 1906, it would probably have been described as a 'normal' fault" (Willis, 1938a, p. 799), because no clear indication of strike-slip faulting was recognized on any of its segments that had been mapped prior to 1906 (Hill, 1981). Many geomorphic features considered today as being characteristic of strike-slip faults, however, were recognized before the earthquake along a linear zone several hundred kilometers long (Fairbanks, 1907, p. 324; Gilbert, 1907, p. 228). It was the nearly 300-km-long surface rupture itself, in fact, which clearly linked the sundry segments together and demonstrated the length and mechanism of the fault. The San Francisco earthquake incontrovertibly taught contemporary geologists that substantial horizontal movements occur and recur along faults, especially in California (Lawson and others, 1908; Wood, 1916; Lawson, 1921). As late as 1950, however, some geologists were not so certain: "to judge from meager reports available, primary displacement of the Californian wholly transcurrent kind is uncommon in other regions on a major scale. Some of the San Andreas fractures seem to be almost unique" (Cotton, 1950, p. 750).

Geologists were very slow to extrapolate the observed movements during earthquakes to repetition of those movements over geologic time, producing displacements of hundreds of kilometers. The early English and Scottish geological literature contains many descriptions of strike-slip faults having from a few tens of meters to a few thousands of meters of horizontal displacement (Cunningham Craig in Horne and Hinxman, 1914, p. 70; Read, 1923), and Argand and his students certainly had a mobilist view of crustal deformation that included room for much horizontal displacement through geologic time (Şengör, in press), but many years of patient geologic mapping were needed in every part of the world to depict geometries and cumulative displacements among correlative rock units over great spans of time and space before acceptable interpretations of many kilometers of strike slip were eventually made.

Noble (1926, 1927) was one of the first investigators to propose tens of kilometers of strike slip on a major strike-slip fault, although Vickery (1925) gave little-noticed geologic documentation for 20 km of right slip on the Sunol-Calaveras fault, a major branch of the San Andreas fault system in central California. Noble correlated some distinctive nonmarine sandstone units across the San Andreas fault in the Mojave Desert in southern California. That correlation required 40 km of strike slip since late Miocene or early Pliocene time, a displacement which now we realize is too small, based on a partially incorrect correlation of a very similar lithofacies (Woodburne, 1975). Some contemporary geologists were receptive, but cautious: "The total horizontal displacement of the southwest mass northwestward past the northeast body [of the San Andreas fault] appears to be very notable, though not definitely determinable" (Willis, 1938a, p. 798), whereas others realized concurrently that much of coastal California is a long, narrow zone of horizontal shear strain with right slip on several faults (Lawson, 1921, p. 580; Buwalda, 1937b), perhaps much of California itself (Locke and others, 1940). Wallace (1949) mapped a 30-km-long segment of the San Andreas and concluded that strike slip on the San Andreas fault must exceed 120 km since mid-Tertiary time, based on his provisional slip rate determined from stream offsets; however, prior to 1950, the opinion of many California geologists coincided with that of the influential N. L. Taliaferro (1941, p. 161), who maintained that the horizontal displacement "has not been greater than one mile and probably less." He had mapped 70 km of the fault in central California, but that was not enough to reveal the magnitude of displacements recognized today. Taliaferro believed that the San Andreas fault was a late Pleistocene structure which coincided with a profound dip-slip fault of Eocene age.

Similar histories of recognition and rejection of great strike slip on faults in other parts of the world could be cited, but one that stands out is Dubertret's (1932) hypothesis for 160 km of left slip in the Dead Sea Rift. It was a remarkable hypothesis for the fact that the 160 km were required by the consequences of *continental drift and rotation* that Dubertret proposed to account for the arrangement of the Sinai and Arabian peninsulas relative to Africa, although the total slip along the Dead Sea transform is presently regarded as 105 km since Miocene time (Bartov and others, 1980). Bailey Willis (1938b), who translated relevant parts of Dubertret's article for the *Bulletin*, used contemporary dogma to reject Dubertret's hypothesis by saying, just as others said to Wegener, that it was an entertaining idea but one which lacked substance because no known force causes continents to drift.

Trümpy (1977, p. 1) related a delightful anecdote regarding the regionality of strike-slip faults and the attendant difficulties in recognizing them:

> In the autumn of 1957, I had the pleasure of guiding Professor Biq Chingchang through part of the Swiss Alps. When we arrived in the Engadine, I told him about our difficulties in correlating the geological structures across the valley. In his very modest and cautious way, he suggested that these discrepancies might be due to a sinistral wrench fault. I thought about it for a moment, and then I slapped my forehead: Professor Biq's explanation was the only possible one, and how could we have been so foolish not to have seen it before?
>
> How indeed the Swiss geologists failed to recognize such a conspicuous feature is a curious and in some ways an instructive story. Much of it is due to parochial shortcomings, because the Engadine line cuts across national boundaries. On a wider scale, Alpine geologists, convinced that they were working in the navel of the geological world, took little account of the discovery of large wrench faults in the Circum-Pacific orogens. . . .

Now the existence of strike-slip faults in the Alps is an acceptable hypothesis (Laubscher, 1971), and recent microstructural studies demonstrate an episode of right slip on the Insubric Line in Neogene time (Ratschbacher, 1986; Schmid and others, 1987).

Outside of California, several workers proposed major strike slip on several faults; these included Hess (1938), who postulated that the left separation of Haiti and Cuba could be attributed to post-Miocene lateral movement on a fault in the Bartlett Trough. I believe, however, that the first work to give a strong geologic basis to significant horizontal displacement on a fault, at least in the western world, was Kennedy's (1946) paper on the Great Glen fault in Scotland, a paper which was read to the Royal Geological Society of London in 1939, but which was not published until seven years later. Kennedy correlated rocks and structures produced in several different geologic events, now sinistrally separated 100 km across the Great Glen fault since middle Carboniferous time. Kennedy was quick to acknowledge assistance in his thinking from E. M. Anderson, who had already published the basis of "Andersonian" fault dynamics in 1905.

I believe Kennedy's paper spawned several nearly simultaneous papers in the late 1940s and early 1950s that presented evidence for great displacements on diverse strike-slip faults the world over, including the Alpine fault (450 km, Wellman in Benson, 1952), the San Andreas fault (more than 560 km, Hill and Dibblee, 1953), and the Dead Sea rift (100 km, Quennell, 1958, 1959).

Following these seminal papers by Kennedy, Quennell, Wellman, and Hill and Dibblee, geologists began to recognize strike-slip faults in many places in space and time, largely on the basis of physiography, and they were increasingly bold about publishing their findings during the 1960s, especially in western North America where the evidence is sufficiently well exposed over adequately long distances. Important papers include those about the San Andreas fault system (Crowell, 1952, 1960, 1962; Allen, 1957; Allen and others, 1960); about strike-slip faults in Alaska (Tocher, 1960; St. Amand, 1957); in Canada (Wilson, 1962; Webb, 1969; Roddick, 1967); and in the Basin and Range province (Nielsen, 1965; Shawe, 1965; Albers, 1967; Hill and Troxel, 1966; Stewart, 1967). Recognition of the geologic activity of strike-slip faults followed also in South America (Wilson, 1940, updated in 1968; Bucher, 1952; Rod, 1956; Campbell, 1968; Feininger, 1970); in the circum-Pacific (Biq, 1959; Allen, 1962, 1965; Sugimura and Matsuda, 1965; Kanenko, 1966); in Europe, the Middle East, and Asia (Ketin, 1948; Ketin and Roesli, 1953; Wilson and Ingham, 1958; Pavoni, 1961a; Burtman and others, 1963; Bagnall, 1964; Wellman, 1965; Burton, 1965), and in Africa (Rod, 1962; de Swardt and others, 1965).

As bold as the writers were in the 1950s and early 1960s, it is almost amusing today to read their struggles to explain the mechanics of the great faults. Many geologists, with some notable exceptions (for example, Schofield, 1960), could not visualize, mechanically, how tens or hundreds of kilometers of strike slip could diminish to zero at the ends of the faults, nor could they distinguish between interpretations obtained from fault geometry (separation) or from fault movement (slip) (Hill, 1981). A coherent kinematic explanation was given by Wilson (1965) in his paper on transform faults. Wilson postulated how tens or hundreds of kilometers of horizontal movement on an oceanic transform fault may be transferred into divergent or convergent movements of the Earth's crust. This concept may have been the key revelation in the formulation of plate tectonics and the mechanics of strike-slip faults (Hill, 1981), because it transformed contemporary thinking from fixed continents dominated by vertical movements to crustal slabs and plates having great horizontal movements.

One of the most cited of all *Bulletin* papers in this regard is Atwater's 1970 paper about the implications of plate tectonics for the Cenozoic tectonic evolution of North America. She used Wilson's transform mechanism for ocean-floor tectonics to formulate a tectonic history among several plates in the eastern Pacific and to bring plate tectonics onto the land. Her hypothesis made use not only of the spatial and temporal implications from sea-floor magnetic lineations, but it also explained the origin and history of one of the Earth's major transforms, the San Andreas fault, in a way consistent with the accumulated tangle of geological data.

Now, within the plate-tectonics framework provided by seismotectonics, geology, seismic reflections, geometry, and deep drilling, strike-slip faults are not as mysterious as formerly, but they are no less complicated. In fact, these same investigative tools and new concepts continually reveal the compound mechanical and tectonic roles and complexities of strike-slip faults. With increasing frequency, investigators are appealing to strike-slip faults to translate, rotate, and juxtapose great slabs of the crust within and along orogenic belts over thousands of kilometers throughout geologic time and in ways perhaps even Wegener could not have imagined.

RECOGNITION OF STRIKE-SLIP FAULTS AND THEIR DISPLACEMENTS

Active strike-slip faults are recognized by coseismic surface displacements (Lawson and others, 1908; Ketin and Roesli, 1953; Ambraseys, 1963), by distinctive physiographic features as outlined below, and by earthquake focal mechanisms (Julian and others, 1982). The modern place and rate of strike slip on active faults are documented by geodetic studies (Reid, 1910; Thatcher, 1979, 1986; Crook and others, 1982; Prescott and Yu, 1986). Their paleoseismic behavior is studied by detailed microstratigraphic studies (Sieh, 1978, 1984).

Physiographic Features

The most distinctive characteristic of active or recently active strike-slip faults is their extreme structural and topographic linearity over very long distances together with an array of distinctive physiographic features which were succinctly described by Noble (1927, p. 37) in reference to the San Andreas fault: "It has a curiously direct course across mountains and plains with little regard for gross physiographic features, yet it influences profoundly the local topographic and geologic features within it." Several of the strike-slip faults of the heavily vegetated parts of Asia and the western Pacific were recognized simply by the great length and linearity of their "rift" topography (Willis, 1937; Biq, 1959; Allen, 1962, 1965). However, recognition of strike-slip faults by geomorphic structures is limited by the durability of small, easily eroded landforms, such as sag ponds and deflected streams whose preservation depends mainly on climate. Even clearly offset cultural features, such as roads, walls, fences, and winery facilities (Steinbrugge and others, 1960; Rogers and Nason, 1971; Zhang and others, 1986, 1987) have a finite life expectancy.

The "rift" or trough may be up to 10 km wide (Gilbert, 1907, p. 234; Lawson and others, 1908, p. 25–52; Noble, 1927, p. 26–27; Davis, 1927) with a variety of fault-formed structures (Fig. 2), including pressure ridges (Wallace, 1949, p. 793), closed depressions called "sag ponds" if presently or once filled with water (Ransome, 1906, p. 286; Cotton, 1950), shutter-ridges (Buwalda, 1937a), and systematically deflected streams (Russell, 1926; Rand, 1931; Cotton, 1952; Wallace, 1968, 1976; Kuchay and Trifonov, 1977; Burtman, 1980; Sieh and Jahns, 1984; Zhang and others, 1987). These characteristic landforms are clearly illustrated in neotectonic strip maps of the Bocono and La Victoria faults in Venezuela (Schubert, 1982a, 1986a), the Agua Blanca fault in Baja California (Allen and others, 1960), and the San Andreas fault (Vedder and Wallace, 1970; Schubert, 1982b; Clark, 1984; Davis and Duebendorfer, 1987). Neither side of the "rift" need be higher than the other. When the opposing blocks of the fault move in strike slip, elongate blocks and slivers subside between parallel or en echelon fault strands, or they warp, sag, or tilt to form

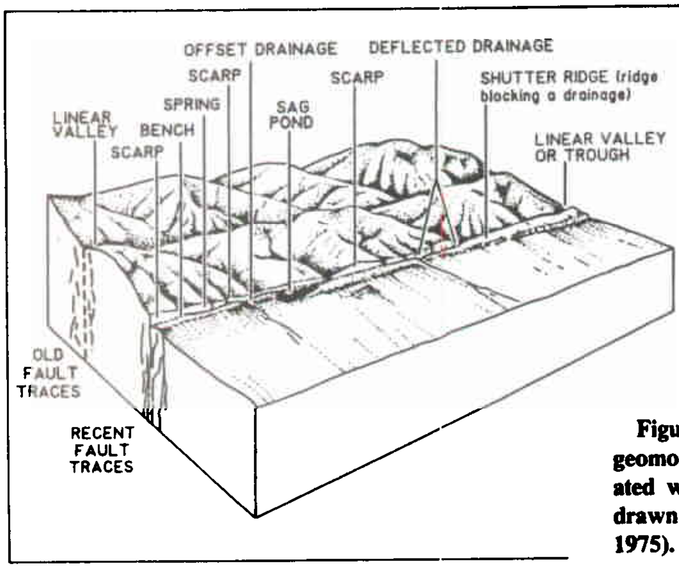


Figure 2. Block diagram of geomorphic structures associated with strike-slip faults (redrawn from Wesson and others, 1975).

closed depressions in the "rift" zone (Gilbert, 1907, p. 228; Fairbanks, 1907, p. 323), and they do so from the smallest to the largest scale. Other blocks may rise, tilt, or slide obliquely to produce pressure ridges. Notched ridges, fault-parallel trenches, or troughs along the fault may reflect increased erosion of the crushed and broken rocks in the fault zone (Fairbanks, 1907, p. 326; Allen and others, 1960; Vedder and Wallace, 1970; Wallace, 1976). Deflected streams must be used with caution to determine direction of lateral displacement. The offsets must be in an uphill direction, or else stream piracy or differential erosion may be invoked to explain them (Higgins, 1961; Allen, 1962, 1965; Wallace, 1976; Patterson, 1979).

The height of a fault scarp generally indicates its minimum vertical displacement, but this may not be true along a strike-slip fault which transects rugged topography, or where the vertical component of slip varies concurrently with the strike of the fault (Peltzer and others, 1988). On a small scale, shutterridges are formed by lateral or oblique displacement on faults transecting a ridge and canyon topography. A displaced part of a ridge shuts off a canyon, hence the term (Buwalda, 1937a; Sharp, 1954). On a larger scale, an entire mountain front may be exposed by lateral displacement of its toes or its other half, as Noble (1932) postulated for the south face of the San Bernardino Mountains in southern California.

Where several scarps are present within a strike-slip fault zone, it is typical that they will have opposing senses of vertical displacements or "scissors" geometry, due to topographic irregularities or because of the variable oblique displacements on the faults themselves. Indeed, a strike-slip fault may display reversals of the direction of throw along its trace, giving rise to descriptive phrases in the older literature as "scissoring" and "propeller faults" (Tomlinson, 1952).

Locally, the relatively uplifted block may slide by gravity or be thrust onto the adjacent, downdropped block. Where "scissoring" has apparently occurred along the strike of the strike-slip fault, a series of thrusts with alternat-

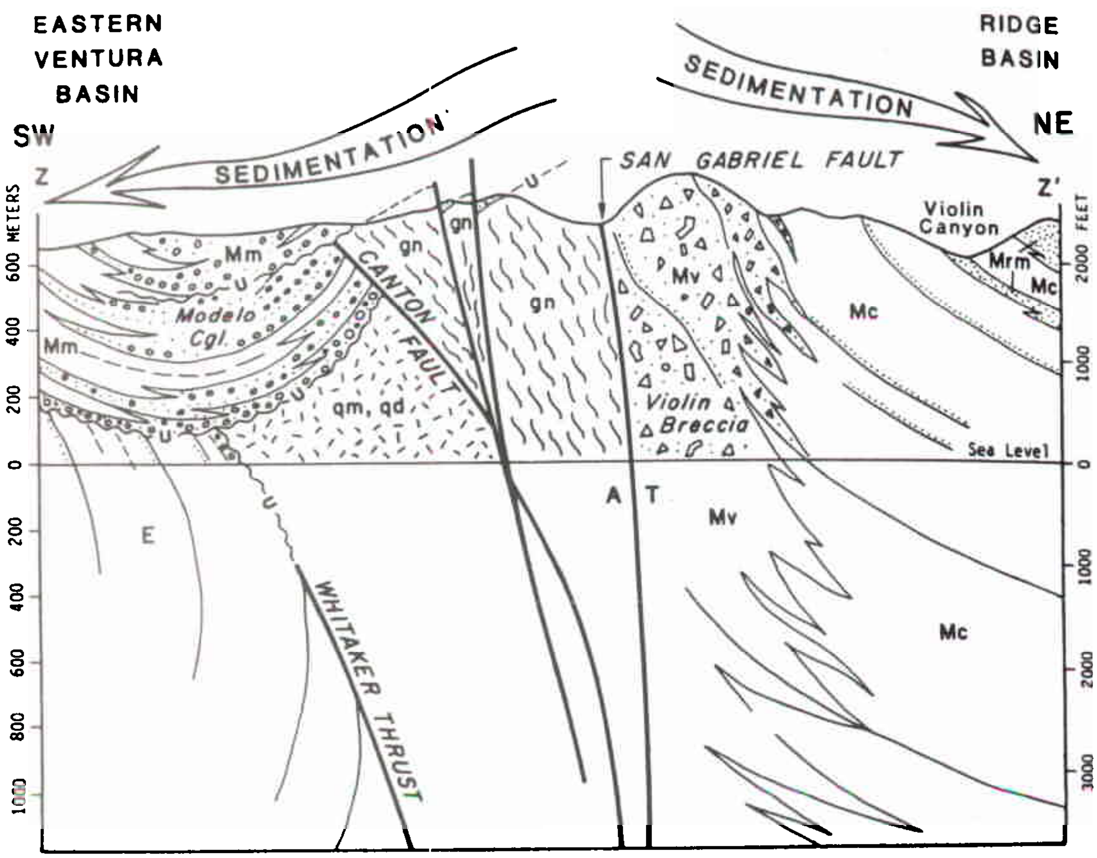


Figure 3. Diagrammatic cross section showing sedimentation mismatch across the San Gabriel fault, southern California (from Crowell, 1982a). Conglomerate of the Modelo Formation, derived from the northeast, now lies faulted against Violin Breccia which was derived from the southwest. The source areas for both stratigraphic units have been displaced by right slip on the fault zone. T = displacement toward observer; A = displacement away from observer. Reproduced by permission of the Pacific Section of Society of Economic Paleontologists and Mineralogists.

ing and opposing throw may be produced, perhaps effectively hiding the deeper presence of the responsible strike-slip fault. T[illegible] may be the explanation for the alternating and opposed verging folds and thrust directions in the Columbia Plateau of Washington (G. Davis, 1985, personal commun.; R. Bentley, 1987, personal commun.).

Geologic Features

The presence of a strike-slip fault is frequently indicated by en echelon arrays of fractures, faults, and folds in narrow elongate zones. In addition to truncating reference features such as stratification, foliation, folds, dikes, sills, and other faults, strike-slip faults juxtapose rocks of dissimilar lithology, facies, age, origin, and structure. Sedimentary facies may be telescoped or "stretched" across some faults. Locally, great mismatches in sedimentation history may be juxtaposed across a fault (Fig. 3). Slickensides or mullions, where present on fault surfaces, are mainly horizontal. The apparent simultaneous development of extended and shortened structures, together with the variable vertical separation along strike, are typical aspects of strike-slip faults. These features are caused by localized and alternating convergent and divergent components of lateral displacement along the fault in combination with alternating local uplift and subsidence of blocks and slices developed within the fault zone over time. Resultant local uplift and erosion of those blocks and slices yield unconformities of the same age as thick, but not laterally extensive, sedimentary sequences that were deposited very rapidly on adjacent subsided blocks. A great discordance between clast size or lithology of detritus in sedimentary units and possible sources across a fault typically reflects the horizontal translation of the source relative to the deposits.

Structural cross sections reveal variable senses of vertical separation on faults near the surface that coalesce downward into the main fault. Serial cross sections may bear little systematic relation to one another in terms of structure, style, throw, and juxtaposed lithology over distances of 1–2 km (Tomlinson, 1952) in contrast to those at the same scale, for example, across a major normal or thrust fault system.

The criteria for determination of the magnitude of horizontal displacement on a strike-slip fault are those that are least affected by depth of erosion (Gabrielse, 1985). They include offset of geologic lines having considerable horizontal extent that yield piercing points (Fig. 4; Campbell, 1948; Crowell, 1959, 1962), such as strand lines and shelf-to-basin transition zones (Addicott, 1968); sedimentary and metamorphic facies boundaries (Suggate and others, 1961; Roddick, 1967); formation pinch-outs; and isopachous lines (Stewart, 1983); channels and shoestring sands; constructional lines such as "fold axes" and intersections of surfaces (Crowell, 1962), of unique isotopic and geochemical trends (Silver and Mattinson, 1986; James, 1986), and of surfaces of considerable vertical extent such as dikes (Smith, 1962; Speight and Mitchell, 1979), plutons (Sharp, 1967), batholiths (Kennedy, 1946), and even suture zones (Şengör, 1979). Ideally the reference lines and surfaces should be at a high angle to the fault. In many other instances, a rare rock type or unique assemblage of rocks has been shed or erupted from one side of a strike-slip fault and deposited on the other side, so that a minimum horizontal displacement may be determined by the distance the deposit has been removed from its source (Crowell, 1952; Fletcher, 1967; Ross, 1970; Ross and others, 1973; Ehlig and others, 1975; Matthews, 1976; Ramirez, 1983). Juxtaposition of provinces having great dissimilarities in geochemistry or paleomagnetic orientations, together with the slicing of long slivers of oceanic crust, are evidences of strike-slip faulting in the Semangko fault zone (Page and others, 1979).

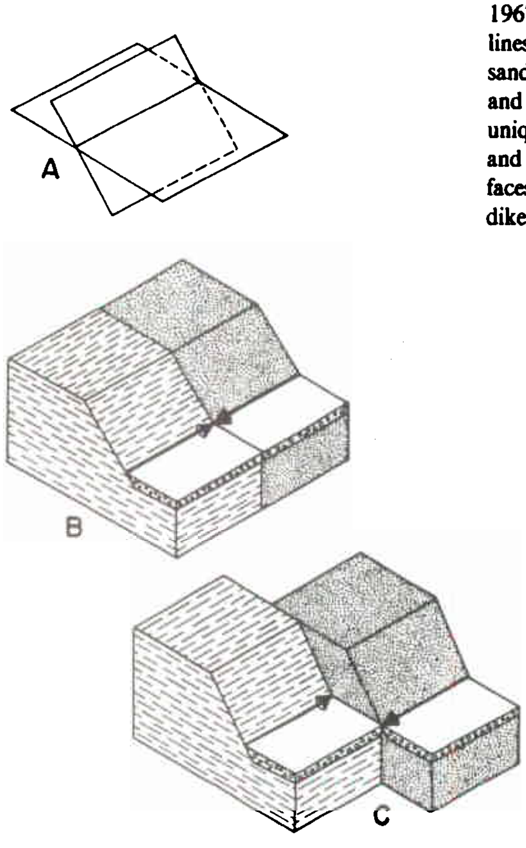


Figure 4. Diagrams illustrating the concept of piercing points of a displaced geological line. (A) Intersection of two surfaces defines a line. Block diagrams show a vertical fault between sandstone (stippled) and shale (dashed). The sloping surface represents a stream bank. (B) The intersection between the stream bank surface and the upper surface of the alluvium layer defines a line. The intersection of that line with the fault defines a piercing point. (C) Amount and direction of slip of the fault blocks can be determined from the displacement of the piercing points.

Syntectonic igneous activity is notably sparse along strike-slip faults except locally in zones of transtension (Şengör, 1979; Şengör and Canitez, 1982; Hempton and Dunne, 1984) and along trench-linked strike-slip faults which are imbedded in the volcanic arc. Major strike-slip faults also lack a conspicuous, peaked heat-flow anomaly which should be present if the average dynamic frictional resistance exceeds a hundred bars in the seismogenic zone (Brune and others, 1969; Lachenbruch and Sass, 1980). The heat flow in the Coast Ranges, within which the San Andreas fault is located, however, is moderately high and is comparable to that in the Basin and Range province (Lachenbruch and Sass, 1980). Where the San Andreas fault is transformed at its south end to a spreading rift, the heat flow is very high and spatially associated with young rhyolitic volcanic rocks and geothermal activity (Müffler and White, 1969; Robinson and others, 1976).

Separation, Slip, and Trace Slip

In any discussion of how to determine the amount and direction of displacement on any fault, it is essential to distinguish carefully between the terms "slip" and "separation" (Hill, 1959; Crowell, 1959, 1962). Separation is the apparent displacement of a plane in two dimensions and has geometric significance only, whereas slip, the actual displacement of lines, has kinematic significance. Separation has clear and long-standing geometric meaning in the context of faulting and should be retained for use in its original context. Regrettably, however, some writers (Rodgers, 1980; Segall and Pollard, 1980; Barka and Cadinsky-Cade, 198) have given the word different meaning in association with strike-slip faults.

Of equal importance is the recognition and understanding of "trace slip" (Beckwith, 1941;

Crowell, 1962) where the displacement is parallel to the trace of reference markers, such as bedding, on the fault surface (Fig. 5). Trace slip especially hinders recognition of strike-slip faults in seismic sections which are sensitive only to vertical separation. Strike slip out of the plane of the section, parallel to the line or trace of the layering on the fault surface, may displace the layers laterally but not vertically. I suspect that many strike-slip faults have gone unrecognized in many areas of nearly flat-lying strata for this reason and will be eventually "discovered" when sufficient regional data are accumulated. Many orogen-parallel strike-slip faults have gone unnoticed also because of regional trace slip, or because geologic reference lines make low angles with the fault.

A criterion which should force the suspicion of the existence of a strike-slip fault in seismic or structure sections is the presence of unresolved space problems after palinspastic restoration of the structural geometry has been attempted (Sylvester and Smith, 1976; Harding and Lowell, 1979; Harding, 1983b), problems which necessitate movement of rock volume in and out of the plane of the section. It is this factor that makes it difficult or impossible to achieve balance in structural cross sections. Conversely, lack of balance in faulted sequences is one of the keys to recognition that strike slip must be considered in the deformation picture (Sylvester and Smith, 1976). A second criterion is the presence of several adjacent, nearly vertical faults which show apparent, opposing senses of vertical separation (Harding and Tuminas, 1988, their fig. 11). The vertical separations may indeed represent true slip, but it is difficult to construct a plausible tectonic story of normal and reverse faulting which alternates over a short distance in time and space. In my experience, two-dimensional separations on steeply dipping faults should be regarded with considerable suspicion and may be better interpreted as vertical components of three-dimensional strike or oblique slip.

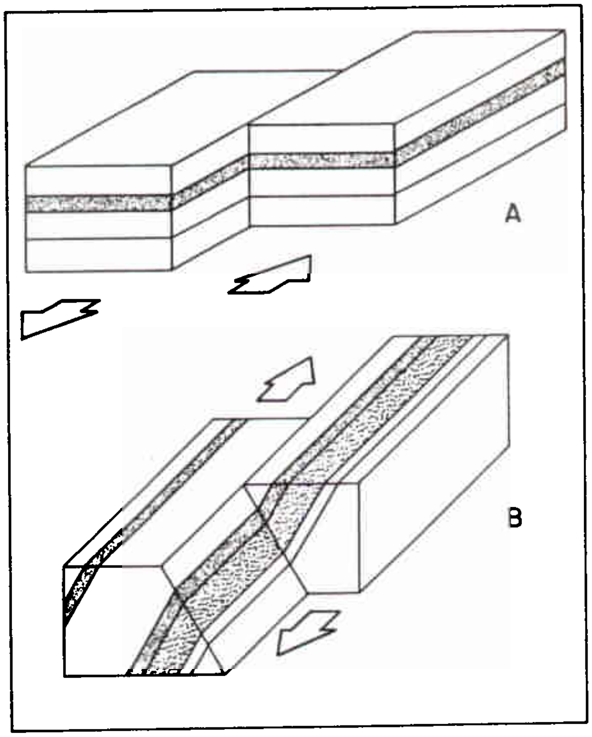


Figure 5. Diagrams illustrating the concept of trace slip. (A) Slip, which is parallel to the trace that the bed makes on the vertical fault surface, is not revealed in a vertical cross section perpendicular to the fault, (B), even if the beds and fault surface are not orthogonal to the ground surface or to each other.

MECHANICS OF STRIKE-SLIP FAULTING

The presence of shortening structures such as folds and thrust faults, of extensile structures including normal faults and dikes, and structures representing horizontal shear on nearly vertical surfaces—all together in a strike-slip regime—is involved in the concept of "wrench tectonics" (Anderson, 1942; Moody and Hill, 1956; Wilcox and others, 1973). The complexity and variety of these structures individually or in combination have three main aspects (Naylor and others, 1986): (1) the en echelon nature of faults and folds; (2) complications due to components of reverse or normal dip-slip on the basement fault; and (3) lateral offsets of basement-involved strike-slip faults which create local extensile or shortening structures. Two principal mechanisms explain the geometric and dynamic relations among these faults and associated structures: pure shear, sometimes called the Coulomb-Anderson model, and simple or "direct" shear (Fig. 6). Pure shear produces relatively short, typically conjugate sets of strike-slip faults which help to accommodate the brittle component of strain in tectonic regimes of crustal shortening, such as overthrust belts. Bulk pure shear is irrotational and has an orthorhombic symmetry. Simple shear has a monoclinic symmetry and rotational component of bulk strain and accounts for the kinematics of strike-slip faults at all dimensions (Tchalenko, 1970).

Pure Shear

This mechanism was originally proposed by Anderson (1905) to explain the orientations of faults relative to a triaxial stress field in a homogeneous medium. In the case of strike-slip faults, it predicts that a conjugate set of complementary sinistral and dextral strike-slip faults will form at an angle of ϕ and $-\phi$ about the shortening direction (Fig. 6a), where ϕ is the angle of internal friction. It predicts that extension fractures or normal faults will form perpendicular to elongation axis, and that folds and thrust faults will form perpendicular to the shortening axis. Notice in Figure 6 that the orientations of structures are shown relative to incremental strain axes rather than to stress axes. In most published diagrams, the principal stress axes are assumed to be parallel to the principal strain axes because of the assumed homogeneity of the medium and of the instantaneous strain, theoretical assumptions which are rarely achieved by heterogeneous rocks in prolonged natural deformation.[1]

The conjugate faults can accommodate irrotational bulk strain as long as they operate simultaneously; otherwise space problems ensue which can be solved only by rotation and alternating differential slip on each of the conjugate faults (Fig. 7). Strike-slip faults in domains of pure shear do not evince offsets measurable in hundreds of kilometers, because of room problems that result from convergence of large crustal masses (Fig. 8). Anderson (1941) himself recognized the space problem caused by the convergence of large crustal masses if the faults have a high junction angle and if they do not operate simultaneously (Fig. 8), but he either did not recognize or acknowledge the role of simple

[1]Readers may wonder about the lack of the term "compression" in this paper, as in the context of "compressional structures." Many geological writers use "compression" indiscriminately for both stress and strain. I follow the convention from rock mechanics that "tension" and "compression" are terms that should be used in the context of *stress*, whereas the corresponding *strain* terms are "extension," "elongation," or "lengthening" and "contraction" or "shortening" or even "constriction."

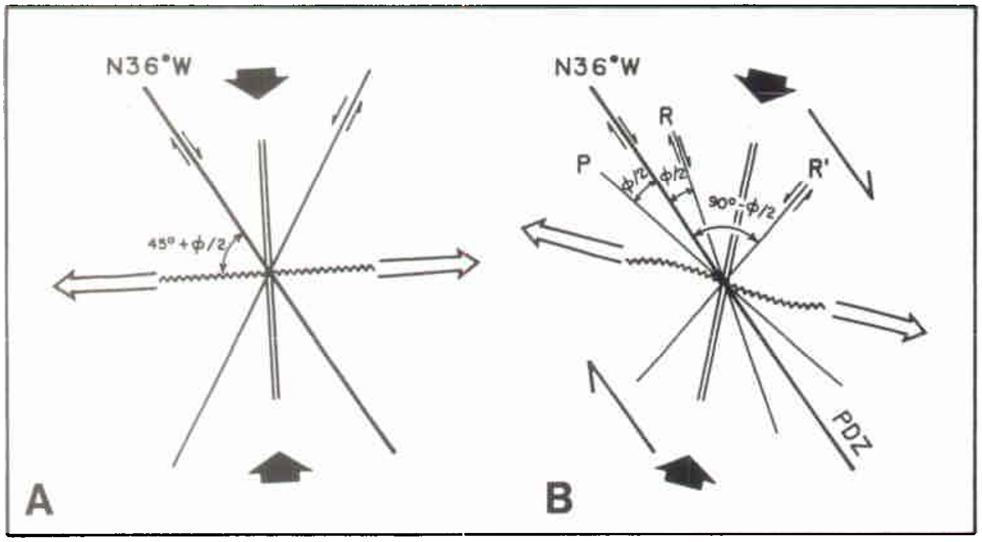


Figure 6. Plan view of geometric relations among structures according to two-dimensional, strike-slip, tectonic models for a vertical fault which strikes N36°W (adapted with modifications from Aydin and Page, 1984). (A) Coulomb-Anderson model of pure shear; (B) Riedel model of right simple shear. Double parallel line represents orientation of extension (T) fractures; wavy line represents orientation of fold axes. P = P fracture, R and R′ are synthetic and antithetic shears, respectively; PDZ = principal displacement zone; ϕ = angle of internal friction. Short black arrows = shortening axis; open arrows = axis of lengthening.

shear in strike-slip faulting of the crust. In fact, in discussion of Kennedy's (1946) paper about strike slip on the Great Glen fault in Scotland, Anderson maintained that strike-slip faults occur in complementary X-shaped pairs, 50° apart, that shear on the Great Glen fault is left-lateral, and that equally great right-lateral faults should be found nearby. Complementary right-lateral faults having a magnitude of strike-slip equivalent to that of the Great Glen fault have not been found in Scotland, because the Great Glen fault is a product of simple shear and not pure shear as Anderson assumed.

The requirements of the pure shear mechanism have been missed by many geologists who have misapplied it to field areas of strike-slip faulting, yielding simplistic misinterpretations of the kinematics and dynamics of strike slip. In southern California, for example, some writers assumed that the Garlock and Big Pine faults were the left-lateral counterparts to the right-lateral San Andreas fault in a pure shear system (Fig. 9), even though the apical angle between them across the presumed direction of shortening is about 120° rather than the standard 60° (Hill and Dibblee, 1953). The interpretation was based on the X-shaped pattern and the senses of slip on the three faults. It is an incorrect interpretation, however, when the amount and timing of the fault displacement are considered. The San Andreas fault came into existence about 24 Ma, accumulating 330 km of dextral slip (Crowell, 1979), whereas movement on the Garlock fault commenced 10 m.y. ago and totals about 60 km of left slip (Davis and Burchfiel, 1973; Burbank and Whistler, 1987; Loomis and Burbank, 1988). The Big Pine fault had an Oligocene episode of displacement, mainly dip-slip, and a post–late Miocene episode of left slip (Crowell, 1962).

The notions of McKinistry (1953) and Moody and Hill (1956), which simplistically explained strike-slip faulting on all scales in terms of the Coulomb-Anderson mechanism, did much to confuse understanding of strike-slip tectonics for several decades in my opinion, and judging from critical discussions of their papers (Prucha, 1964; Maxwell and Wise, 1958; Laubscher, 1958). The Coulomb-Anderson mechanism is an easy concept to grasp, and many writers did so, misapplying the concept to domains of simple shear, but finding that all of the structures were not explained, wondering why, and struggling with *ad hoc* arguments to explain the lack of the conjugate set of equal magnitude that the Andersonian theory promises (Rod, 1958; Allen and others, 1960). They did this in spite of the fact that a considerable volume of literature was available, especially from the 1920s, about simple shear experiments and their relation to observed structures in the field (Mead, 1920; Fath, 1920; Hubbert, 1928; Brown, 1928; Cloos, 1928; Riedel, 1929).

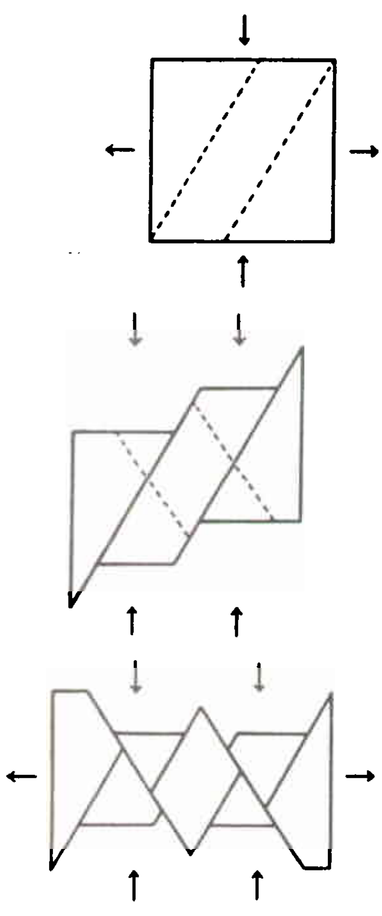


Figure 7. Bulk pure shear strain by alternating slip on conjugate shears with rigid body rotation (from Anderson, 1905).

Figure 8. Space problems with conjugate slip in bulk pure shear (adapted with modifications from Ramsay, 1979). (A) Synchronous faulting; (B) alternating fault activity.

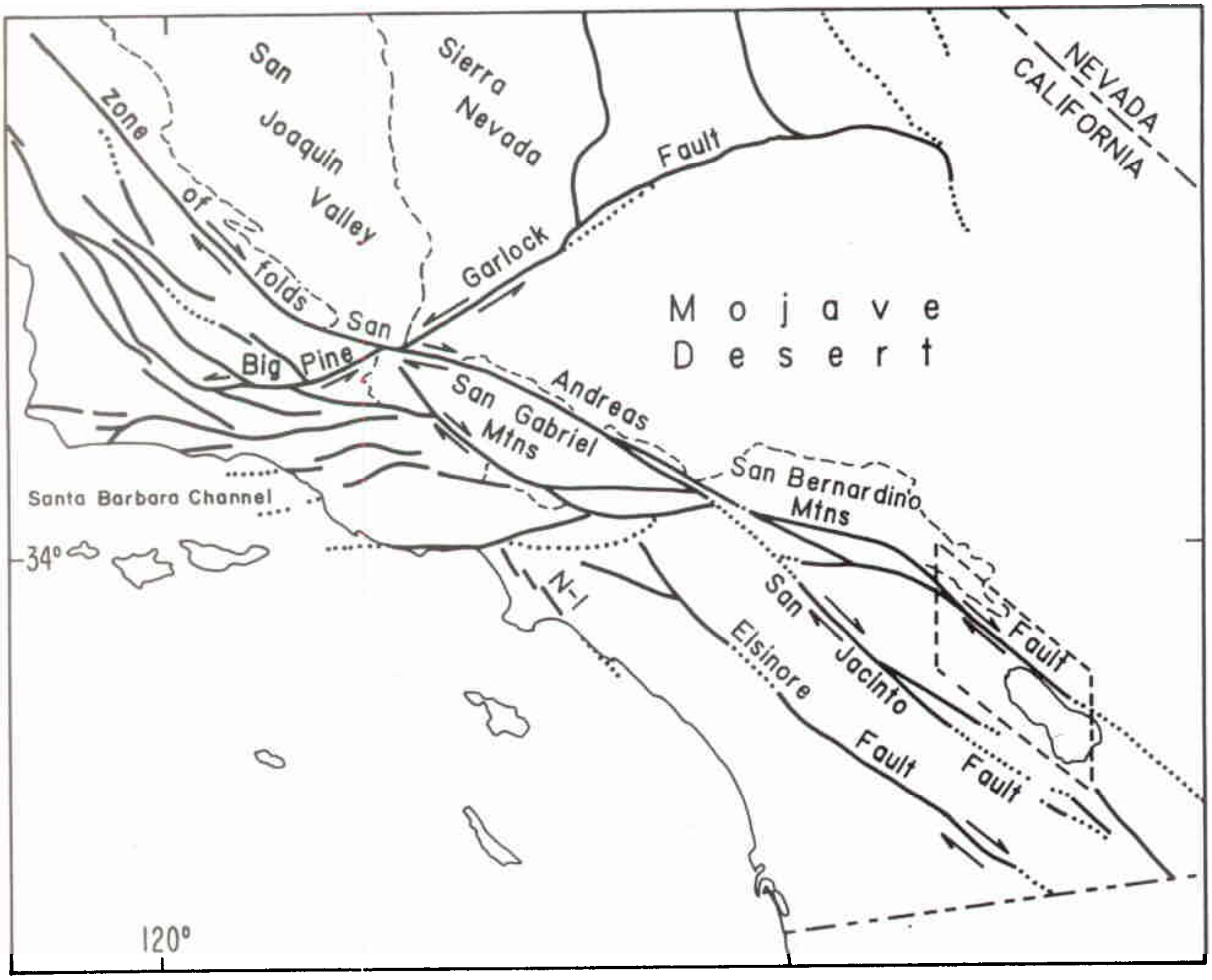


Figure 9. Principal faults in southern California. The "big bend" in the San Andreas fault is located at the south end of the San Joaquin Valley and at the northwest end of the San Gabriel Mountains where the Big Pine and Garlock faults intersect the fault. The dashed parallelogram indicates the area shown in the map in Figure 23.

Domains of regional pure shear with conjugate strike-slip faults having many kilometers of displacement are real, however, and are well documented. Typically these domains are in fold-thrust belts where conjugate strike-slip faults transect the fold trends. A good example is in the Wopmay orogen of northern Canada where the Asiak foreland thrust-fold belt (Hoffman and St-Onge, 1981; Tirrul, 1982, 1984) is offset by a conjugate set of east-northeast–striking, right-slip faults which are typically longer than 50 km, and by west-northwest–striking, left-slip faults (Figs. 10 and 11). The faults are developed at all scales and are in mutually exclusive domains for a given scale (Tirrul, 1984). As much as 15 km of displacement has occurred across some of the faults. Geometric arguments and palinspastic map reconstructions indicate that both fault sets formed initially at 25°–30° to the east-west shortening direction and rotated thereafter about a vertical axis away from it (Tirrul, 1984), as Freund (1970a) postulated for conjugate faults in the Sistan District of Iran (see also Tirrul and others, 1983). The bulk strain approaches the magnitude of the regional pure shear: east-west shortening is up to 25% with north-south extension (Tirrul, 1984). Other notable regional domains of strike slip caused by bulk pure shear in areas of crustal convergence have been identified in the Apennines of central Italy (Lavecchia and Pialli, 1980, 1981), the Carpathian Pannonian basin in southeastern Europe (Royden and others, 1982), the Makran of southwest Pakistan (Platt and others, 1988), the southern Chilean Andes (Katz, 1962), and in the accretionary prism of the Aleutian trench (Lewis and others, 1988).

Simple Shear

The major strike-slip faults of the world are in domains of simple shear which may be thousands of kilometers long and tens of kilometers wide, and they have displacements measured in hundreds of kilometers. Within the domain of simple shear, the most recently active strand may be a zone of active faulting that is only a few meters wide.

Simple shear has a monoclinic symmetry of strain because it is rotational, and a greater variety of structures forms in simple shear than in pure shear (Fig. 6b). The structures typically form en echelon arrangements in relatively narrow zones (Fig. 12). Five sets of fractures form in simple shear in model experiments, in experimental deformation of homogeneous rocks under confining pressure, and in alluvium deformed by surface rupturing during earthquakes (Fig. 6b): (1) Riedel (R) shears (Tchalenko, 1970) or "synthetic" (Cloos, 1928) or "pinnate" (Ma and Deng, 1965) strike-slip faults; (2) conjugate Riedel (R′) shears (Tchalenko, 1970) or "antithetic" (Cloos, 1928) strike-slip faults; (3) secondary synthetic strike-slip faults at an angle of $-\phi/2$ to the direction of applied shear (P shears of Skempton, 1966; Tchalenko, 1970;

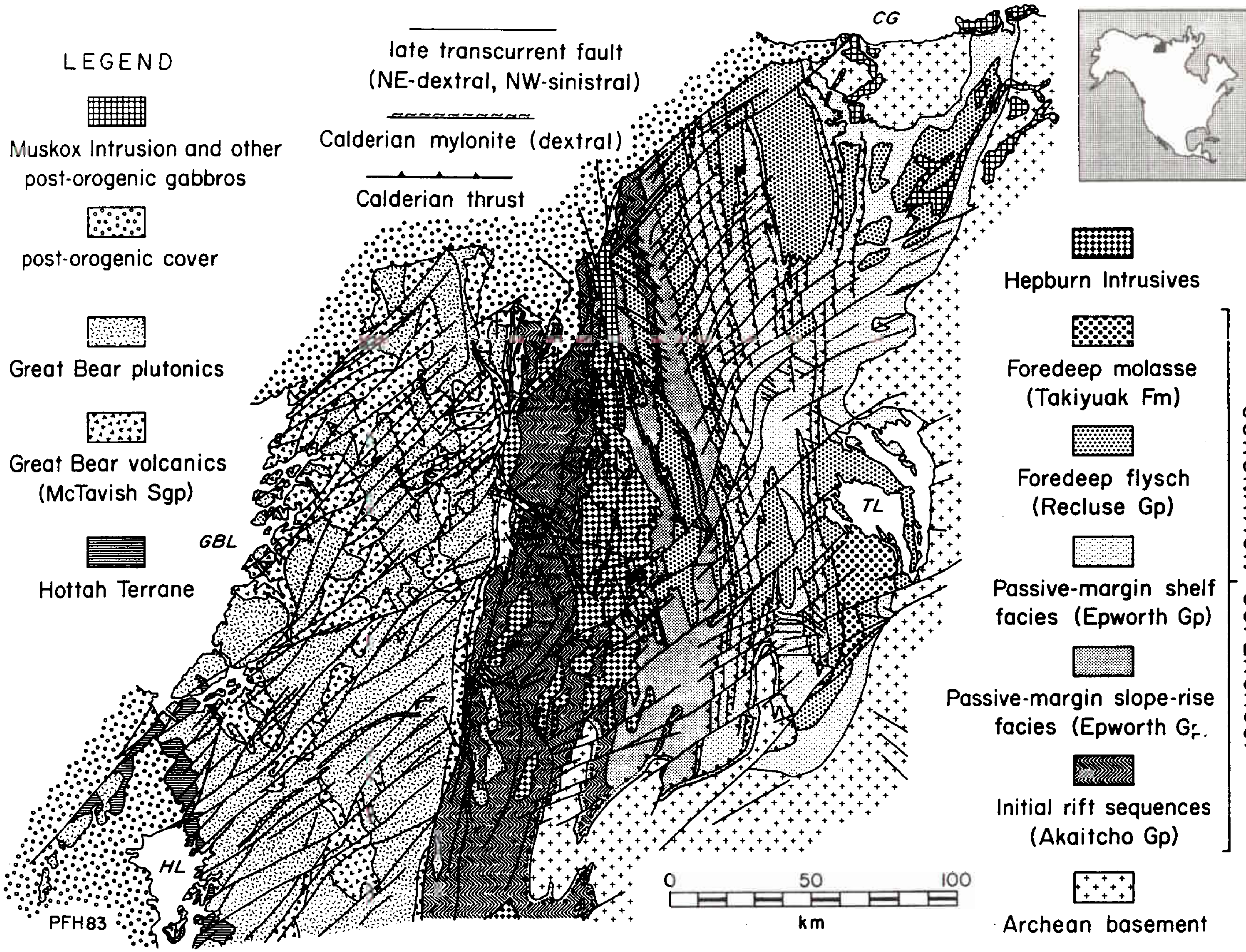
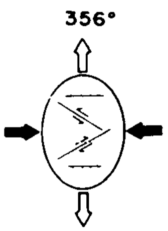
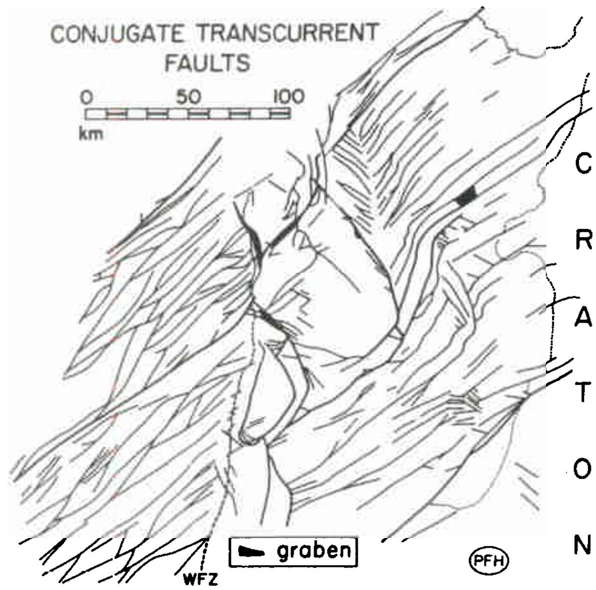


Figure 10. Simplified geologic map of the northern two-thirds of the Wopmay orogen, showing regional setting of conjugate strike-slip faults which transect the Asiak fold-thrust belt (from Hoffman and others, 1984). Northeast-striking faults are right-slip faults; northwest-striking faults are left-slip faults. Inset shows location of map. Prominent water bodies are: CG, Coronation Gulf; GBL, Great Bear Lake; TL, Takijug Lake; HL, Hottah Lake.

Figure 11. Distribution of strike-slip faults in the northern Wopmay orogen. As indicated in the ellipsoidal inset, all northeast-striking faults are dextral, northwest faults are sinistral, and east-west faults are normal. WFZ = Wopmay fault zone. Dark arrows indicate shortening axis; open arrows indicate the extension axis (P. F. Hoffman, M. R. St-Onge, and R. Tirrul, Geological Survey of Canada, unpub.).





Tchalenko and Ambraseys, 1970); (4) extension fractures (T fractures of Tchalenko and Ambraseys, 1970) or normal faults which develop at about 45° to the principal displacement zone; and (5) faults parallel to the principal displacement zone (Y shears of Morgenstern and Tchalenko, 1967).

The laboratory studies (Cloos, 1928; Riedel, 1929; Tchalenko, 1979; Wilcox and others, 1973) simulated a rigid basement by two stiff boards overlain by a cake of unbroken clay, analogous to a cover of sedimentary rocks. When the boards slipped parallel to one another at depth, the first-formed structures in the overlying clay were en echelon R shears whose overstepping sense is directly related to the sense of slip of the underlying boards: that is, left-stepping in right simple shear, and right-stepping in left simple shear, although reversals locally occurred. The width of the zone of fracturing in plan view is a function of the thickness of the clay cake. These experiments have been repeated numerous times by academic and industry scientists, with results that vary according to the type of material being sheared (Skempton, 1966; Morgenstern and Tchalenko, 1967; Emmons, 1969; Hoeppener and others, 1969; Lowell, 1972; Courtillot and others, 1974; Mandl and others, 1977; Groshong and Rodgers, 1978; Graham, 1978; Bartlett and others, 1981; Gamond, 1983; Deng and Zhang, 1984; Macdonald and others, 1986; Hempton and Neher, 1986; Naylor and others, 1986). The same arrangement of structures also forms in alluvium deformed by surface rupturing during earthquakes (Gianella and Callaghan, 1934; Florensov and Solenenko, 1963; Brown and others, 1967; Clark, 1968; Tchalenko and Ambraseys, 1970; Philip and Megard, 1977; Terres and Sylvester, 1981; Sharp and others, 1972; Deng and Zhang, 1984).

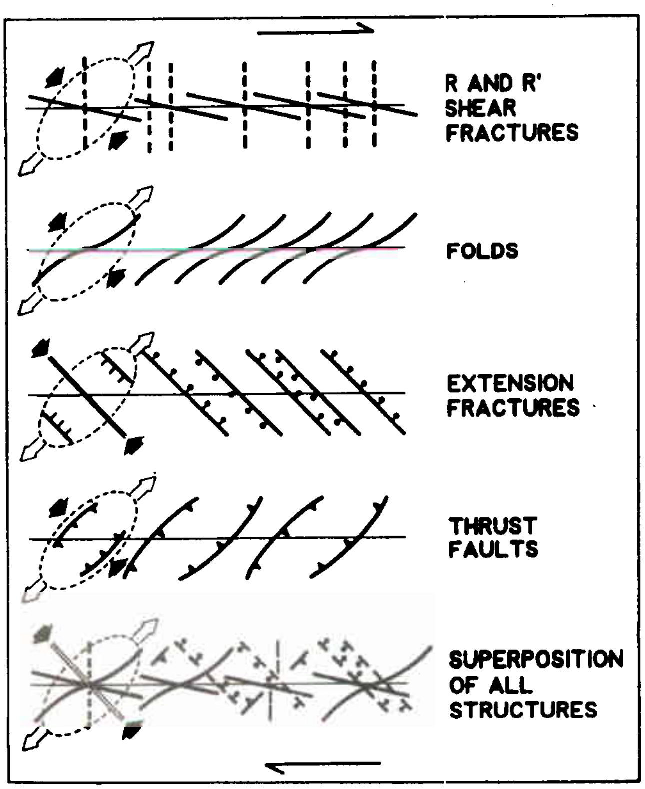


Figure 12. Orientation of folds and faults in bulk right simple shear.

The sense of strike slip along the R, P, and Y shears is the same as that of the basement fault, whereas that of the R′ shear is opposite. All of the faults, except the thrust faults, are nearly vertical when they form. The R and R′ shears make angles of $\phi/2$ and $90°-\phi/2$, respectively, with the principal displacement zone (Fig. 6b), where ϕ is the angle of internal friction. This means that the R shears strike from 15° to 20° to the principal displacement zone; and the R′ shears, from 60° to 75° (Tchalenko and Ambraseys, 1970, p. 56). In sand model experiments, the actual angle depends on the thickness of "overburden" above the "basement" fault, being at a low angle when the "overburden" is thin, and being greater than 15° when the "overburden" is thick (Naylor and others, 1986). The extension fractures bisect the angle between the R and R′ shears and are oriented parallel to the incremental axis of shortening and at an angle of 45° to the direction of applied shear. R′ shears rarely develop in nature (Keller and others, 1982) except where there is a substantial overlap between adjacent R shears (Tchalenko and Ambraseys, 1970; Naylor and others, 1986).

Soon after the initial formation of the R shears in the model studies, the incremental strain field is locally modified in the deformation zone, giving rise to the development of P shears and short-lived splay faults (Naylor and others, 1986). Splay faults form at the tips of the R shears and curve toward parallelism with the extension fractures so that an R shear will be a strike-slip fault in the central part of the deformation zone, but it will be a normal fault with little or no strike slip at its extremities. The P shears form as a consequence of the reduction of shearing resistance along the R shears, so that all of the basement displacement cannot be taken up on the discontinuous en eche' R shears in the overburden (Tchalenko, 1970). Thus, typically between two overlapping R shears, the shortening axis will be reoriented toward the R shear, producing new faults in this local strain field that strike at an angle of $-\phi/2$ to the principal displacement zone (Naylor and others, 1986).

The most advanced stage of the deformation yields a principal displacement zone of braided or anastomosing, vertical fractures wherein the main surficial shear strain is centered on an irregular, through-going fracture whose path may be composed variously of the R and P fractures (Naylor and others, 1986). Tchalenko (1970) also showed that the assemblage of fractures formed in simple shear is geometrically similar through the microscopic to the macroscopic scales.

Folds and thrust faults form initially perpendicular to the axis of shortening (Fig. 12), and thus, also at an angle of 45° to the principal displacement zone. If deformation continues, then the fold axes will rotate according to the amount of shearing: as much as 19° for a shear strain of unity (Ramsay, 1967, p. 88).

The theory of simple shear in the strict sense (Ramsay, 1967, 1980) is probably not applicable to any large part of the Earth's crust with any mathematical rigor because of uncertainties in mechanical behavior and strain profile and because of the heterogeneous nature of rocks (Aydin and Page, 1984). "Simple shear," however, is a general approximation of the theoretical concept when applied to strike-slip faulting as was done by Tchalenko (1970) and by Wilcox and others (1973). Typically it is applied to any zone between strike-slip faults where crustal strain is the direct result of pervasive horizontal shear in a consistent sense and direction (Aydin and Page, 1984). Even though the magnitude and rate of shear strain may vary greatly from place to place within the zone, the entire domain of strike slip is made up of many subzones, each undergoing simple shear at a particular rate (Aydin and Page, 1984).

Only a few good geological field examples (Keller and others, 1982; Erdlac and Anderson, 1982) show the idealized among R and R′ fractures depicted in Figure 6b, because natural structures develop sequentially rather than nearly instantaneously as they do in laboratory models and earthquakes, because rocks are heterogeneous, and because early-formed structures may be internally rotated with protracted shear strain. Thus, many of the early-formed fractures may be cut off and will be inactive during subsequent shear on the principal displacement zone. The resultant plethora of R, R′, P, T, and Y fractures in nature is perplexing to unravel sequentially, especially in strike-slip fault zones that have had a long history of movement. The style of natural structures is also affected by convergent or divergent strike slip, as is discussed more fully in a subsequent section. The end result may be that some of the folds and faults will form or be rotated into orientations parallel to the principal displacement zone. The resultant faults will look like dip-slip growth faults, and the folds will have all of the characteristics of drape folds when viewed in two-dimensional seismic or structure sections (Harding and others, 1985).

In some cases, the strike-slip at depth is sufficiently small or deep that a through-going fracture fails to develop at the surface (Naylor and others, 1986). Instead, only a long, narrow zone of en echelon normal faults or R shears forms (Erdlac and Anderson, 1982), perhaps associated with en echelon folds as is illustrated, for example, in the Columbus basin of offshore Trinidad (Leonard, 1983). En echelon arrays composed only of extension fractures, gash fractures, or normal faults are common above a buried strike-slip fault. Thus, most of the individual faults of the Lake Basin and Nye-Bowler fault zones in Montana are normal faults that strike ~45° to the trend of the strike-slip fault zone (Wilson, 1936; Alpha and Fanshawe, 1954; Smith, 1965). R shears and even extension fractures may have significant strike slip on them, as does the Newport-Inglewood zone in southern California (Barrows, 1974; Yeats, 1973; Harding, 1973), where as much as 800 m of right slip has occurred across the fault zone, truncating early-formed en echelon folds into half-anticlines and domes. The simple shear mechanism for the Newport-Inglewood zone was modeled and clearly understood by Ferguson and Willis (1924) and later by Wilcox and others (1973).

R fractures have a helicoidal shape in three dimensions (Fig. 13), which is a consequence of three factors (Naylor and others, 1986): (1) the en echelon nature of the shears at the surface; (2) their concave-upward geometry when formed in shear without components of convergence or divergence; and (3) the need to join a single basement fault at depth. In cross section, such a fault looks like an upthrust at depth and like a landslide near the surface: Crowell

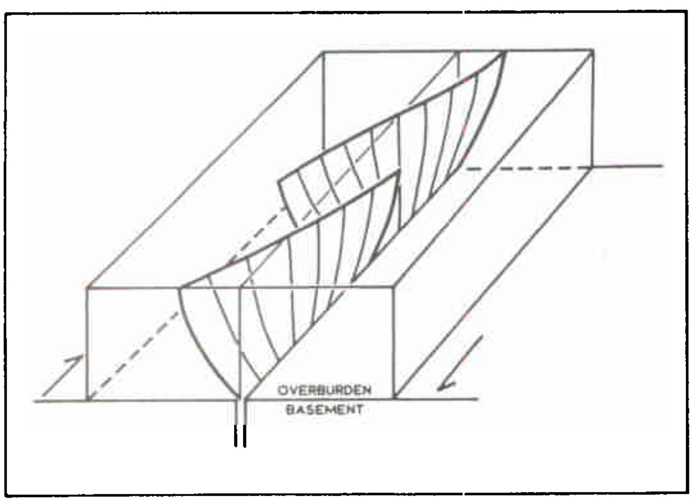


Figure 13. Helicoidal form of individual Riedel shears in right simple shear, reconstructed from horizontal serial sections in sandbox model experiments (redrawn from Naylor and others, 1986). Reproduced with permission of *Journal of Structural Geology*.

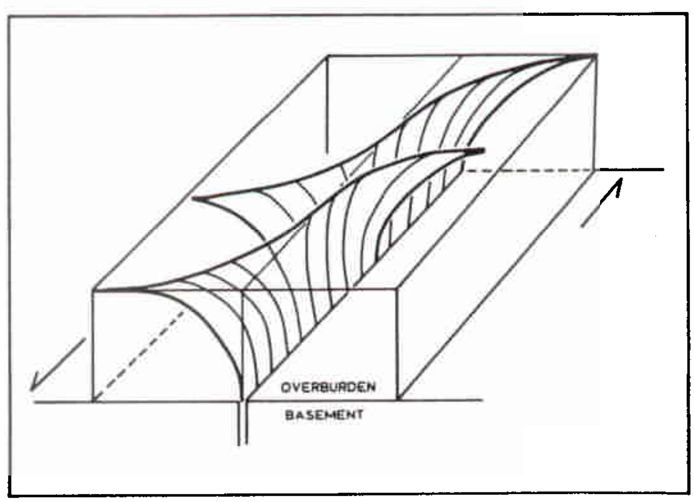


Figure 14. Helicoidal form of axial surfaces of two en echelon folds in left simple shear. Inspired from analogous diagram for R fractures (Naylor and others, 1986) and from descriptions of natural folds (Sylvester and Smith, 1976; Gamond and Odonne, 1983).

(1982a, p. 40) informally refers to this fault form as a "slust" or a "thride." The imprecise and non-euphonic terminology represents the difficulty of characterizing and describing with a single word or phrase an oblique-slip fault, the surface of which may be vertical at depth and convex-upward but which flattens to horizontal at the surface (Allen, 1965; Lowell, 1972; Wilcox and others, 1973; Sylvester and Smith, 1976) or which is nearly vertical at the surface and concave-upward but flattens at depth into the main strike-slip fault (Harding, 1983a; Harding and others, 1985; Naylor and others, 1986).

EN ECHELON FOLDS

Folds associated with strike-slip faults are typically arranged in en echelon pattern oblique to the principal direction of shear, and they have received much attention from the petroleum industry, because they are attractive prospective traps for hydrocarbons (Harding, 1974; Dibblee, 1977b; Harding and Lowell, 1979; Harding and Tuminas, 1988). Typically, en echelon folds are distributed in a relatively narrow and persistent zone above or adjacent to a master strike-slip fault. They may form in a broad zone between two major strike-slip faults as they do in the East Bay Hills in the San Francisco Bay region of California (Aydin and Page, 1984). The presence of en echelon folds or faults parallel to a zone of deformation, however, is not restricted to strike-slip faulting (Sherill, 1929; Campbell, 1958; Christie-Blick and Biddle, 1985). En echelon folds may reflect the complex influence of basement structures at depth, or they may represent the superposition of differently oriented folds in time and space (Harding, 1988).

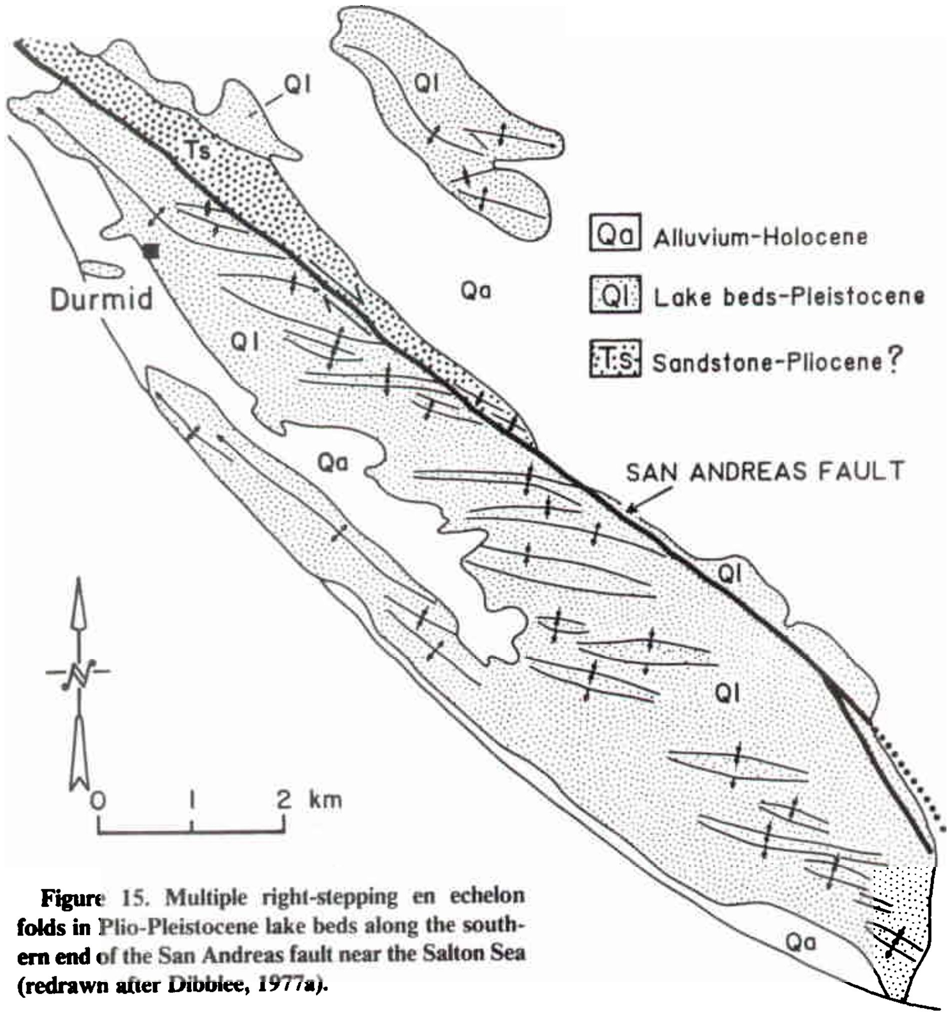


Figure 15. Multiple right-stepping en echelon folds in Plio-Pleistocene lake beds along the southern end of the San Andreas fault near the Salton Sea (redrawn after Dibblee, 1977a).

Ideally, the crestal traces of en echelon folds should make an angle of 45° in plan view to the shear direction (Fig. 6b), representing the shortening component of the bulk strain, but trends of real folds are gently twisted and vary from 10° to 35° to the strike of the fault zone (Harding and Lowell, 1979). In three dimensions (Fig. 14), the axial surfaces of echelon folds in a sequence of strata overlying a rigid basement are nearly vertical and parallel to the fault at basement level, but higher in the overburden, they flatten upward and twist away from the strike of the fault (Gamond and Odonne, 1984; Koral, 1983), and they plunge away from the principal displacement zone (Harding and Tuminas, 1988). The axial surface of an en echelon fold has a helical geometry, similar to that of an R shear (Naylor and others, 1986), except that an R shear steepens upward, whereas a fold axial surface flattens upward. This implies that the observed angular relation between en echelon fold axes and principal displacement zone in plan view may depend locally on the depth of erosion, as well as on the amount of internal rotation within the shear zone.

En echelon folds have been popularly called "drag folds" (Moody, 1973) where they curve into parallelism with a strike-slip fault, but the term and notion are misnomers in the context of strike-slip faulting, because the folds are born in an en echelon orientation as many model studies show (Pavoni, 1961b; Wilcox and others, 1973; Dubey, 1980). Another reason why natural folds do not always fit the optimal orientation predicted by heterogeneous simple shear, therefore, is that they may be rotated or internally sheared by piecemeal slip on R fractures and smeared, thereby, into a "dragged" appearance.

En echelon folds are useful structural indicators, because they tell three things about the associated fault and its related structures: (1) that a strike-slip fault is probably nearby laterally or at depth; (2) the direction of slip on that fault by the overstepping direction of the folds; and (3) the expected orientations of related faults. These concepts are discussed in the next two paragraphs.

Clay-model studies of en echelon folds imply that the folds form symmetrically above the principal zone of displacement (Wilcox and others, 1973; Odonne and Vialon, 1983). At the south end of the San Andreas fault (Fig. 15), however, en echelon folds are rare on the northeast side of the fault, relative to their abundance on the southwest side (Babcock, 1974; Dibblee, 1977a). Their apparent lack on the northeast side is partly due to the fact that they are covered by younger deposits; but it is just as reasonable to expect that folds which formed on the northeast side were displaced out of the picture by strike slip; or that the rocks on the northeast side were incapable of folding as readily as the thin-bedded, gypsiferous, lacustrine strata on the southwest side. Rather than project the most likely locus of the strike-slip fault symmetrically beneath the field of folds, as one would tend to do in analogy with clay-model experiments, it is better to infer only that a strike-slip fault is nearby or at depth.

The direction of the horizontal movement on the strike-slip fault is revealed by the stepping direction of the folds (Fig. 16): Right-stepping folds form in right slip; left-stepping folds form in left slip. Then from the geometry of strain in simple shear, one can deduce the expected directions of associated Riedel shears, normal faults, and thrust faults (Fig. 6b).

In divergent strike slip, the crestal traces of folds are typically parallel to the principal displacement zone and have a parallel or relay pattern (Burkart and others, 1987) resembling those having formed in pure shear. The folds have a cross-sectional geometry and evolution similar to those of drape, or forced folds that form above normal fault blocks complete with anticlines or monoclinal knees next to, and parallel with, the relatively higher side of the principal displacement zone; they also have synclines or monoclinal "ankle flexures" adjacent to, and parallel with, the edge of the apparently downdropped fault block (Harding, 1974; Harding, 1983a; Harding and others, 1985). The faults beneath the folds may have the geometry of growth faults in cross section, and may well have played a "growth fault" role during syntectonic sedimentation (Sylvester and Smith, 1976). In convergent strike slip, en echelon folds may have any or all of the profile geometries found in other convergent tectonic styles, even thrust-associated types.

The factors which control the style and development of natural en echelon folds in simple shear have not been clearly defined. Initially they form perpendicular to the shortening axis within a shear couple in clay-model studies (Fig. 6b), but that relation may be an oversimplification, because their formation in model studies requires a ductile material or interlayered member, such as a thin sheet of tin foil, rubber, or plastic, which will deform continuously rather than by shearing (Mead, 1920; Wilcox and others, 1973). Thus, in order to form a neat arrangement of en echelon folds in clay, Wilcox and others (1973) placed a thin sheet of plastic film beneath the surface in some of their clay models. They found that the fold spacing, orientation, size, and rate of growth depended on the cohesion of the clay, the strain rate, and the degree of strike-slip convergence of the boards beneath the clay cake. Therefore, in convergent strike slip, a component of shortening is imposed above the zone of strike-slip deformation, and the folds form readily and in a distinct, consistent en echelon arrangement (Wilcox and others, 1973; Babcock, 1974).

Harding and Lowell (1979) depicted the maturation of a strike-slip fault zone and its associated structures from the evolutionary sequence of structures observed in clay-model studies (Wilcox and others, 1973). Early in the deformation history, an en echelon array of simple folds is formed (Fig. 17). With more deformation, R shears break the surface and deform the folds into domes, half anticlines, and synclines. Then the deformation zone broadens, and parts of the early-formed structures may be deeply eroded or faulted out of the picture, leaving vestiges of steep, secondary faults or parts of folds now cut complexly by thrust and normal faults and by R shears. Folding extends progressively farther from the principal displacement zone with increasing displacement over time. The largest amplitude folds are at depth near the fault, and the most recent folds are farthest from the fault at the margins of the deformation zone (Harding, 1976). Some tectonicists object that such a narrow zone of actual faulting can produce a zone of deformation as wide as 30 km. The main zone of modern strike slip is narrow indeed, but the zone of pervasive simple shear is wide and heterogeneous over geological time; at the latitude of central California, it may be from 500 km (Hamilton, 1961; Minster and Jordan, 1987; Ward, 1988) to 1,000 km wide (Atwater, 1970).

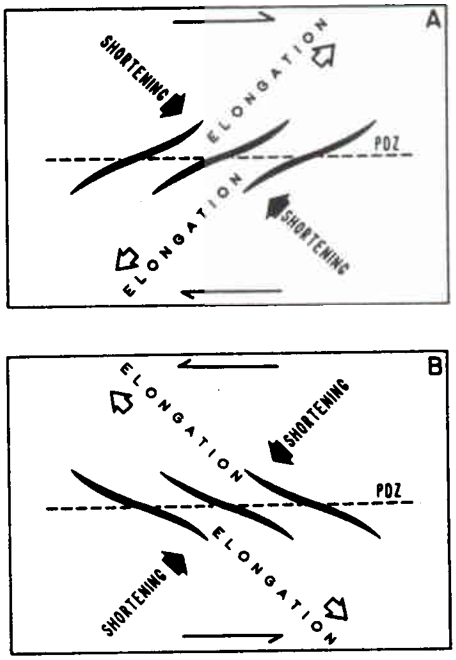


Figure 16. Geometry of en echelon folds in right simple shear (A) and left simple shear (B). PDZ = principal displacement zone. Bulk strain axes are labeled.

The simple shear mechanism and scale of the process involved in the formation of the remarkable series of en echelon folds associated with the San Andreas fault in the San Joaquin Valley was appreciated and understood clearly by Lawson (1921, p. 580): "With regard to the force which is responsible for the lateral movement on the these faults, all the evidence tends to show that there is a northward creep of the mass beneath the ocean with respect to the mainland mass or shield of the Sierras [sic], and that the region of the coast ranges represents the shear zone beneath these two great masses. The mass beneath the ocean, moving northward, presses against the northwesterly trending coast line and slides northwest along it, exerting, of course, great northeasterly pressure against it which is responsible for the parallel northwesterly trending folds in the more sharply folded portions of the coast ranges."

A strong case has been made by Harding (1976) in support of Lawson's (1921) and Hamilton's (1961) suggestion that the wide zone of contemporaneous faults and en echelon folds in San Joaquin Valley of southern California (Fig. 9) was produced by large-scale, convergent strike slip along the San Andreas fault rather than by contraction at a high angle, as was often assumed. The stratigraphy recorded by Harding in the folds, which are some of the most prolific producers of petroleum in North America, shows that they originated in mid-Miocene time when the San Andreas fault became active. Their growth increased and new folds developed during the Pliocene and Pleistocene epochs when the rate of movement increased on the San Andreas fault (Harding, 1976). If the Pliocene and early Pleistocene strata are palinspastically "unfolded," then the number of pre-Pliocene anticlines and synclines that remain in the underlying strata are minimal except on those structures that are closest to the fault. If the exercise is repeated for Miocene formations, then only rarely can a coincident pre-Miocene fold be shown to have existed at the same site, although underlying Paleogene and Cretaceous rocks are generally discordant with respect to the Miocene strata. In addition, the older, more tightly folded cores of anticlines nearest the San Andreas fault are commonly disrupted by young, southwest-dipping thrust faults or northeast-dipping reverse faults (Dibblee, 1973; Fuller and Real, 1983; Stein, 1983, 1984; Stein and King, 1984), faults which Harding (1976) regards as late-stage reactions to prolonged deformation.

An alternate hypothesis for the formation of the en echelon folds outside the narrow zone of strike slip in the San Joaquin Valley maintains that present-day shortening deformation is not controlled by distributed shear associated with drag on a high-friction San Andreas fault. Instead, based on interpretation of structural styles (Namson and Davis, 1988) together with borehole elongations and well breakouts, the present maximum stress is inferred to be oriented nearly perpendicular to the fault and therefore, "transpressive tectonics in central California can be better described as decoupled transcurrent and compressive deformation, operating simultaneously and largely independently" (Mount and Suppe, 1987, p. 1146). Separate and largely in-

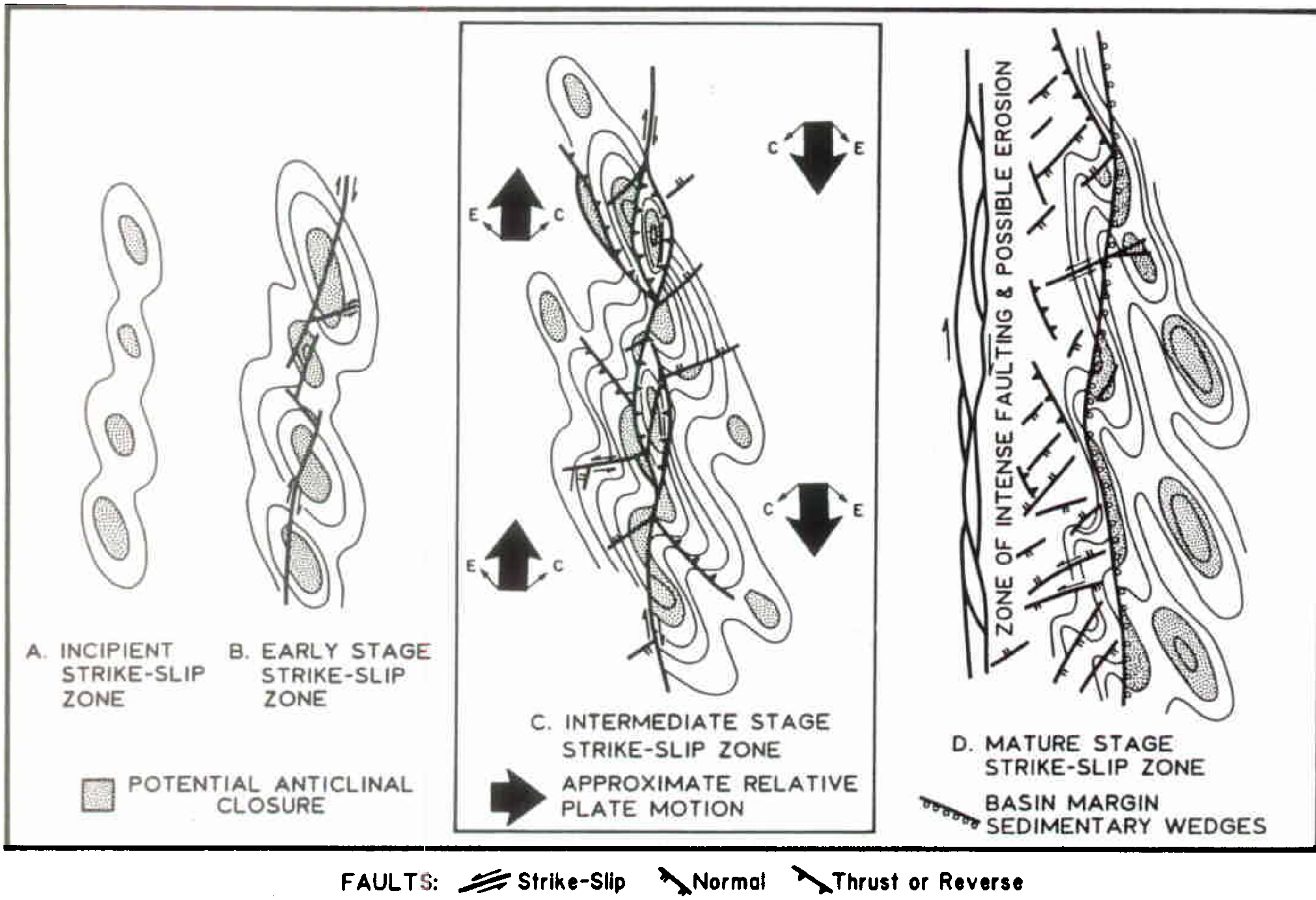


Figure 17. Schematic diagrams of structural assemblage associated with major strike-slip fault and their evolutionary history (redrawn from Harding and Lowell, 1979). Arrows C and E represent inferred principal directions of contraction and extension, respectively, that arise in the right simple shear couple represented by the heavy black arrows. Reproduced with permission of American Association of Petroleum Geologists.

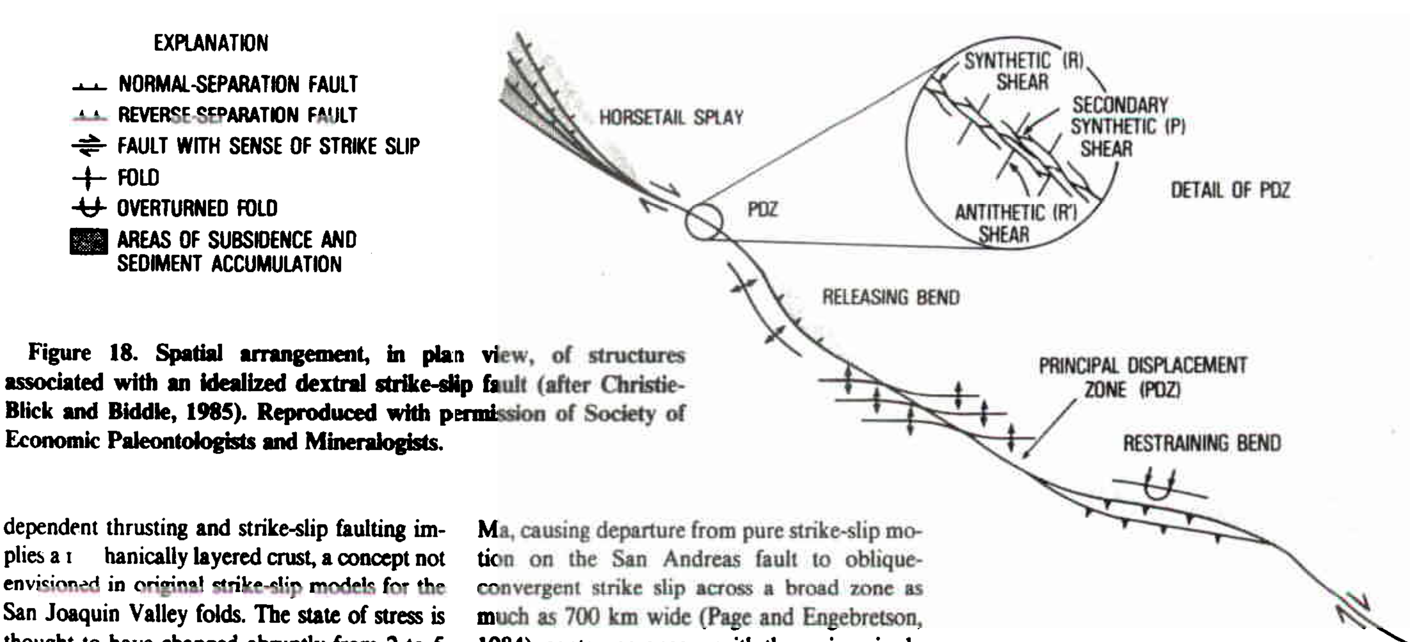


Figure 18. Spatial arrangement, in plan view, of structures associated with an idealized dextral strike-slip fault (after Christie-Blick and Biddle, 1985). Reproduced with permission of Society of Economic Paleontologists and Mineralogists.

dependent thrusting and strike-slip faulting implies a r hanically layered crust, a concept not envisioned in original strike-slip models for the San Joaquin Valley folds. The state of stress is thought to have changed abruptly from 2 to 5 Ma, causing departure from pure strike-slip motion on the San Andreas fault to oblique-convergent strike slip across a broad zone as much as 700 km wide (Page and Engebretson, 1984), contemporaneous with the main episode

A

B

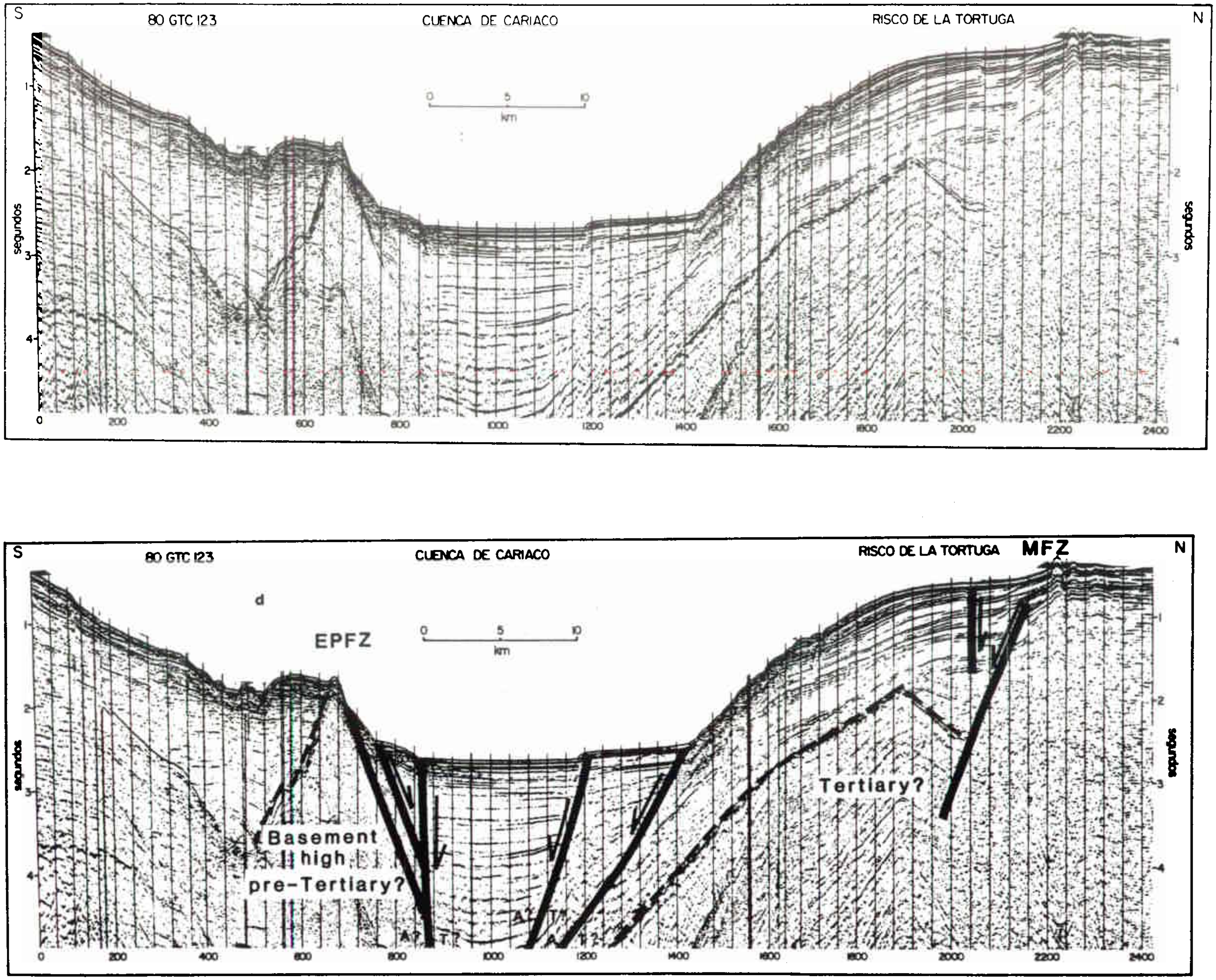


Figure 19. Unmigrated south-north seismic reflection profile across Caricao pull-apart basin (Cuenca de Cariaco) and the Tortuga Rise (Risco de la Tortuga), northern Venezuela (Schubert, 1986b). Vertical exaggeration is approximately 10:1. (A) Uninterpreted; (B) interpreted by C. Schubert, 1988. EPFZ = El Pilar fault zone; MFZ = Morón fault zone; segundos = seconds; A = movement away from observer; T = movement toward observer.

of crustal shortening and folding which created the California Coast Ranges. The young, contractile faults in the cores of the en echelon anticlines are therefore believed to accommodate recent interplate crustal shortening of about 5 mm/yr across the California Coast Ranges (Minster and Jordan, 1984), whereas the San Andreas fault takes up the much greater transform displacement (~35 mm/yr).

Excellent descriptions are in hand to explain the existence, timing, and developmental sequence of en echelon folds, but their mode of coupling to strike-slip faults is not really understood, nor is the influence of mechanical layering in their development, and it is not known why folds rather than fractures should form at one place or another. Natural en echelon folds lack interbedded sheets of tin foil, rubber, or plastic needed for their formation in experimental models, but the vast literature on the genesis of buckle and bending folds, wherein such factors as bedding thickness, strain rate, and bedding-plane slip are of fundamental importance, will certainly have relevance to the problem of generating en echelon folds in strike-slip fault zones.

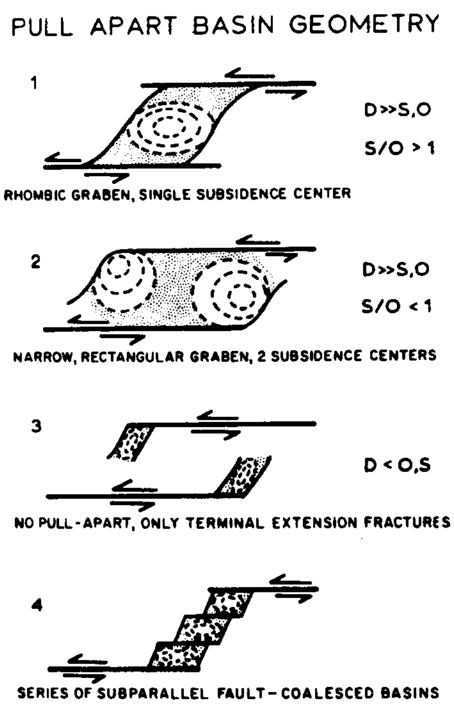


Figure 20. Geometry and locations of basins in overstepping domains of parallel faults (with modifications after Deng and others, 1986). D = depth to basement; S = spacing between parallel or en echelon fault strands; O = overstep. Reproduced with permission of Birkhaüser Verlag AG, Basel, Switzerland.

CONVERGENT AND DIVERGENT STRIKE-SLIP FAULT TECTONICS

The geometry and style of structures associated with strike-slip faulting depend greatly on several factors at different times and different places along and within a strike-slip fault zone, including the nature of the rocks being deformed, the configuration of pre-existing structures, the amount of horizontal slip, the contribution of the vertical component of slip, and the strain rate, which is a major consideration for clay models wherein the strain rate is rapid.

The most important factor governing uplift or subsidence along a strike-slip fault is the bending geometry of the fault surface relative to its slip vector (Fig. 18), because that determines whether local convergence or divergence will occur (Fairbanks, 1907, p. 322; Pakiser, 1960; Clayton, 1966; Crowell, 1974b). Where strike-slip movement is inhibited by restraining bends (Crowell, 1974b), convergent strike slip or transpression (Harland, 1971) occurs associated with crowding, crustal shortening, and uplift. Releasing bends (Crowell, 1974b) provide for transtension (Harland, 1971) or divergent strike slip accompanied by stretching, crustal extension, subsidence, and formation of pull-apart basins.[2]

An appreciation of crustal mobility and flexibility, both vertical and horizontal, is of particular importance for arriving at tectonic understanding of deformation in strike-slip fault zones (Crowell and Sylvester, 1979). Even while crustal blocks move laterally over time, they may alternately rise and sink (Kingma, 1958; Clayton, 1966). In experimental studies, elongate blocks in the zone of strike-slip faulting are bounded by various arrangements of R and P fractures (Woodcock and Fisher, 1986; Bartlett and others, 1981; Tchalenko, 1970; Wilcox and others, 1973; Naylor and others, 1986). Similarly, detailed mapping of strike-slip fault zones (Grose, 1959; Bridwell, 1975; Moore, 1979) shows that strike-slip faults, which are typically belts of braided, nearly vertical shears, bound elongate blocks which are squeezed upward along those shears to make high-standing source areas for sediments (Crowell, 1974b). Where strike-slip movement is divergent, crustal blocks may sag, subside, or tilt between or adjacent to bounding faults, making local sites for deposition of sediments whose stratigraphic characteristics reveal much about the related tectonic activity (Ballance and Reading, 1980). Crowell and Sylvester (1979) drew an analogy between this structural process and a porpoise swimming parallel to the fault strike, alternately arching above the sea surface and diving below it.

Bailey Willis (1938b, p. 664) was well aware of the uplift and subsidence of elongate ridges of rocks, bounded by steep, inward-dipping minor faults in strike-slip fault zones: "Furthermore, it is a common mechanical result of continued pressure in a zone traversed by vertical shears that the crushed rock in the zone is both squeezed up and down, with the result that the stresses are carried back into the adjacent masses. Thrust faults are thus developed in the latter, and they curve upward in the direction of least resistance. They thus become up-curving thrusts or ramps," and (Willis, 1938a, p. 795) "blocks within such an area (fault zone) are completely isolated by minor faults. They are pushed about, mayhap up or down, mayhap over or under, perhaps lengthwise along strike." Willis termed this process "wedge-block faulting," and he regarded the uplifts as wedges both in horizontal and vertical views similar to the notion that Wallace (1949) had for the origin of "center-trough ridges" or pressure ridges, as they are now called.

Basins Related to Strike-Slip Faults

Form and Shape. Becker (1934) and Lotze (1936) were among the first to realize that strike-slip faulting may generate large and complex basins (Şengör and others, 1985). Between curved or releasing overstepped fault segments, depressions develop as sharp, rhomb-shaped basins (Crowell, 1974b; Garfunkel, 1981) or as lazy S- or Z-shaped basins (Schubert, 1980; Mann and others, 1983) due to local crustal extension. The basins range in size from small sag ponds along a strike-slip fault to rhomb-shaped basins up to 500 km long and 100 km wide, such as the Cayman trench located between the releasing overstepped Oriente and Swan Islands transform faults along the north edge of the Caribbean plate (Mann and others, 1983; Rosenkrantz and others, 1988).

Considerable literature concerns the geometry of basins related to strike slip (Dibblee, 1977a; Aydin and Nur, 1982; Mann and others, 1983; Harding and others, 1985), their three-dimensional structure (Ben-Avraham and others, 1979; Ben-Avraham, 1983; Howell and others, 1980; Ginzburg and Kashai, 1981; Fuis and others, 1982; Bally, 1983; Schubert, 1982b), sedimentation (Crowell, 1974a; Ballance and

[2] This kind of basin has also been called a tectonic depression (Clayton, 1966), strike-slip graben (Belt, 1968), rhomb graben (Freund, 1971), negative flower structure (Harding, 1983a), tulip structure (Naylor and others, 1986), or rhombochasm (Carey, 1958) if it has a volcanic floor like the Cayman trough (Rosenkrantz and others, 1988) and the Black and southern Caspian Seas (Apol'skiy, 1974).

Reading, 1980; Hempton 1983; Hempton and Dunne, 1984), origin (Crowell, 1974b, 1981, 1987; Christie-Blick and Biddle, 1985), evolution (Bahat, 1983; Şengör and others, 1985), tectonic setting (Crowell, 1974b, 1981), and thermal history (Royden, 1985; Karner and Dewey, 1986). The close attention focused on strike-slip basins reveals that they are much more varied and complex than originally envisioned in the early, frequently cited paper about the Death Valley "pull-apart" by Burchfiel and Stewart (1966).

Crowell (1974a) depicted pull-apart basins as deep, rhomb-shaped depressions bounded on their sides by two, subparallel, overlapping strike-slip faults, and at their ends by perpendicular or diagonal dip-slip faults, termed "transfer faults," which link the ends of the strike-slip faults. Early writers inferred that the bounding strike-slip faults merge at depth to a single master fault (Kingma, 1958; Clayton, 1966; Sharp and Clark, 1972), and whereas some do merge at depth (D'Onfro and Glagola, 1983), others just as clearly must be vertical at depth and bound a down-dropped block in between; for example, the Dead Sea basin (Manspeizer, 1985; Eyal and others, 1986) and the Caricao basin of Venezuela (Fig. 19).

Typical pull-apart basins have an aspect ratio of 3:1 in plan view (Aydin and Nur, 1982), although that value may vary widely, depending on whether the structural, physiographic, or active dimensions of the basin are measured. Crowell (1974b) implied, and Quennell (1959) and Freund and Garfunkel (1976) suggested, that the dip-slip faults at each end of a rhombic pull-apart were a single fault prior to extension between the parallel segments of the main strike-slip fault. In such a model, the length of the graben would therefore reflect the amount of horizontal displacement (Eyal and others, 1986), and the graben would be characterized by great depth relative to its areal dimensions. This may be true for some small or young basins such as those between closely spaced strike-slip faults; for example, Mesquite basin in Imperial Valley, southern California, is only 5 km wide and 5 km long, but it is filled by at least 5 km of sedimentary rocks (Fuis and Kohler, 1984). Other pull-apart basins are much more complex than the simple pull-apart concept implies; a range of possibilities is depicted by Harding and others (1985).

High-resolution seismic data from the Gulf of Elat illustrate some of the main structural features of pull-apart basins (Ben-Avraham and others, 1979; Ben-Avraham, 1985). There major rift faults having a component of left slip are arranged en echelon, and between them are three sharp, north-trending, pull-apart basins, separated by low sills, wherein the structural relief may exceed 5 km. The sedimentary fill, composed of turbidites and pelagic sediments, is more than 7 km in the deepest of the three basins. The basin floors tilt eastward in the form of a half-graben so that the stratal thicknesses are asymmetric, being thickest on the east sides of the basins. Locally the basins are evidently being shortened in an east-west direction, because the strata are arched upward in a series of large anticlines; however, the seismic lines were not spaced sufficiently close to determine the continuity of the folds from one profile to another. Ben Avraham and his colleagues concluded that the folds trend north-south, parallel to the main basin-bounding faults, but it is also possible that they have an en echelon arrangement, related to syntectonic strike slip associated with the opening of the basins.

The structural geometry of the sedimentary cover in the overstep between strike-slip fault segments depends on the length of overlap, the width of the gap between the fault segments, and the depth to the main fault in the basement (Fig. 20). Recent analyses of the patterns of basin formation and faulting related to overstepped strike-slip faults were published by Mann and others (1983), Rodgers (1980), Segall and Pollard (1980), and Hempton and Dunne (1984). They analyzed a discontinuous fault composed of interacting segments. Rodgers based his analysis on infinitesimal strain theory and assumed constant slip along the entire length of the fault. Segall and Pollard, however, maximized the displacement near the middle of the fault segments and allowed it to go to zero at the ends of the faults, comparable to what is observed in nature. When taken together, both analyses provide clues to the displacement geometry of pull-apart basins, the orientation and kinds of faults which may be found in these basins, as well as the state of stress in and around those basins.

The extended domain within a releasing bend has been depicted as having a meshlike arrangement of extension fractures and strike-slip faults between segments of the bounding strike-slip faults (Sibson, 1986, 1987), an arrangement which was confirmed for the North China Basin by analysis of the complex 1976 Tangshan earthquake sequence (Nábělek and others, 1987). The basin is a large, hydrocarbon-producing basin that began to form in early to middle Eocene time. It is located, and evidently evolved, between two, right-stepping, master, right-slip faults of a larger dextral-slip fault system, and the dominant focal mechanism of the earthquake sequence was right slip. Lesser right-slip, normal, and thrust faults are present within the basin in a geometrical arrangement that mimics the pattern of master faults, and they produced lesser shocks in the earthquake sequence that revealed the structural nature and tectonic mechanism of the basin. The interplay among the intrabasin faults outlines domains of uplift and subsidence within the basin. Secondary pull-aparts in the basin form unconnected domains of local subsidence which, taken in combination, impart an intense, wholesale subsidence to the entire basin. Unfortunately, the earthquake data are insufficient to determine how the basin-bounding faults project to depth or to determine the nature of the basin floor (Nábělek and others, 1987).

With regard to the nature of their floors, pull-apart basins seem to lie between two end members.

At one end are true rifts that extend at depth into hot rocks of the upper mantle, such as those expected above an oceanic spreading ridge. Older rocks, largely continental, are ripped asunder, first by attenuation of the upper crust, and then by actually breaking apart as the mantle material wells up into the widening gash. Under these circumstances, older parts of the basin floor on which sediments are laid down may be missing completely. These basins lack a true basement, and a well drilled to depth would go through sediments into a sill and dike complex of volcanic rocks. Because they intrude the oldest sediments in the basin, the volcanic rocks are younger than sediments at the base of this type of pull-apart. If the well were drilled vastly deeper, it would presumably reach hot and even molten rocks of the lower crust and upper mantle. The Salton basin is an example of this kind of basin.

At the other end of the spectrum of pull-apart basin types are those that bottom-out on a detachment or decollement surface within the upper crust, and they may be grouped into two subtypes: 1) those that bottom against flat tectonic surfaces, or the detachments or flat faults themselves, and 2) those that bottom unconformably against older basement where the detachment is deeper still. Seismic profiling by COCORP and CALCRUST and other geophysical studies, along with down-dip extrapolation of surface observations, show that much of southern California is underlain by subhorizontal tectonic surfaces (Cheadle et al., 1986; Frost and Okaya, unpub. ms.). It is quite likely that some of the through-going reflections are from structural discontinuities on which crustal blocks pull apart and rotate. The profiles disclose several suspect detachments, however, so it is not yet clear on which, if any, the rotations and pull-aparts occur. Perhaps the reflections are stacked decollements, and intermediate blocks between them rotate differently from their underlying and overlying neighbors (J. C. Crowell, unpub. data).

Examples of basins that terminate in a detachment at a relatively shallow level include the Vienna basin (Royden, 1985) and the Dead Sea basin (Manspeizer, 1985).

SEDIMENTATION RELATED TO STRIKE-SLIP FAULTS

Much literature has been written about sedimentation in basins along strike-slip faults, the most recent and comprehensive of which has been edited by Ballance and Reading (1980), Crowell and Link (1982), and Biddle and Christie-Blick (1985). Many basins are typified by high sedimentation rates, scarce igneous and metamorphic activity, abrupt facies changes, abrupt thickening of sedimentary sequences over short distances, numerous unconformities which reflect syntectonic sedimentation, and the presence of a locally derived, skewed fan-body of fault-margin breccia facies representing talus detritus or alluvial fans (Crowell, 1974a, 1974b; Mitchell and Reading, 1978; Hempton and others, 1983; Dunne and Hempton, 1984; Nilsen and McLaughlin, 1985). The coarse, basin-margin facies forms a narrow band along the fault at the edge of the basin. It is volumetrically subordinate to, and contrasts strongly with, the main sequence of flood basin and lacustrine strata with which it interfingers and mixes in the basin, and which is much finer grained, farther traveled, and commonly deposited by turbidity currents (Hempton and others, 1983; Sadler and Demirer, 1986).

The most distinctive stratigraphic feature of basins that form in association with strike slip is the extreme thickness of onlapping sedimentary sequences in pull-apart basins relative to their area (Fig. 21). This happens because of migration of the depocenter by means of syndepositional strike slip (Crowell, 1974b, 1982a). The center of deposition migrates in the direction opposite to that of strike-slip movement of the basin, so that the basin lengthens over time, and the sediments are deposited in an overlapping "venetian blind" arrangement or "stratal shingling" which youngs toward the depocenter (Crowell, 1982a, 1982b; Hempton and Dunne, 1984). The areal extent of Hornelen Basin of western Norway is less than 1,250 km^2, but the stratigraphic thickness of the Devonian sedimentary sequence therein approaches 25 km in a basin 60–70 km long, 15–25 km wide (Steel and others, 1977; Steel and Gloppen, 1980). The true maximum vertical thickness of the succession at any point, however, is probably less than 8 km (Steel and Gloppen, 1980). Ridge basin in southern California is 30–40 km long, 6–15 km wide, about 400 km^2 in areal extent, with a cumulative fill of about 13 km (Crowell and Link, 1982), but the thickness of strata in any single drill hole would be considerably less (Fig. 21). Such thick, asymmetric, sedimentary fillings are characteristic of other basins of various sizes, ages, and tectonic settings (Aspler and Donaldson, 1985; Guiraud and Seguret, 1985; Manspeizer, 1985; Nilsen and McLaughlin, 1985).

Uplifts Related to Strike-Slip Faults

Convergent strike slip or transpression (Harland, 1971; Sanderson and Marchini, 1984) provides a component of horizontal shortening across the strike-slip fault zone which is necessarily accompanied by compensatory uplift of rocks in the fault zone. This is clearly demonstrated in laboratory-model studies where an elongate, fault-bounded welt forms above the zone of principal displacement (Fig. 22) because of the accommodation of the component of shortening strain by uplift (Lowell, 1972; Wil-

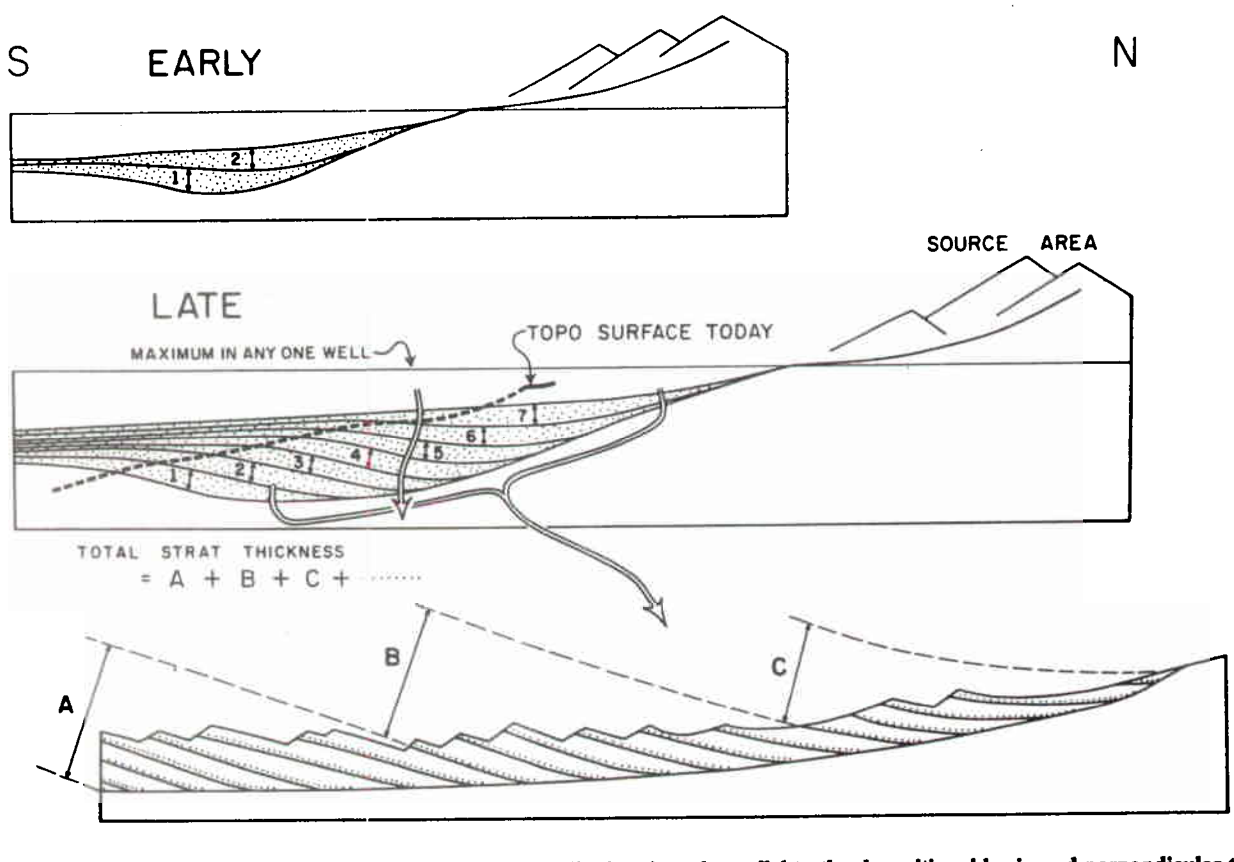
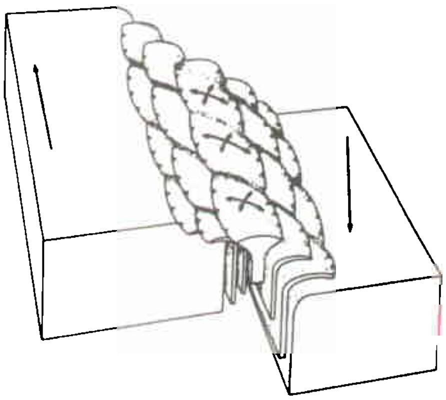
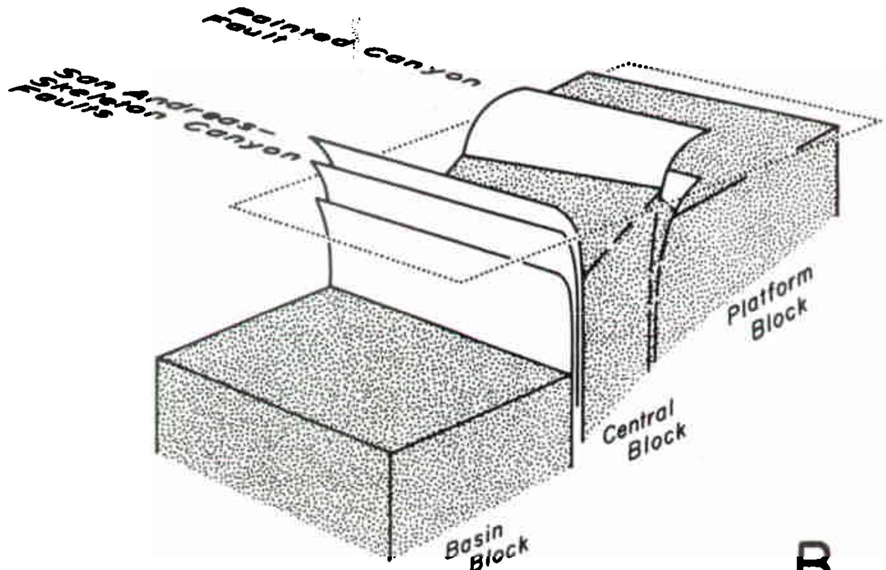


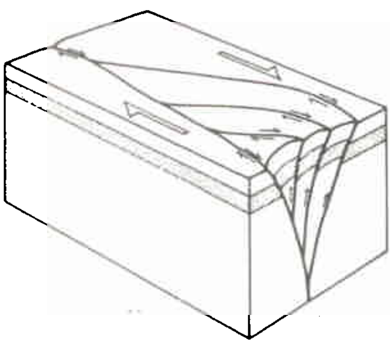
Figure 21. Diagrammatic arrangement of stratal units in Ridge Basin, viewed parallel to the depositional basin and perpendicular to a right slip fault which lies between the source area and the basin (from Crowell, 1982a). The center of deposition migrates toward the principal source area which lies north of the basin and across the fault behind the basin. The upper and center cross sections show the arrangement at early and late states, and the lower cross section shows the way the region is exposed today after uplift and erosion. The total stratigraphic thickness is obtained by adding thicknesses measured along the topographic surface today, and is reproducible by any stratigrapher. Reproduced with permission of Society of Economic Paleontologists and Mineralogists.



A


Painted Canyon Fault
San Andreas-Skeleton Canyon Faults
Platform Block
Central Block
Basin Block

B



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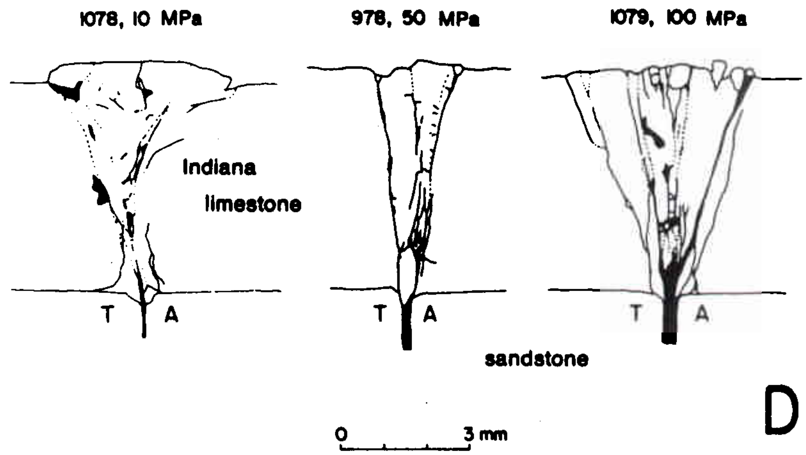
1078, 10 MPa
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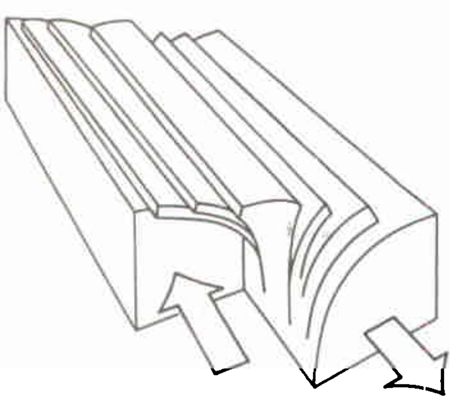

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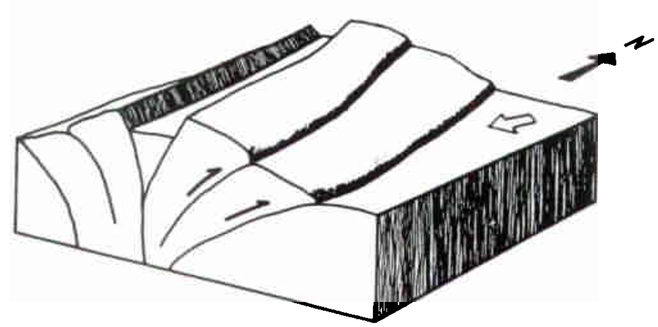

N

F

Figure 22. Conceptual diagrams of palm tree structures in right simple shear. (A) After Lowell (1972, p. 3099), reproduced with permission of Geological Society of America and of Lowell, 1988; (B) after Sylvester and Smith (1976), reproduced with permission of American Association of Petroleum Geologists and of Sylvester, 1988; (C) after Woodcock and Fisher (1986), reproduced with permission of Journal of Structural Geology; (D) after Bartlett and others (1981), reproduced with permission of Elsevier Science Publishers and of Bartlett, 1988; (E) adapted with modifications from Ramsay and Huber (1987, p. 529); (F) with axial graben after Steel and others (1985), reproduced with permission of Society of Economic Paleontologists and Mineralogists.

cox and others, 1973; Bartlett and others, 1981). The welt is bounded by sinuous faults which, in cross section, are nearly vertical at depth and flatten upward, carrying parts of the welt short distances outward upon the adjacent, stable blocks. This upward-branching arrangement of faults in strike-slip fault zones is mentioned by Willis (1938a), Kingma (1958), and Clayton (1966); was mapped by several writers (Wallace, 1949; Sylvester and Smith, 1976; Burke, 1979; Davis and Duebendorfer, 1982); and was produced especially clearly in experimental studies in layered media (Emmons, 1968; Bartlett and others, 1981; Naylor and others, 1986).

Lowell's (1972) conceptual block diagram of transpression (Fig. 22A) shows a central zone of vertical slabs which rise upward and outward on convex-up faults over the adjacent blocks like a stack of imbricate thrusts. Wilcox and others (1973) termed the upward and outward branching arrangement of faults in Figure 22A "positive flower structure."[3] Because the surficial traces of the faults are sinuous in plan view, most geologists would interpret field outcrop and map patterns of the faults as evincing thrusts rather than strike-slip faults, because, almost by definition, strike-slip faults are "transcurrent"; that is, they cut straight across rocks, structures, and topography as vertical, throughgoing faults.

[3]Descriptions of the two-dimensional arrangement of the faults in interpretations of seismic sections and structural cross sections have a variety of botanical appellations, including "positive flower structure" (Wilcox and others, 1973) or "palm tree structure" (Sylvester and Smith, 1976), as well as "pop-up," "squeeze-up," and "tectonic wedge" in convergent strike slip, and "negative flower structure" (Harding and Lowell, 1979) or "tulip structure" (Naylor and others, 1986) in divergent strike slip. Because real flowers have a variety of shapes as well as appellations, I prefer the term "palm tree structure" to describe the convex-upward geometry of faults in profile that bounds an uplifted block in a strike-slip fault zone, even though flower structure has a few years' precedence (Biddle and Christie-Blick, 1985). I believe the term "tulip structure" evokes a clear image of the concave-upward geometry of faults in profile that form in divergent strike-slip, and I propose that "tulip structure" be used in place of "negative flower structure," even though these structures are not circular in plan view.

The near-surface, low-angle segments of strike-slip faults are common along many major convergent strike-slip faults, particularly where the faults mark the base of a steep mountain front. In the Mecca Hills (Fig. 22B), the San Andreas, Skeleton Canyon, and Painted Canyon faults dip 60°–70° toward the central block in the deepest exposures. The faults flatten upward into short, oblique-slip thrust faults beneath rocks of the central block that have been thrust from 50 to 200 m upon the footwall of the adjacent block.

The upward flattening of strike-slip faults in zones of transpression has been documented also on the Alpine fault of New Zealand (Wellman, 1955), and along the Banning and San Jacinto faults of southern California (Allen, 1957; Sharp, 1967). Similar, more extensive nappes traveled up to 1 km in the "big bend" segment of the San Andreas fault (Davis and Duebendorfer, 1982), and as much as 5 km in West Spitsbergen (Lowell, 1972; Kellogg, 1975; Craddock and others, 1985).

Wellman (1955) considered that the surficial thrusting results from downslope creep under gravity which bends the fault in mass movement, but according to Allen (1965, p. 84): "Another important causal factor here and elsewhere may be related to the origin of the steep mountain front itself: if, as seems likely, the presence of the mountain front is caused by a local vertical component of displacement along the predominantly transcurrent fault, then vertical motion constrained at depth to a vertical plane must necessarily result in localized low-angle thrusting at the surface, as has been demonstrated analytically and in models by Sanford (1959)" and by Hafner (1951). Allen pointed out that these surficial thrusts may conceal the major underlying strike-slip faults and may have delayed recognition of the dominance of horizontal displacements on faults in many parts of the world.

Allen's hypothesis is a good explanation for the local surficial flattening of strike-slip faults, but I believe that the larger-scale thrusting results from a mechanical delamination of an uplifted block coupled with shortening that permits thin structural flakes to move obliquely across the adjacent block and appear as though they have been thrust up and out of the deformation zone. This must be the explanation for those thrust segments which seemingly have come out of a strike-slip fault zone which is very much narrower than the thrust segment itself.

Sylvester and Smith (1976) were influenced by Lowell's block diagram to the extent that they also visualized an uplifted central block nearly 2 km wide underlain by a zone of crushed metamorphic rocks between the two, subparallel faults (Fig. 22B). Analyses of fractured and faulted gneiss layering show that the crystalline basement responded to convergent strike slip by cataclastic flow and piecemeal slip along fractures and faults at all scales, just as Willis (1938b) realized for strike-slip faults elsewhere. The overlying sedimentary sequence between the faults, however, deformed passively in response to cataclastic flow and differential uplift of the basement (Sylvester and Smith, 1976). Outside of the main fault zone, the Painted Canyon and lesser strike-slip faults branch away from the San Andreas fault to form an array of splay faults (extensile fan in the terminology of Woodcock and Fischer, 1986; Fig. 22C) in plan view (Fig. 23), just as they do in experimental studies (Naylor and others, 1986). The lesser faults are characterized by normal separation where they strike at high angles to the San Andreas fault (Harding and others, 1985; Fig. 23).

The well-exposed structure of the footwall at the edges of the relatively down-dropped blocks and beneath the thrust in the Mecca Hills (Fig. 24) shows that the uplift necessitated by the convergent strike slip is accommodated by an accordion style of folding and by a variety of faults, some of which project into the main fault that bounds the uplifted central block, and others which dip away from the central block as back-thrusts and die out in bedding surfaces (Sylvester and Smith, 1976). Together, all of the faults bound a triangular-shaped domain of folded strata which resembles the triangle or delta zone (Jones, 1982; Butler, 1982; Lowell, 1985, p. 282) seen at the toe of some overthrust faults, including the Prospect thrust near Jackson, Wyoming (Dorr and others, 1977; Dixon, 1982), the Findley structure in the northern

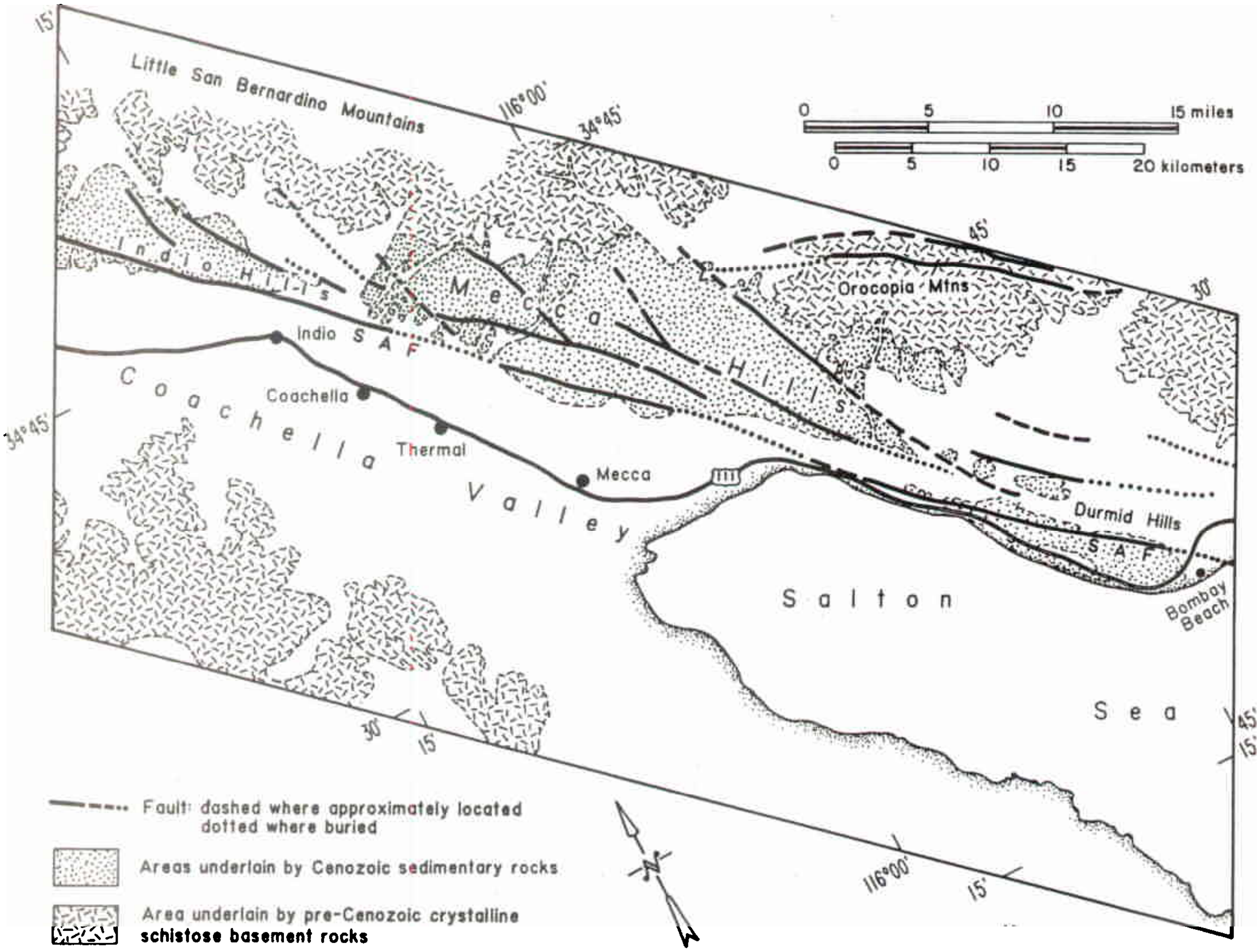


Figure 23. Simplified geologic map of fault splays and of tectonic culminations along the fault zone, represented by the Indio Hills, Mecca Hills, and Durmid Hills at the southern end of the San Andreas fault (SAF) in Salton Trough.

foothills of Alberta, Canada (Jones, 1982), that beneath the Nushman anticline in the western foothills belt of south-central Taiwan (Suppe, 1980), and the toe of the Alpine front and Molasse Basin, Bavaria (Bachman and Koch, 1983). The axes of the folds in the triangle zone along strike-slip faults are oriented at low angles to the main faults and constitute a transected series of en echelon folds. Similar, larger triangle zones along convergent strike-slip faults are alleged to have been documented in the subsurface of producing offshore oil fields of southern California and in the Taranaki basin of New Zealand.

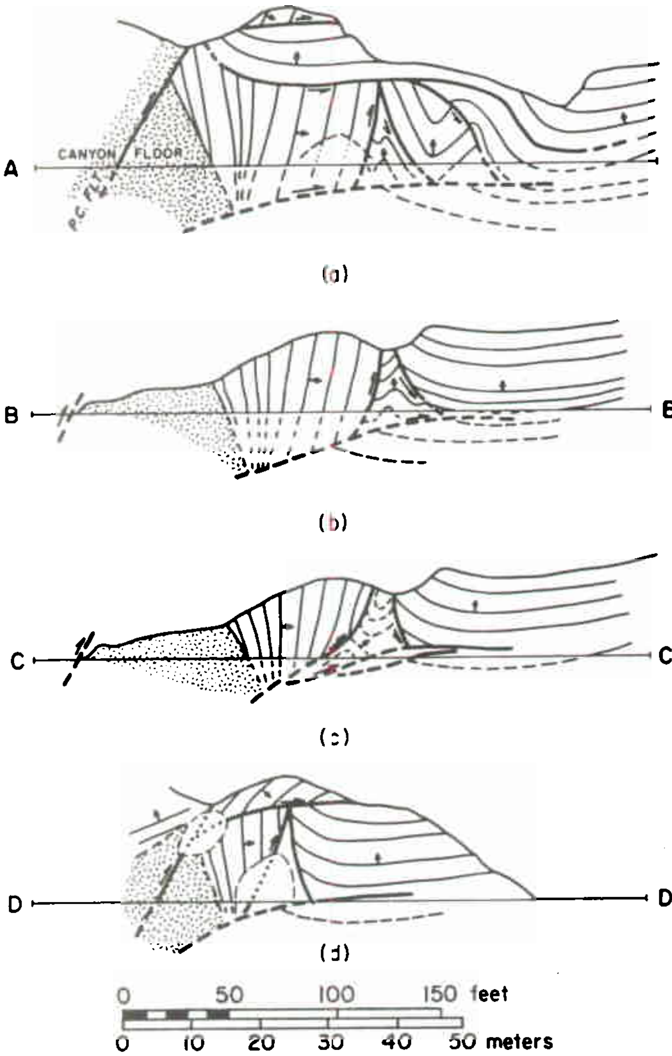
Seismic profiles have been obtained over palm tree structures (for example, Fig. 25) and tulip structures (Harding and Lowell, 1979; Bally, 1983; Harding, 1985) and are typified by having reverse- and dip-separation faults side by side across the crest of the structure (Figs. 22D, 22F). To be sure, some of the faults are indeed thrusts that reflect the over-all uplift of the central block and shortening across it, but others are just as clearly normal faults reflecting the extension across the top of the uplifted and laterally spreading block as documented by Sylvester and Smith (1976) and shown diagrammatically in Figure 22F. To a greater degree, however, the variable apparent vertical displacement is due to oblique slip or pure strike slip in and out of the plane of the vertical section.

SEISMOTECTONICS

Segmentation

Different segments of active strike-slip faults behave differently in terms of (1) the maximum magnitude of earthquakes they have generated or are capable of generating; (2) the maximum amount of surface displacement they display with those earthquakes; (3) the return frequency of earthquakes; and (4) the rate of aseismic fault creep (Allen, 1968; Wallace, 1970). For example, the strike of the San Andreas fault is within 5° of the plate slip vector, with three main exceptions where the fault trace departs from this trend: one north of San Francisco and located largely offshore (Thatcher and Lisowski, 1987), the second on the north side of the San Gabriel Mountains (Fig. 9), and the third along the south side of the San Bernardino Mountains (Fig. 9). Each is a restraining bend and a center of great earthquakes where the fault is considered to be locked (Allen, 1968) relative to the straight, creeping segment of the fault in central California which is typified by rather frequent minor to major earthquakes (Brown and Wallace, 1968; Wallace, 1970).

Fractal analyses reveal that the San Andreas fault has slight, but statistically significant, varia-



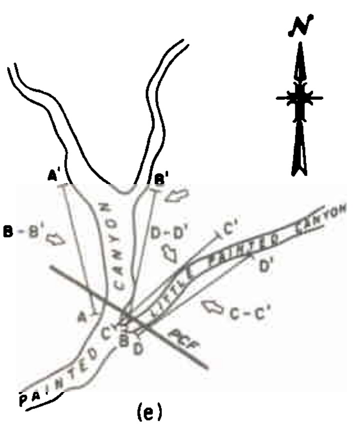


Figure 24. Field sketches of triangle zone exposed on canyon walls beneath Painted Canyon fault (P.C. fault; PCF), Mecca Hills, southern California. (a) View west-southwest; (b) view east-southeast (reversed); (c) view north-west; (d) view southeast (reversed); (e) locations of cross sections relative to each other.

tions in fractal dimensions from segment to segment, variations that correlate with Allen's (1968) subdivisions (Okubo and Aki, 1987; Aviles and Scholz, 1987). Although fault jaggedness increases southeastward along the strike of the fault, fault segments having variations of creep, seismic slip, or microearthquake activity cannot be distinguished from one another on the basis of their fractal dimension (Aviles and Scholz, 1987).

The jagged surface trace of the San Andreas fault may be subdivided into 12- to 13-km segments which have abrupt changes in trend of 6° ± 2° (Clark, 1984). Tectonic depressions are present where the segments are parallel to the plate slip vector (N40°W), and uplift occurs where the fault segments are oblique to the plate slip vector (Bilham and Williams, 1985; Fig. 26). Across the San Andreas opposite the uplifts are some of the deepest parts of the Salton Trough; judging from gravity data (Biehler and others, 1964), the depth to basement in the basin block is 4,000 m only a few kilometers from the southwest edge of the Mecca Hills (Fig. 22B).

Many authors have recently addressed the short-term implications of segmentation insofar as earthquake mechanics and hazards are concerned, but the long-term geological implications have yet to be elucidated. Presently defined segment lengths are determined by the depth to the seismogenic zone, by the presence and position of releasing or restraining bends, by fault bends, and by "asperities" (Tang and others, 1984; Barka and Kadinsky-Cade, 1988). Within major zones of strike slip, geologists have recognized and mapped numerous shears which predate the presently active trace, and which may be reactivated in future earthquakes. How and why should slip transfer from segment to segment across strike, and how does this slip-swapping affect the length of the fault segments? When a strike-slip fault is bent beyond some angle that accommodates easy slip, does a new fault take a shorter path, thereby short-cutting the bend and abandoning it? The curved San Gabriel fault (Fig. 9), active in Miocene and Pliocene time, is regarded as an old segment of the San Andreas fault which was abandoned when the San Andreas fault shifted to its younger, straighter locus on the northeast side of the San Gabriel Mountains (Crowell, 1979). How do strike-slip faults become bent in the first place?

The average strike of the San Andreas fault is N40°W north of the Transverse Ranges, it strikes exactly east-west for a distance of 10 km in the "big bend" segment where the Big Pine and Garlock faults intersect the San Andreas fault (Fig. 9), and then it strikes about N60°W for ~300 km into the Salton Trough. Several writers have speculated that the "big bend" is a result of left slip on the Garlock fault as it plays its role as an *intraplate transform* (Fig. 27). In each of the models, Basin and Range crust north of the Garlock fault extends westward, relative to a stationary Mojave Desert, carrying also westward the Sierra Nevada off the Kingston Peak core complex (G. A. Davis, 1988, personal commun.), and the Great Valley, together with the San Andreas fault (Eaton, 1932; Hamilton and Myers, 1966; Davis and Burchfiel, 1973; Wright, 1976; Hill, 1982; Bohannon and Howell, 1982). At the east end of the Garlock fault at the Kingston Range, therefore, the slip is evidently zero; at the center part of the fault, sinis-

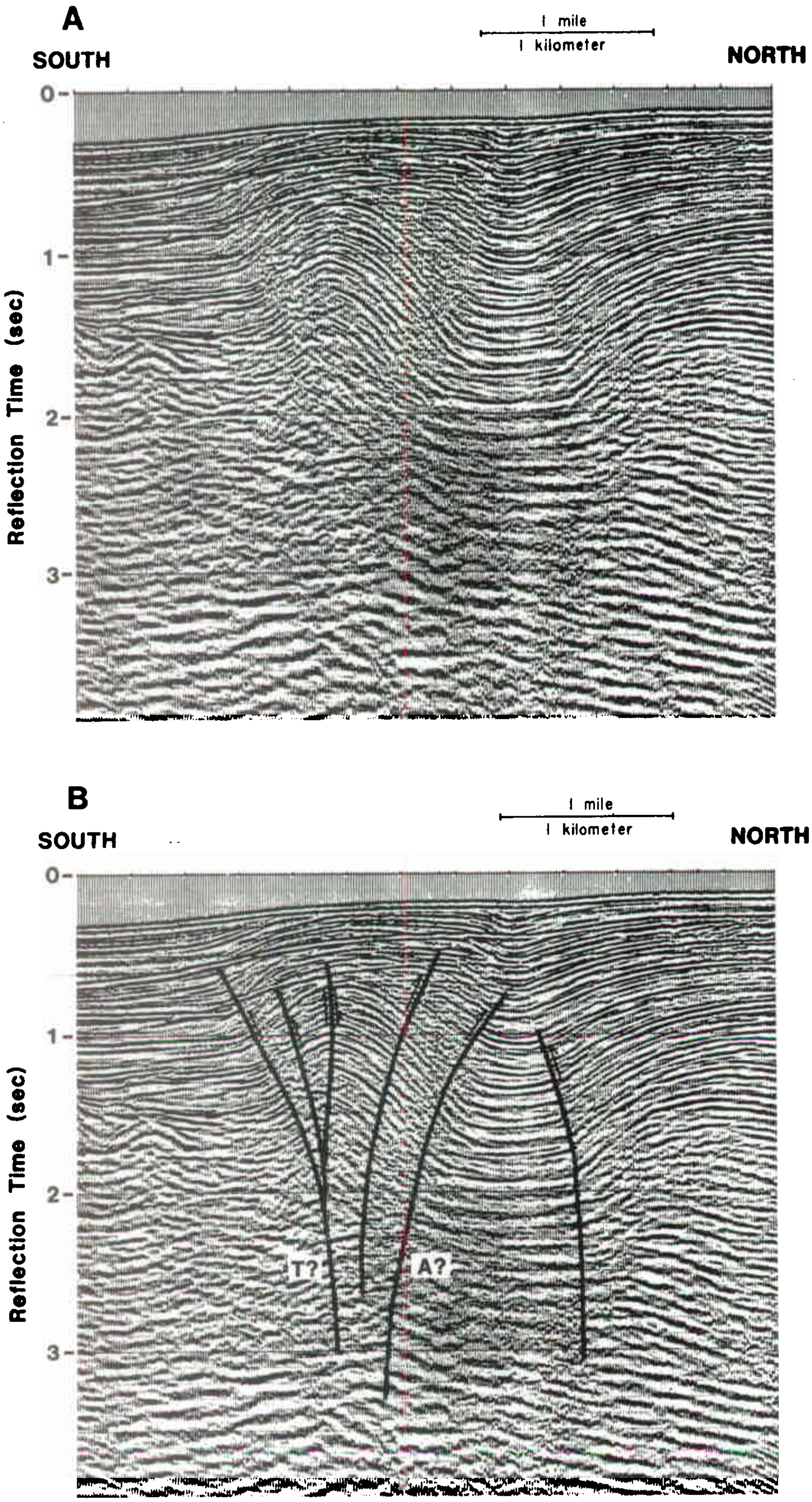


Figure 25. Seismic expression of a palm tree structure along the subsea extension of the Oak Ridge fault in the Santa Barbara Channel (Fig. 9), southern California. (A) Migrated reflection profile; (B) interpreted, migrated reflection profile. A = movement away from observer; T = movement toward observer. The near-surface ends of the faults cannot be projected to the sea floor because of interference of reflectors by sea-floor multiples. The east-west length of the structure is nearly 6.5 km.

tral displacement is about 60 km, and it increases westward to the San Andreas fault (Davis and Burchfiel, 1973). Because of its low angle of intersection, one might guess that sinistral displacement during a large earthquake on the Garlock fault will transfer into the dextral San Andreas fault for a short distance. Then when the San Andreas fault slips right laterally in a subsequent earthquake, the displacement will bypass the end of the Garlock fault and cancel the bit of sinistral slip imparted to it by the earlier event on the Garlock fault (Bohannon and Howell, 1982).

The net effect of the westward transport of blocks north of the Garlock fault is to place a restraining bend in the edge of the North American plate around which the Pacific plate must pass. The difficulty in that passage is reflected by the profound structural complexities in southern California around the "big bend" manifested by the Transverse Ranges, only one of two east-west–trending mountain ranges in North America.

Paleoseismicity

One of the most far-reaching papers concerning frequency of earthquakes and slip rate on any kind of fault was published by Kerry Sieh in 1978. He defined the timing of Holocene movements on the San Andreas fault and, thereby, quantified the earthquake hazard for the major population centers of southern California. By means of microstratigraphic analyses of peat beds interlayered with strata of fluvial sand and gravel in a drained sag pond, Sieh (1978) recognized vertical separation, stratigraphic disruption, and liquefaction related to nine events that deformed the strata similar to the way that the M 8 earthquake of 1857 disturbed the same strata. He showed that nine earthquakes, probably like that of 1857, occurred in the past 2,000 yr, yielding a slip rate of 30 mm/yr and an average recurrence interval of

140 ± 40 yr, but with a range of from 70 to 300 yr. Dendrochronologic studies, however, show that major earthquakes on this part of the San Andreas fault do not cluster closely about the 131-yr average, nurturing "doubts about hypotheses of uniform fault-strain accumulation and relief" (Jacoby and others, 1988, p. 196). Recent ultra-precise ^{14}C ages have revised the average of the intervals to 131 ± 20 yr, and a range from 60 to 400 yr (Sieh and others, unpub. ms.). That is quite frequent, given the fact that similar studies on faults, mainly normal faults, in other tectonic domains of western North America yield return frequencies on many time scales, ranging in general from several thousands of years (Swan and others, 1980; Bucknam and others, 1980; Scott and others, 1985; Malde, 1987; Pearthree and Calvo, 1987; Lubetkin and Clark, 1988) to several hundreds of thousands of years (Bull and Pearthree, 1988; Wallace, 1987), as do major strike-slip faults in other parts of the world (Japan: Okada, 1983; China: Allen and others, 1984; Peru: Schwartz, 1988; north Africa: Meghraoui and others, 1988, Peltzer and others, 1988). Field sites having the favorable combinations of well-preserved, readily datable material, intimately interlayered with a distinctive stratigraphic sequence on the active segment of a fault are rare along the San Andreas fault. Other major strike-slip faults therefore need to be searched and given the kind of systematic paleoseismic attention that the San Andreas fault has received to provide more information about how these faults work.

Earthquakes are temporally clustered in the Pallett Creek segment of the San Andreas fault: the earthquakes in each cluster are separated by decades, but the clusters are separated by dormant periods from two to three centuries (Sieh and others, in press). The occurrence of large historic earthquakes on the North Anatolian fault in Turkey and on predominantly normal faults in the Basin and Range province are also clustered in time (Ambraseys, 1970, 1971; Wallace, 1987). On a longer time scale, Clifton (1968) found sedimentary cycles suggestive of tectonic "events (or closely spaced flurries of events) with a periodicity of tens of thousands of years" possibly related to movements on the San Andreas fault.

King (1987), following Savage's (1971) hypothesis that migratory pulses of earthquake activity along the southern half of the San Andreas fault are driven by creep waves induced by episodic magma injections at the East Pacific Rise, has offered a comprehensible method of loading episodic horizontal strain energy into the fault system. The injections propagate along the broad transform boundary between the Pacific and North American plates at subseismogenic depths.

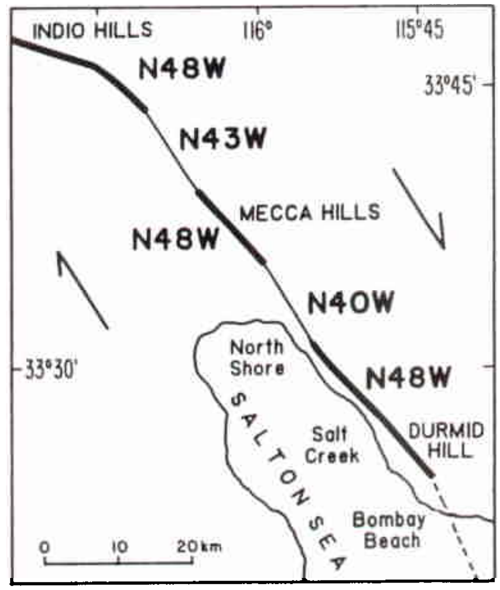


Figure 26. Segments of parallel slip and convergent slip along the southern end of the San Andreas fault (redrawn from Bilham and Williams, 1985). The plate slip vector is parallel to the N40°W segment. Folding and uplift occur on the northeast sides of the N48°W fault segments. Compare with Figure 23.

Recently Sibson (1987) proposed an intriguing idea of wide interest that links strike-slip earthquakes, ore deposits, rock mechanics, segmentation, and fluid flow. "Paleoseismic studies show that segments of some faults tend to rupture at fairly regular intervals in characteristic earthquakes of about the same size" (Sibson, 1985, p. 248; Schwartz and Coppersmith, 1984). The regularity of the ruptures has been questioned recently (Wallace, 1987; Sieh and others, in press), but the characteristic size and rupture dimensions are evidence that segment terminations, which are typically releasing or restraining bends or oversteps, arrest or perturb fault rupture propagation (Sibson, 1985, 1987; Barka and Kadinsky-Cade, 1988). At releasing bends, abrupt extension locally reduces fluid pressure, leading to brecciation by hydraulic im-

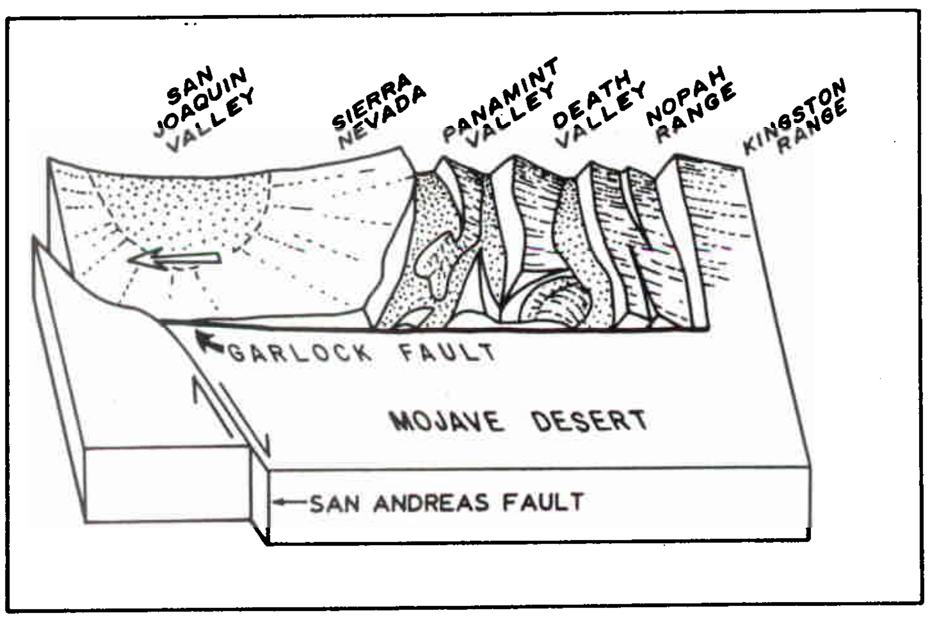


Figure 27. Northward diagrammatic view of the Garlock fault, southern California, as an intracontinental transform between the northern, westward-extended Basin and Range province, and the southern, nondistended Mojave Desert (redrawn with modifications from Davis and Burchfiel, 1973, p. 1417, and reprinted with permission of Geological Society of America and of Davis, 1988).

plosion and to sudden, concentrated influx of fluid into the bend. Arrest of the rupture may be followed by delayed slip transfer through the bend during fluid pressure re-equilibration by diffusion (Sibson, 1986). Where high boiling events are triggered by the arrest mechanism, Sibson (1987) postulated that episodic mineral deposition is induced in the top 1–2 km of the bends by the dynamic effects of rupturing on the flanking strike-slip faults. Other writers have already borrowed these ideas to explain how strike-slip faults controlled the structure of some hydrothermal mineral deposits (Willis and others, 1987), and I suspect this is only the beginning of a renaissance in structural economic geology.

Aseismic Slip

One of the most curious aspects of strike-slip faults not readily demonstrated by other kinds of faults is that of aseismic slip, or creep, which is a surficial phenomenon restricted to the upper 100 m or so of the fault (Goulty and Gilman, 1978; Sharp and others, 1986a). The phenomenon has been documented on only two faults to date: the North Anatolian fault in Turkey (Ambraseys, 1970; Aytun, 1980) and the San Andreas and related faults in central and southern California (Steinbrugge and others, 1960; Brown and Wallace, 1968; Schulz and others, 1982; Louie and others, 1985; Galehouse, 1987). The maximum rate of creep on the central segment of the San Andreas fault is 35 mm/yr (Burford and Harsh, 1980) where the fault zone is narrow and straight and is contained in oceanic basement rocks of the Franciscan Formation (Irwin and Barnes, 1975). That creep rate is equal to the historic slip rate for the entire fault zone as determined from geodetic evidence (Thatcher, 1979), and for Holocene time as inferred from paleoseismic studies (Sieh, 1978, 1984; Sieh and Jahns, 1984). The creep rate is only about 3.4 mm/yr at the southern end of the San Andreas fault (Louie and others, 1985).

Creep is believed to be driven by elastic loading of the crust at seismogenic depths. Its occurrence at the surface represents either deep aseismic motion which is expressed as seismicity in one area and to aseismic slip in another, or the accumulation of seismic dislocations away from slipping faults (Louie and others, 1985).

The tectonic significance of creep is a topic of debate (Wallace, 1970; Sylvester, 1986; Williams and others, 1988): some investigators believe that creep represents steady-state slip which relieves buildup of stress on strike-slip faults so that large earthquakes are precluded in a creeping fault segment (Brown and Wallace, 1968; Prescott and Lisowski, 1983), and that notion seems to have gained support, at least for the San Andreas fault, during the past 15 yr by a massive amount of horizontal strain data (Langbein, 1981). Alternatively, creep is postulated to be the first step in progressive failure leading to a large earthquake (Nason, 1973, 1977), although it is clear that several southern California strike-slip earthquakes were not preceded by surficial pre-seismic creep (Cohn and others, 1982).

Strike-slip faults also exhibit a greater degree of *afterslip* than do other kinds of faults so far as is known. Afterslip is fault slip that occurs on the fault in the days, weeks, or even months following the main earthquake. It may be gradual and continuous or episodic, but its principal character is that the slip rate decreases logarithmically over time (Wallace and Roth, 1967; Sylvester, 1986; Wesson, 1987). The amount of displacement of afterslip following strike-slip earthquakes may equal or exceed the coseismic slip (Smith and Wyss, 1968; Ambraseys, 1970; Burford, 1972; Bucknam and others, 1978), whereas it is very much less in thrust and normal fault earthquakes: For example, Sylvester and Pollard (1975) found that afterslip totaled only 1% of the coseismic slip during the year following a thrust earthquake which produced a maximum of 2 m of coseismic net slip at the surface. Whether afterslip is truly aseismic has not been clearly established, although the afterslip following the 1979 Homestead Valley, California, earthquake (M_L = 5.8) was much greater than the summed moment of the aftershocks, leading Stein and Lisowski (1983) to conclude that the afterslip, which constituted about 10% of the seismic slip, was aseismic. Wesson (1987) modeled afterslip with a "stuck" or locked patch at depth on the fault surface, surrounded by an area that creeps in response to the applied stress.

Still another unique seismotectonic phenomenon exhibited to date only by strike-slip faults is that of *triggered slip*, which is coseismic slip on a fault or faults other than the causative fault outside the epicentral area of the main shock (Sylvester, 1986). The phenomenon has been observed repeatedly in moderate earthquakes in the Salton Trough of southern California where up to 30 mm slip has occurred on faults as far as 40 km from the causative fault and its epicenter (Allen and others, 1972; Fuis, 1982; Sieh, 1982; Williams and others, 1988). The mechanism for the triggered slip is problematic, although geodetic and seismologic evidence in the California earthquakes suggests that parts of the affected faults may be variably prestressed or have different shear strengths. Thus those faults which are near failure will slip small amounts either because of shaking dynamically induced by the main earthquake (Fuis, 1982; Sieh, 1982), or because the regional static strain field is perturbed so that stress is concentrated on other faults which then yield by creep. Allen and others (1972) preferred the mechanism of dynamic strain to explain slip triggered on faults in the Salton Trough by the 1968 Borrego Mountain earthquake (M = 6.8), because they found that the change in the static strain field was an order of magnitude less than the dynamic strain due to ground shaking.

Resolution of the various problems of creep will require more and continued monitoring of slip by geodetic and instrumental methods, careful searches for minor slip on associated faults near the seismogenic faults, and better determinations of components of pre-seismic slip, coseismic slip, and afterslip in historical and young prehistoric offset data, not only for the San Andreas fault, but also on other active strike-slip faults.

TECTONIC ROTATION IN SIMPLE SHEAR

Tectonic rotation of slabs of the Earth's crust about a vertical axis in simple shear was suspected or postulated for parts of the American Pacific coast and the Dead Sea regions by several writers (Hamilton and Myers, 1966; Freund, 1970a, 1970b; Teissere and Beck, 1973; Garfunkel, 1974; Beck, 1976; Jones and others, 1976; Simpson and Cox, 1977; Hamilton, 1978) before the notion took root in southern California and flowered.

Hamilton and Myers (1966) postulated that the Transverse Ranges rotated clockwise about a vertical axis at their east end from a position alongside Peninsular Ranges. Jones and others (1976), noting the 90° difference in structural trends in Mesozoic (?) rocks on Catalina and Santa Cruz Islands, relative to the north-south trend in the southern Sierra Nevada foothills, postulated that the islands had rotated 90° without specifying the sense.

Luyendyk and others (1980, 1985), following these and other earlier indications of rotation (Teissere and Beck, 1973; Beck, 1976; Simpson and Cox, 1977; Greenhaus and Cox, 1979), and building on theoretical considerations of crustal rotation (Freund, 1974), thought that the direction and amount of rotation could be determined paleomagnetically. Accordingly, Luyendyk and his students systematically sampled early Neogene rocks, principally volcanic rocks in the western Transverse Ranges, which should have been rotated clockwise by distributed late Neogene shear within the San Andreas fault system (Kamerling and Luyendyk, 1979, 1985; Terres and Luyendyk, 1985). Their results supported the clockwise rotation inferred by Hamilton and Myers (1966), but not the mechanism. More importantly, they were able to determine dimensions of the domains of rotation (Fig. 28), postulating that several blocks or slabs in south-

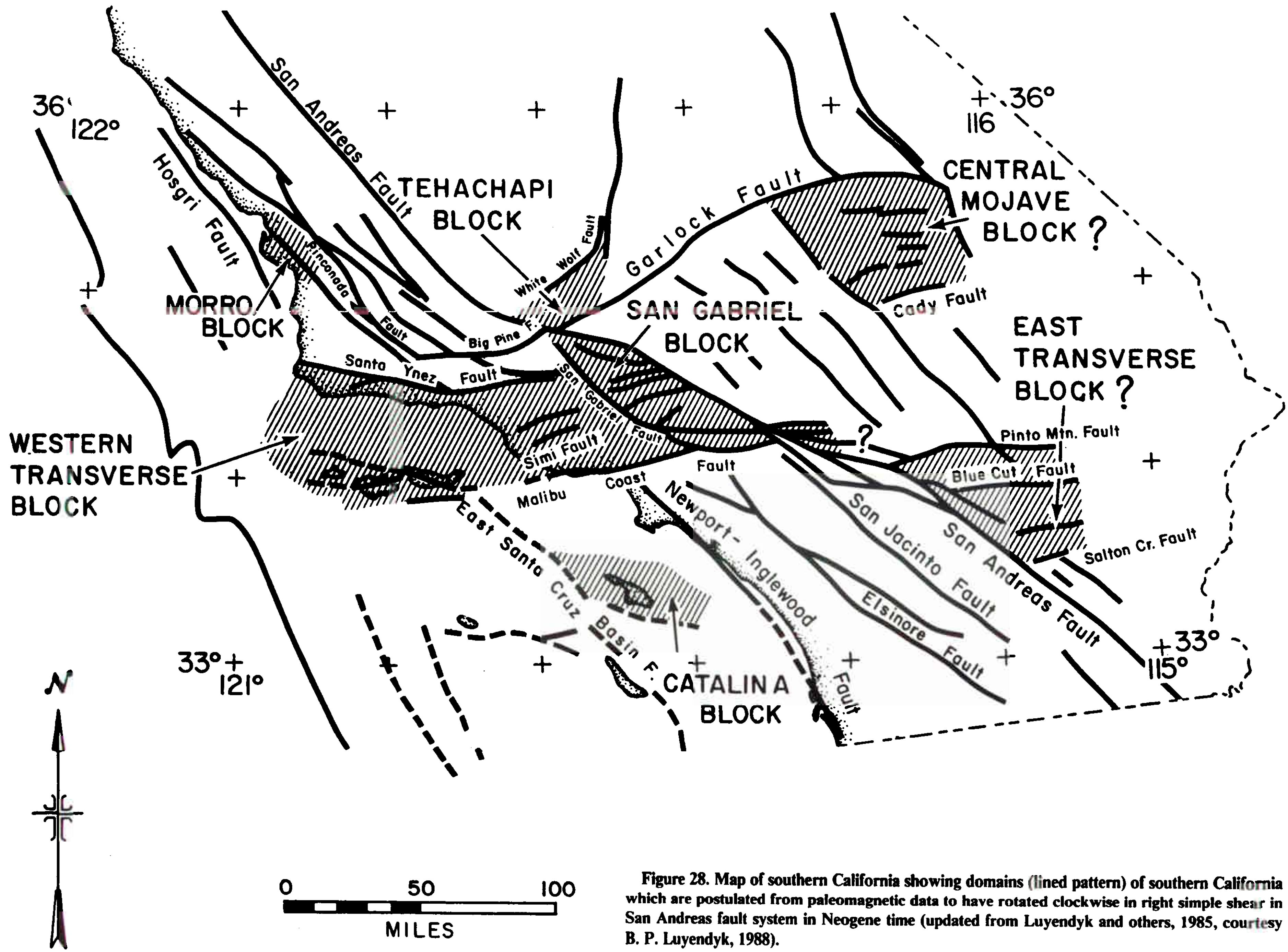


Figure 28. Map of southern California showing domains (lined pattern) of southern California which are postulated from paleomagnetic data to have rotated clockwise in right simple shear in San Andreas fault system in Neogene time (updated from Luyendyk and others, 1985, courtesy B. P. Luyendyk, 1988).

ern California had rotated at least 90° clockwise in a broad zone of simple shear, and perhaps as much as 120° in Neogene time. Initially, they predicted that blocks bounded by east-west-striking, left-slip faults and by northwest-striking, right-slip faults would also be found to have rotated clockwise across a broad zone of simple shear between the two great crustal plates. By and large, subsequent paleomagnetic studies by Luyendyk and his colleagues have substantiated this model. Hornafius and others (1986) extended the hypothesis into the fourth dimension by showing that the magnetic declination vectors having the greatest degree of clockwise rotation are in the older rocks, and the rotations decrease progressively in progressively younger rocks. The western Transverse Ranges (Fig. 28) therefore rotated from 20 Ma to 4 Ma.

Whereas several writers embraced the rotation concepts and found paleomagnetic evidence for rotations in other major strike-slip fault zones (Rotstein, 1984; Ron and Eyal, 1985; Ron and others, 1986; Kissel and others, 1987), other workers have had difficulties understanding how such relatively long, narrow crustal slabs could behave so rigidly during rotation (Nelson and Jones, 1987), and why supporting geologic evidence of the paleomagnetic rotation is not especially evident. This is because existing models of the proposed rotations are geometrically simplified or theoretical (Fig. 29), but few of those models are defined by compelling field data. We lack sufficient information about the three-dimensional geometry of natural strike- and oblique-slip faults, the mechanism of translation and rotation of fault blocks around strike-slip fault bends, or about the sequential development of structures in the complex zone of heterogeneous strain. Geologists have searched for geologic evidence, such as associated deformation, or rotation of paleocurrent directions of Paleogene rocks away from their provenances with variable success (Karner and Dewey, 1986) or inconclusive results (Crowell, 1987, p. 227; Howard, 1987). In southern California, the paleomagnetic evidence seems to point strongly to kinematics of clockwise rotation, but the present models need to be tested by more detailed mapping and structural studies of well-exposed areas.

It seems clear that the rotated blocks must in fact be flakes (Oxburgh, 1972), slabs, or crustal panels (Dickinson, 1983), which detach on a shallow horizontal shear surface as Brown (1928) observed in model studies that rotation of upper layers occurred where horizontal shear took place in a weak, underlying layer. In a 1-m-wide zone of right slip associated with the 1979 Imperial earthquake, Terres and Sylvester (1981) found that elongate blocks of soil, which were defined and bounded by pre-earthquake plow cuts, rotated clockwise from 20° to 40°, forming the deltoid "basins" and slipping internally on the left-slip faults predicted by kinematic models (Freund, 1974; Luyendyk and others, 1980). The rotated blocks had detached at a mechanical anisotropy 15 cm below the surface: the dry-soil–wetted-soil interface.

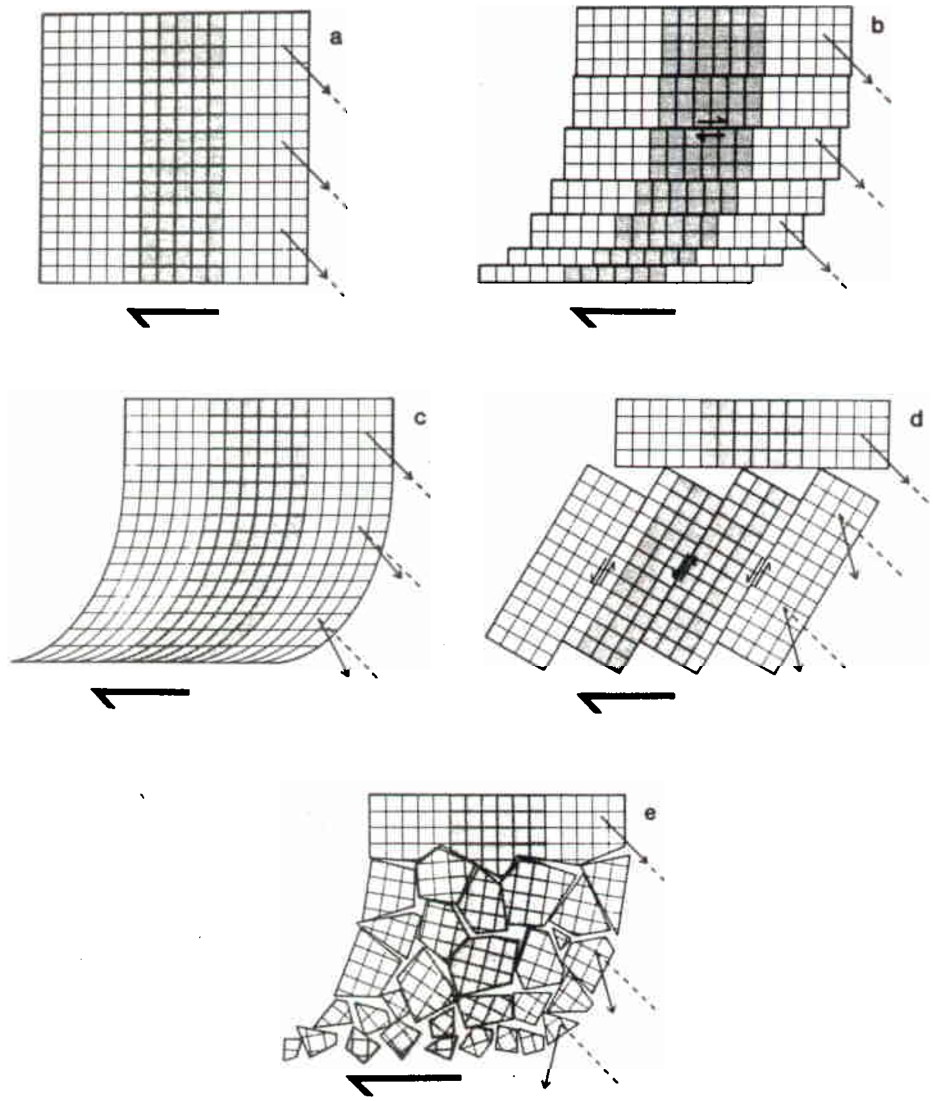


Figure 29. Mechanisms of oroflexural bending and rotation in simple shear zone (after Nelson and Jones, 1987). (a) Undeformed domain; (b) shearing on faults parallel to the main shear zone without rotation; (c) pervasive, continuous bulk simple shear; (d) block rotation with internal antithetic shear; (e) small block model with variable, internal rotation.

On a larger scale, Wilson (1960) described how slabs of ice rotated in a simple shear zone between the Filchner Ice Shelf and the Antarctic mainland. He saw that the slabs were "pinned" on either side of the chasm and rotated rigidly in response to the oceanic current which moves the ice shelf. The ice slabs were bounded by extension fractures or R shears which, during prolonged shear across the ice chasm, twisted and extended in response to rigid body rotation of the slabs (Fig. 30).

Evidence that a mechanical discontinuity or detachment (or several) must underlie the Transverse Ranges in order for them to rotate comes from analyses of P-wave delays from quarry blasts and earthquakes in southern California (Hadley and Kanamori, 1977; Nicholson and others, 1986). The presence of horizontal zones of mechanical detachment is also supported by deep reflection seismic profiling in the Mojave Desert and Transverse Ranges which shows a series of nearly flat reflectors in the

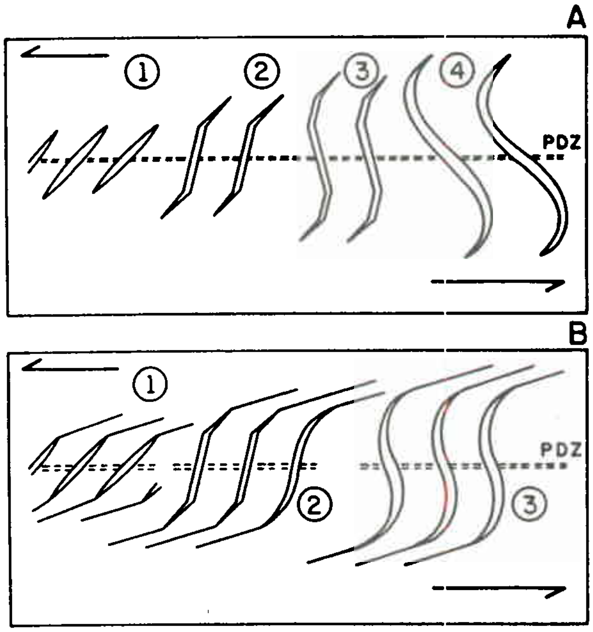


Figure 30. Rotation and progressive propagation of fractures in a left-shear couple (redrawn after Wilson, 1960). Extension (T) fractures form (1), then in progressive simple shear, they rotate counter-clockwise but continue to propagate either parallel to the direction of the initial fracture (A) or parallel to the direction favorable for the formation of R fractures (B), that is, at an angle $\phi/2$ to the principal displacement zone (PDZ). See also Figure 6.

upper crust, the highest one of which shoals eastward and rises to the surface at the traces of one or two major faults in the Mojave Desert (Cheadle and others, 1986), although several of the COCORP reflectors have been reinterpreted recently as "side-swipe" effects (Serpa, 1988).

Additional P-wave analyses revealed that the distribution of relatively "fast" and "slow" crust is moderately well outlined by the positions of the major faults in the upper crust, but not in the lower crust, suggesting that the major strike-slip faults do not extend beneath the seismogenic zone into the lower crust (Hearn and Clayton, 1986a, 1986b). When the focal mechanisms of a very large sample of earthquakes are analyzed, the only ones with thrust mechanisms correct for a detachment below rotated slabs are not only the deepest in southern California, but are also just those deep earthquakes located beneath the central Transverse Ranges (Hearn and Clayton, 1986a, 1986b). Not coincidentally, the central Transverse Ranges are not isostatically compensated; they lack a root.

These studies, together with the thrust-fault interpretation of the Whittier Narrows earthquake of 1 October 1987 (Hauksson and others, 1988), have revived the notion of "flake tectonics" (Oxburgh, 1972), that is, of "thin-skinned tectonics" in southern California (Yeats, 1968, 1981), wherein much or most of the Transverse Ranges is detached on deep and blind, north-dipping thrust faults (Webb and Kanamori, 1985). Above them, high-strength, flake-like slabs up to 15 km thick have detached from the lower crust to offset strike-slip faults at depth (Hearn and Clayton, 1986a, 1986b; Dewey and others, 1986; Lemiszki and Brown, 1988). In this model, the slabs may slide away from one another and are free to rotate externally about a vertical axis in the prevailing regime of simple shear. Webb and Kanamori (1985) postulated that perhaps even the San Andreas fault is offset at depth, but interpretations of seismic reflection data across several segments of the fault seem to negate that hypothesis (Lemiszki and Brown, 1988).

These are profound hypotheses and revelations in southern California, where notions of a simple transform plate boundary have prevailed since Hamilton (1961) and Wilson (1965), following Carey (1958), gave plate-tectonic bases to explain how horizontal displacements, measurable in hundreds of kilometers on the San Andreas fault, terminated abruptly at its southern and northern ends. The earthquake hazards posed by the blind thrust faults beneath the Transverse Ranges give southern California considerable reason for concern.

STRIKE-SLIP FAULTS IN CONVERGENT PLATE MARGINS

Orogen-Parallel Strike Slip

Suture zones make ideal strike-slip boundaries to accommodate much of the horizontal plate motion, as proposed for the Indo-Eurasia collision (Fitch, 1972; Karig, 1980; Tapponier and others, 1986). These motions result in part from the conversion of the horizontal component of oblique convergence onto a discrete strike-slip fault behind or within the magmatic arc, resulting in strike-parallel shuffling of terranes along major, trench-linked, strike-slip faults (Sarewitz and Karig, 1986). Orogen-parallel strike-slip faults are particularly prevalent within present-day, subduction-arc complexes characterized by oblique plate convergence (Fitch, 1972; Oxburgh, 1972; Saleeby, 1978), the most popularly cited example of this phenomenon being the Semangko fault system (Page and others, 1979; Karig, 1980; Hla Muang, 1987). Now that we have a better idea of the various roles and settings of strike-slip faults, a revival is occurring in the recognition of strike slip along the length of many ancient orogenic belts (Wilson, 1962; Reed and Bryant, 1964; Webb, 1969; Tapponier, 1977; Bradley, 1982; Gates and others, 1986; Sarewitz and Karig, 1986; Costain and others, 1987; Ferrill and Thomas, 1988; Şengör, in press).

Strike-slip faulting is a ubiquitous process in volcanic arcs of most subduction zones that have a continental overriding plate (Jarrard, 1986), and arc volcanoes such as Mount St. Helens are located in extensile zones above a deep shear zone which itself is preferentially oriented to accommodate horizontal strike slip within the volcanic arc (Weaver and others, 1987). It is noteworthy that the majority of volcanic eruptions are preceded by earthquakes having strike-slip focal mechanisms (Zobin, 1972; Weaver and others, 1981).

Geologic and paleomagnetic evidence is building an interesting and complex story of orogen-parallel lithospheric translation in mid-Jurassic to early Tertiary time in the great Cordilleran orogenic belt from the west coasts of Mexico, United States, and Canada on into Alaska as predicted by Atwater (1970), and which may be a precursor of the kinds of motions to be expected in orogenic belts elsewhere. In the Cordilleran orogen, the zone of transport lay within the trench or bordered the volcanic arc, perhaps as a series of trench-linked, strike-slip faults or boundary transforms at various places in time and space. Strands of the fault system had 175 km of dextral displacement in Late Cretaceous or early Tertiary time in what is now the trans-Mexican volcanic belt (Gastil and Jensky, 1973), as much as 1,000 km of dextral displacement along western Mexico (Karig and others, 1978), at least 1,200 km of dextral displacement west of what is now Baja California (Hagstrum and others, 1985), and at least 900 km of dextral displacement off the present coasts of northern California and Oregon (Champion and others, 1984; Bourgeois and Dott, 1985). The orogen is bent in Washington and northeast Oregon (Carey, 1958; Wise, 1963); then it regains its straight northwest trend north of the Shuswap terrane and emerges again in the northern Rocky Mountain–Tintina Trench system in western Canada, where at least 750 km, probably more than 900 km, of dextral displacement occurred in Middle Jurassic to early Cenozoic time (Gabrielse, 1985).

Opposed to these dextral motions are mid-Cretaceous sinistral movements of from 700 to 800 km that have been postulated for other major faults in the orogen, including the Sur-Nacimiento fault in central California (Seiders, 1983; Dickinson, 1983), and the Mojave-Sonora megashear in northern Mexico in mid-Mesozoic time, which is postulated to extend into eastern California (Silver and Anderson, 1974; Anderson and Schmidt, 1983). The direction of fault slip, whether dextral or sinistral, on the various strike-slip faults in the trench-linked, strike-slip fault system in the Cordilleran orogen would certainly have depended on the angle of oblique convergence and the relative motion of one plate relative to the other at any given time. Displacement along faults which bound translating terranes may also have opposing senses of movement at any given time.

Much of the discussion about the origin, movement, and eventual resting place of suspect terranes along the west margin of the Cordilleran orogen in Mesozoic time (Jones and others, 1983; Beck, 1986) is involved in this history, simply because of the requirement of horizontal translation for terrane movement as is seen today along active, oblique convergent plate boundaries (Sarewitz and Karig, 1986).

Tectonic Escape

Along irregular convergent or collision fronts, nodes of constriction cause discrete pieces of crustal fragments and splinters termed "scholles"[4] to be expelled sideways on strike-slip faults having opposed senses of slip toward zones of overthrusting or free faces formed by subduction zones. This process of "tectonic escape" (Burke and Şengör, 1982) is a consequence of the inability of trenches to consume continental crust (McKenzie, 1972) and is possible because of the extreme heterogeneity and low shear strength of continental rocks (Şengör and others, 1985). The boundary force is produced by the buoyancy of the continental crust, not its strength. "The surface expression of this buoyancy is a mountain belt, and it is therefore possible to regard the gravitational forces caused by surface elevations as the driving forces for the motion of small plates" (McKenzie, 1972, p. 181).

One of the best examples of tectonic escape is illustrated by the Anatolian scholle in Turkey (Dewey and Şengör, 1979); it is driven westward between the dextral North Anatolian fault and the sinistral East Anatolian fault in response to the northward collision of the Arabian plate into the southern part of the Eurasian plate (Fig. 31). Similarly and on a larger scale, a great wedge of south China is being expelled eastward and southeastward between enormous intracontinental transform faults, the sinistral Altyn Tagh and Kunlung faults and the dextral Red River fault, onto the Philippine and Pacific plates (Tapponier and others, 1982, 1986). There, escape tectonics has progressed from south to north, causing reversals of movement on the participating strike-slip faults, and shunting ever larger scholles of the Asian lithosphere eastward as India continues its northward penetration into southern Asia (Tapponier and others, 1986). On a smaller, intracontinental scale, the wedge-shaped Mojave Desert in southern California (Fig. 9) is a scholle being driven eastward between the sinistral Garlock fault and the dextral San Andreas fault in response to the localized upper-crustal shortening between irregular edges of the Pacific and North American plates (McKenzie, 1972; Cummings, 1976).

An interesting case of the tectonics of a block mosaic (Hill, 1982) is in southern China where north-south shortening and east-west elongation of a tectonic domain composed of a number of blocks cause horizontal slip of several blocks relative to one another, thus opening large basins between blocks that are wedged away from one another along strike-slip faults (Wu and Wang, 1988).

FOUR PROBLEMS

Much of this paper has focused on the peculiarities of the San Andreas fault, but just how representative are the kinematics, dynamics,and seismic behavior of the San Andreas fault, of strike-slip faults in general, and for boundary transform faults in particular? The answer is important for earthquake-hazard assessment and for understanding strike-slip mechanics and tectonics. The lessons learned from the boundary transform in California will be watched with interest by the tectonic community worldwide, and just as many strike-slip concepts have evolved in southern California and spread elsewhere, so also will these latest revelations.

Four fundamental problems bearing on the geophysical nature of strike-slip faults are exemplified by lessons learned from the San Andreas fault. (1) What are the implications of mechanical stratigraphy for determination of strike-slip structural styles? (2) Why does the San Andreas fault lack a heat flow anomaly (Lachenbruch and Sass, 1980)? (3) Why are measurements of present surficial stress discordant with predictions made from kinematic models (Mount and Suppe, 1987)? (4) Why do paleoseismic determinations of relative plate velocity across the fault fall short of those determined from paleomagnetic data from the sea floor (Weldon and

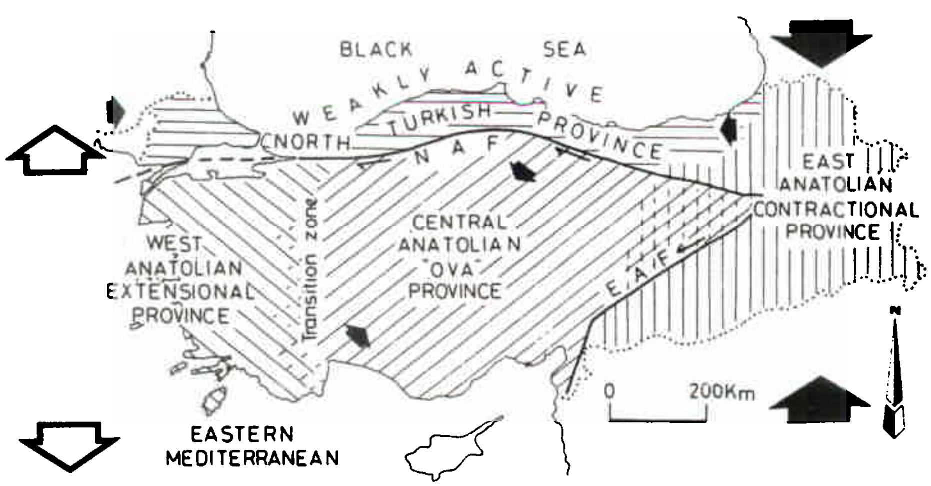


Figure 31. Westward escape of Anatolian scholle between North Anatolian (NAF and East Anatolian (EAF) faults due to collision and continued convergence of the Arabian and Eurasian plates (from Şengör and others, 1985). Reproduced with permission of Society of Economic Paleontologists and Mineralogists and of Şengör, 1988.

[4] The term "Scholle" is a German word for a floe (of ice), a flake, or a clod (of soil), and it is an excellent descriptive word for a tectonic "flake," "slab," "block," "crustal panel," or even "terrane," where the tectonic domain in question is detached from the lithosphere and therefore cannot be termed a "plate" (Dewey and Şengör, 1979). Burke and Şengör (1982) seem to have extended scholle to include small lithospheric domains, but I believe that "miniplate" or "microplate" are good terms for small plates, and I recommend that scholle be used as Dewey and Şengör originally intended.

Humphreys, 1986; Minster and Jordan, 1984, 1987)?

Mechanical Stratigraphy

Only recently have the implications of mechanical stratigraphy become apparent in considerations of strike-slip structural styles. Although Sylvester and Smith (1976) certainly acknowledged the role of detachments in the transpressional deformation that molded the Mecca Hills, it has been the work by Yeats (1981), Terres and Sylvester (1981), Webb and Kanamori (1985), Nicholson and others (1986), Namson and Davis (1988), and others in central and southern California that has raised many fundamental questions about the widely held assumption that strike-slip faults are vertical to great depth. That dogma grew chiefly out of clay-model laboratory experiments in which the clay lacked mechanical layering. Just as the revelations of detachment have altered our concepts of extensional deformation, so also will they alter our views of strike-slip deformation.

Heat Flow

At this writing, a deep hole is being drilled near the San Andreas fault where it transects the Transverse Ranges to answer the second two questions posed above. Near-surface stress measurements (0 to 1 km) show an increase of stress with depth consistent with estimates based on laboratory frictional experiments (about 90 bars/km). Extensive conductive heat-flow measurements made near the fault, however, show no discernible effects of frictional heating, suggesting that there is little frictional heating on the fault, and that the upper limit for the average shear stress on the fault is less than 200 bars. The near-surface stress measurements therefore cannot be extrapolated to depths greater than 3 km without violating existing heat-flow constraints. Several hypotheses have been advanced to explain the lack of a heat-flow anomaly. Among them is the idea that movement along the fault is in response to very low stress, or that some unknown "cooling mechanism" is at depth, such as subsurface water flow, which reduces temperature (O'Neil and Hanks, 1980). Geologic mapping, seismic studies, and regional gravity data suggest that the shallow low-angle thrusts or detachment faults believed to underlie the eastern Transverse Ranges would disperse the heat and support a thermo-mechanical model that explains the low heat flow on the main fault trace.

Recent geophysical studies suggest that the surface trace of the San Andreas fault is offset from 50 to 200 km westward on a subhorizontal detachment from its continuation at depth down to the top of the seismogenic zone (Webb and Kanamori, 1985; Meisling and Weldon, in press). Furlong (1987, p. 1509) has outlined the thermo-mechanical controls on the three-dimensional geometry and lithospheric evolution in the San Francisco Bay region as follows: "Because of the lithospheric geometry inherited from the period of active subduction, the thermal regime evolves to produce a mechanical environment where shear deformation in the lower crust and upper mantle is concentrated in a near vertical region offset approximately 50 km east of the surface San Andreas. The offset produces a complex geometry for the plate boundary in which a sub-horizontal detachment (shear zone?) must connect the surface and deeper segments of the plate boundary." Furlong continued: "the near surface fault zone formed in this way will have an orientation controlled by the orientation of the deeper shear boundary rather than that associated with the regional stress regime. In addition, through this process slivers of the North American plate are 'acquired' by the Pacific plate, producing the complex pattern of juxtaposed terranes outboard of the San Andreas fault in central California."

The presence and structure of deeper shear boundaries and detachments will be difficult to elucidate, because the seismogenic zone is confined to the upper continental crust in regions of active strike-slip faulting (Chen and Molnar, 1983; Sibson, 1983). This means that seismicity cannot give direct information on the location and nature of aseismic shear zones and detachments that may exist in the lower crust and mantle.

The Stress Paradox

Certainly it is an oversimplification to state that the oblique resolution of stress across a strike-slip fault yields folds or thrusts perpendicular to the shortening axis, especially where the evidence at hand says that the present maximum compression, which is ostensibly parallel to the shortening axis, is normal to the San Andreas fault both at Cajon Pass (Zoback and others, 1987) and in central California (Mount and Suppe, 1987). Such stresses would have to be maintained by the plate motion; otherwise, they would be relaxed quickly by earthquakes. Mackenzie (1972, p. 176) maintained that "it is therefore simpler at present to neglect the stress field, which must be exceedingly complicated, and account for the features produced by major deformation directly in terms of plate motion."

The stresses, however, have to be accounted for eventually, and so these kinds of models must be refined and studied more extensively to resolve the paradox that observed fault-parallel well-breakouts and bore-hole elongations imply that the compression axis is normal close to the fault, whereas simple shear theory says that the regional compression axis ought to be 45° to the fault. Stress measurements at various levels in the Cajon Pass deep hole suggest provisionally that *fault-normal compression results from a generally high level of deviatoric stress in the crust but which, in the vicinity of a weak fault, must be oriented in such a manner that the level of shear stress resolved on the fault is low* (Zoback and others, 1987); that is, either the least or the greatest principal stress must be nearly perpendicular to the fault.

Plate Velocity Discrepancies

Paleomagnetic reversals provide the principal control on the velocity of the Pacific relative to the North American plate and yield a rate of 55 mm/yr (Weldon and Humphreys, 1986; Minster and Jordan, 1984, 1987), whereas historic geodetic evidence yields a rate of about 35 mm/yr (Thatcher, 1979). This discrepancy has been known for quite some time, and the practice has been to explain it by apportioning the residual on faults east of the San Andreas fault well into the Great Basin, or on faults west of the San Andreas fault in the Pacific Ocean. Another rational explanation is that plate movements are episodic on a scale of thousands of years, so that the historic geodetic movements give too short a sample of geologic time. Other writers say that the discrepant rates prove that the plate motion has decreased from 55 mm/yr to 35 mm/yr. Still others have reappraised the interpretations of the paleomagnetic data and conclude that the rate may be only about 45 mm/yr. The lower figure, then, yields a smaller discrepancy which might be explained by combinations of all of the other hypotheses. I believe that the solution will come from acquisition of more geodetic data over longer time periods more representative of geologic time, together with more geologic work focused on the timing of local tectonic events to gain a better understanding of Neogene plate movements. The question may be resolved in the next decade with the Global Positioning Satellite system providing data on the scale of the plates themselves at distances, sampling frequencies, and precision undreamed of only a decade ago.

CONCLUDING STATEMENT

The recent revelations pertaining to concepts of strike-slip faults have dramatically broadened our views about the tectonic roles of strike-slip faults in time and space. For example, the terrane concept has changed notions about how deformation in mobile belts is accomplished by bringing a third dimension to views of microplate tectonics along continental margins. Strike-slip faults are the key structures in tectonic migration along plate boundaries of terranes. Now intensive efforts should focus on reviewing the tectonic nature and history of plate boundaries to place limits on their reconstructed geometry.

Of at least equal importance to understanding strike-slip tectonics are recent interpretations of seismic reflection studies of the intermediate and deep crust which imply the presence of detachments, perhaps at several levels within the crust, and which indicate that some strike-slip faults are cut at depths by those detachments. These interpretations have shaken the widely held dogmas about all strike-slip faults being vertical to great depth, about the orientation and partitioning of stress at different crustal levels near strike-slip faults, and about the strength of strike-slip faults and their capacity to store seismic strain energy. The implications of these concepts are far-reaching and concern such diverse disciplines as mineral and hydrocarbon exploration, paleotectonics, paleogeography, and seismic hazards. These revelations provide enormously stimulating ideas to challenge those who study tectonics in general and strike-slip faults in particular, and they certainly portend exciting times of discovery and understanding in the future.

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