

## DEFORMATION AND BASIN FORMATION ALONG STRIKE-SLIP FAULTS<sup>1</sup>

NICHOLAS CHRISTIE-BLICK,

*Department of Geological Sciences and  
Lamont-Doherty Geological Observatory of Columbia University  
Palisades, New York 10964;*

AND

KEVIN T. BIDDLE

*Exxon Production Research Company  
P.O. Box 2189  
Houston, Texas 77252-2189*

**ABSTRACT:** Significant advances during the decade 1975 to 1985 in understanding the geology of basins along strike-slip faults include the following: (1) paleomagnetic and other evidence for very large magnitude strike slip in some orogenic belts; (2) abundant paleomagnetic evidence for the pervasive rotation of blocks about vertical axes within broad intracontinental transform boundaries; (3) greater appreciation for the wide range of structural styles along strike-slip faults; (4) new models for the evolution of strike-slip basins; and (5) a body of new geophysical and geological data for specific basins. In the light of this work, and as an introduction to the remainder of the volume, the purpose of this paper is to summarize the major characteristics of and controls on structural patterns along strike-slip faults, the processes and tectonic settings of basin formation, and distinctive stratigraphic characteristics of strike-slip basins.

Strike-slip faults are characterized by a linear or curvilinear principal displacement zone in map view, and in profile, by a subvertical fault zone that ranges from braided to upward-diverging within the sedimentary cover. Many strike-slip faults, even those involving crystalline basement rocks, may be detached within the middle to upper crust. Two prominent characteristics are the occurrence of en echelon faults and folds, within or adjacent to the principal displacement zone, and the co-existence of faults with normal and reverse separation. The main controls on the development of structural patterns along strike-slip faults are (1) the degree to which adjacent blocks either converge or diverge during strike slip; (2) the magnitude of displacement; (3) the material properties of the sediments and rocks being deformed; and (4) the configuration of pre-existing structures. Each of these tends to vary spatially, and, except for the last, to change through time. It is therefore not surprising that structural patterns along strike-slip faults differ in detail from simple predictions based on the instantaneous deformation of homogeneous materials. In the analysis of structural style, it is important to attempt to separate structures of different ages, and especially to distinguish structures due to strike-slip deformation from those pre-dating or post-dating that deformation. Distinctive aspects of structural style for strike-slip deformation on a regional scale include evidence for simultaneous shortening and extension, and for random directions of vergence in associated thrusts and nappes.

Sedimentary basins form along strike-slip faults as a result of localized crustal extension, and, especially in zones of continental convergence, of localized crustal shortening and flexural loading. A given basin may alternately experience both extension and shortening through variations in the motion of adjacent crustal blocks, or extension in one direction (or in one part of the basin) may be accompanied by shortening in another direction (or in another part of the basin). The directions of extension and shortening also tend to vary within a given basin, and to change through time; and the magnitude of extension may be depth-dependent. Theoretical studies and observations from basins where strike-slip deformation has ceased suggest that many strike-slip basins experience very little thermally driven post-rift subsidence. Strike-slip basins are typically narrow (less than about 50 km wide), and they rapidly lose anomalous heat by accentuated lateral as well as vertical conduction. Detached or thin-skinned basins also tend to be cooler after rifting has ended than those resulting from the same amount of extension of the entire lithosphere. In some cases, subsidence may be arrested or its record destroyed as a result of subsequent deformation. Subsidence due to extension, thermal contraction, or crustal loads is amplified by sediment loading.

The location of depositional sites is determined by (1) crustal type and the configuration of pre-existing crustal structures; (2) variations in the motion of lithospheric plates; and (3) the kinematic behavior of crustal blocks. The manner in which overall plate motion is accommodated by discrete slip on major faults, and by the rotation and internal deformation of blocks between those faults is especially important. Subsidence history cannot be determined with confidence from present fault geometry, which therefore provides a poor basis for basin classification. Every basin is unique, and palinspastic reconstructions are useful even if difficult to undertake.

Distinctive aspects of the stratigraphic record along strike-slip faults include (1) geological mismatches within and at the boundaries of basins; (2) a tendency for longitudinal as well as lateral basin asymmetry, owing to the migration of depocenters with time; (3) evidence for episodic rapid subsidence, recorded by thick stratigraphic sections, and in some marine basins by rapid deepening; (4) the occurrence of abrupt lateral facies changes and local unconformities; and (5) marked differences in stratigraphic thickness, facies geometry, and occurrences of unconformities from one basin to another in the same region.

### INTRODUCTION

Strike-slip deformation occurs where one crustal or lithospheric block moves laterally with respect to an adjacent block. In reality, most "strike-slip" faults accommodate oblique displacements along some segments or during part of the time they are active; and most are associated with an assemblage of related structures including both normal and reverse faults. A component of oblique slip is also required for the formation of sedimentary basins. However,

the title of this paper and that of the volume were chosen to emphasize tectonics and sedimentation in regions where strike-slip deformation is prominent.

In this paper, we follow Mann et al. (1983) in using the term "strike-slip basin" for any basin in which sedimentation is accompanied by significant strike slip. We acknowledge that at any given time some strike-slip basins are hybrids associated with regional crustal extension or shortening, and most are composite, influenced by varying tectonic controls during their evolution (for various recent perspectives about sedimentary basins in a plate-tectonic framework see Green, 1977; Dickinson, 1978; Bally and Snelson, 1980; Klemme, 1980; Bois et al., 1982; Dewey,

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1982; Reading, 1982; Kingston et al., 1983a; Perrodon and Masse, 1984). Strike-slip basins also occur in a wide range of plate-tectonic settings including (1) intracontinental and intraoceanic transform zones; (2) divergent plate boundaries and extensional continental settings; and (3) convergent plate boundaries and contractional continental settings.

Among numerous articles on aspects of strike-slip deformation, basin formation, and sedimentation published in the last decade or so, several have been influential (Wilcox et al., 1973; Crowell, 1974a, b; Freund, 1974; Sylvester and Smith, 1976; Segall and Pollard, 1980; a collection of articles edited by Ballance and Reading, 1980a; Reading, 1980; Aydin and Nur, 1982a; Mann et al., 1983). Sylvester (1984) provides a compilation of classic papers on the mechanics, structural style, and displacement history of strike-slip faults, with emphasis on examples from California. In view of these existing summaries, we draw attention here to significant recent advances in understanding the large-scale characteristics of strike-slip faults and strike-slip basins, including several topics not discussed or only briefly mentioned in Ballance and Reading (1980a) and Sylvester (1984).

The remainder of this summary paper introduces two broad themes, which are elaborated in the articles that follow: (1) the characteristics of and controls on structural patterns along strike-slip faults; and (2) the processes and tectonic settings of basin formation, and distinctive aspects of the stratigraphic record of strike-slip basins.

#### PROGRESS DURING THE DECADE 1975-1985

Examples of advances in understanding the geology of strike-slip basins during the decade 1975 to 1985 include the following: (1) paleomagnetic and other evidence suggesting the very large magnitude of strike slip in some orogenic belts; (2) abundant paleomagnetic evidence where continents are intersected by diffuse transform plate boundaries for pervasive rotation of blocks about vertical axes, with implications for processes of deformation, basin formation and palinspastic reconstruction of strike-slip basins; (3) greater appreciation for the range of structural styles along strike-slip faults, both on the continents and in the ocean basins, and for the processes by which those styles arise; (4) new theoretical and empirical models for the evolution of strike-slip basins; and (5) new geophysical and geological data for many strike-slip basins, some of which are reported in this volume.

#### *Large-Magnitude Strike Slip in Orogenic Belts*

The role of strike slip in the evolution of orogenic belts and associated sedimentary basins has been recognized for several decades (references in Ballance and Reading, 1980b; and Sylvester, 1984), but the possible magnitude of such deformation may have been underestimated. A combination of paleomagnetic, faunal, and other geological data now indicate that in some complex orogens such as the North American Cordillera, individual elements of the tectonic collage (Helwig, 1974), termed terranes, have moved thousands not merely hundreds of kilometers with respect to each other along the trend of the orogen (Jones et al., 1977;

Irving, 1979; Beck, 1980; Coney et al., 1980; Irving et al., 1980; Champion et al., 1984; Eisbacher, 1985 this volume). Although part of this longitudinal displacement in the Cordillera pre-dates accretion, a significant component occurred after the terranes were sutured to North America. The Stikine Terrane, for example, appears to have been displaced northward by 13° to 20° since early Cretaceous time, or on the order of 1,500 km with respect to the North American craton (Jones et al., 1977; Irving et al., 1980; Chamberlain and Lambert, 1985; Eisbacher, 1985 this volume).

Large-scale strike-slip deformation also permits lateral tectonic escape in zones of continental convergence, such as between India and Eurasia (Molnar and Tapponnier, 1975), or on a smaller scale, in Turkey (Şengör et al., 1985 this volume). Molnar and Tapponnier (1975) estimated that about one third to half of the relative plate motion between India and Eurasia since the onset of continental collision in Eocene-Oligocene time (at least 1,500 km) could be accounted for by a comparable amount of strike-slip faulting in China and Mongolia.

We do not imply that such huge displacements characterize all orogenic belts, or that available paleomagnetic results are in every case without ambiguity, but only that the possibility of large-scale strike slip should be seriously entertained unless precluded by firm data. In the Appalachian-Caledonide orogen, for example, paleomagnetic studies initially suggested cumulative sinistral offset of as much as 2,000 km in late Paleozoic time (Kent and Opdyke, 1978, 1979; van der Voo et al., 1979; van der Voo and Scotese, 1981; Kent, 1982; van der Voo, 1982, 1983; Perroud et al., 1984). The timing, magnitude, and sense of displacement have recently been questioned on the basis of (1) new determinations of the Early Carboniferous pole for cratonic North America (Irving and Strong, 1984; Kent and Opdyke, 1985); (2) doubt about the age of the magnetization directions measured in some of the samples from eastern North America and Scotland (Donovan and Meyerhoff, 1982; Roy and Morris, 1983; Cisowski, 1984); and (3) the difficulty of finding appropriate faults on which to distribute the displacement (Ludman, 1981; Bradley, 1982; Donovan and Meyerhoff, 1982; Parnell, 1982; Winchester, 1982; Smith and Watson, 1983; Briden et al., 1984; Haszeldine, 1984). Where large displacements have occurred, however, even relatively young basins may have been dismembered and strung out over huge distances, and some geological mismatches may be resolved only by considering the history of an entire orogenic belt.

#### *Rotations About Vertical Axes*

The rotation of blocks about vertical axes and the bending of segments of orogenic belts have long been postulated on structural grounds (Carey, 1955, 1958; Albers, 1967; Freund, 1970; Garfunkel, 1974; Dibblee, 1977). Paleomagnetic data now confirm that such rotations tend to be pervasive in strike-slip regimes over a wide range of scales, especially where continents are intersected by diffuse transform plate boundaries (Figs. 1, 2), and the data suggest additional constraints on the timing of rotation and on the kinematics of deformation (Beck, 1980; Cox, 1980; Lu-

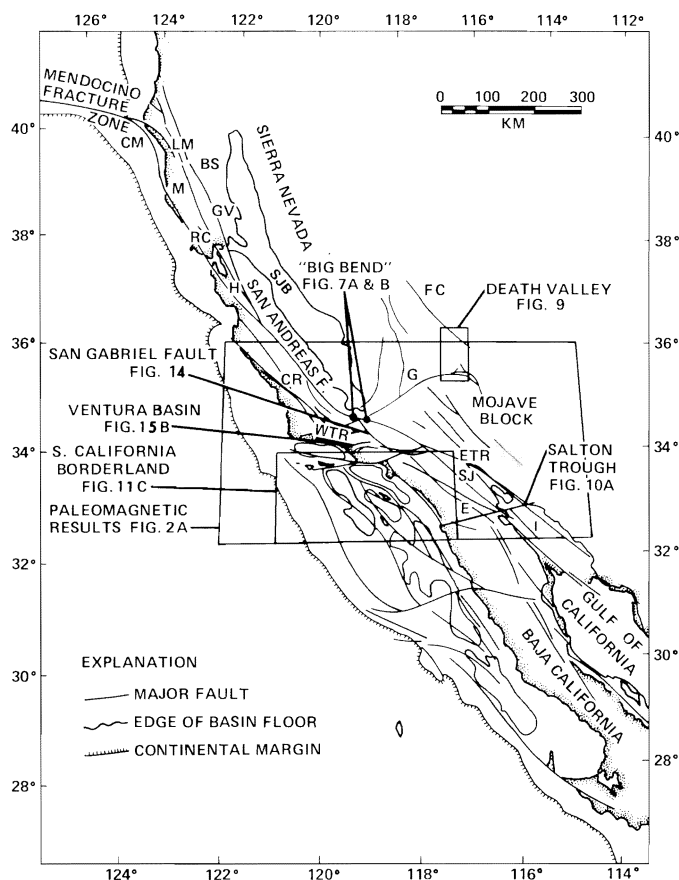


Fig. 1.—Major faults in the diffuse transform plate boundary of California and adjacent parts of Baja California, together with selected basins and elevated blocks mentioned in the text: BS, Bartlett Springs fault zone; CM, Cape Mendocino; CR, Coast Ranges; E, Elsinore fault; ETR, eastern Transverse Ranges; FC, Northern Death Valley-Furnace Creek fault zone; G, Garlock fault; GV, Green Valley fault zone; H, Hayward fault zone; I, Imperial fault; LM, Lake Mountain fault zone; M, Maacama fault zone; RC, Healdsburg-Rodgers Creek fault zone; SJ, San Jacinto fault; SJB, San Joaquin Basin; WTR, western Transverse Ranges (from King, 1969; Crowell, 1974a; Herd, 1978; Nilsen and McLaughlin, 1985 this volume). Map shows locations of areas and cross sections illustrated in Figures 2A, 7, 9, 10A, 11C, 14, 15B.

yendyk et al., 1980, 1985; Ron et al., 1984; Ron and Eyal, 1985).

Paleomagnetic evidence from Neogene rocks of southern California indicates that blocks such as the western Transverse Ranges (WTR in Fig. 1), bounded by east-striking left-slip faults, have experienced net clockwise rotations of between  $35^\circ$  and  $90^\circ$ , with sites near one major right-slip fault being rotated by more than  $200^\circ$  (near DB in Fig. 2A; Luyendyk et al., 1985). In contrast, the Mojave block, located between the San Andreas and Garlock faults, and characterized by northwest-striking right-slip faults, seems to have been rotated counterclockwise by about  $15^\circ \pm 11^\circ$  since 6 Ma (c in Fig. 2A; Morton and Hillhouse, 1985), but with large variations in declination over distances of 30 to 120 km from one sub-block to another. In the Cajon Pass region (f in Fig. 2A), there has been no significant rotation since 9.5 Ma (Weldon et al., 1984). These results generally support the tectonic models of Freund (1970), Garfunkel (1974), and Dibblee (1977), in which major crustal blocks

deform internally like a set of dominoes. Garfunkel (1974) suggested as much as  $30^\circ$  of rotation for the Mojave block, approximately twice that determined paleomagnetically, but he probably overestimated by a factor of two to three the magnitude of displacement on the strike-slip faults (Dokka, 1983).

The dimensions of the rotating blocks are uncertain for three reasons: (1) As noted above, blocks tend to deform internally, producing dispersion in declination anomalies; (2) In some cases, rotation seems to have occurred during deposition or eruption of the sedimentary and volcanic rocks studied, so that there is a systematic variation in the magnitude of rotation with age (Luyendyk et al., 1985); (3) Similar rotations may characterize blocks that were actually behaving independently. It is also likely that the boundaries of crustal fragments have changed through time, as displacement occurred sequentially along different strike-slip faults. Some blocks may have undergone, at different times, both clockwise and counterclockwise rotation. A possible example is the San Gabriel block between the San Gabriel and San Andreas faults (SB and e in Fig. 2A), although available data are inconclusive (Ensley and Verosub, 1982; Luyendyk et al., 1985). Small blocks can rotate at an alarming speed. Plio-Pleistocene sediments of the Vallecito-Fish Creek Basin in the western Imperial Valley (d in Fig. 2A) have been rotated  $35^\circ$  since 0.9 Ma (Opdyke et al., 1977; Johnson et al., 1983), and there is thus no assurance that rotation accumulates at a uniform rate, any more than does displacement.

Paleomagnetic and structural data from northern Israel indicate that Neogene intraplate deformation was accommodated by block rotation and strike-slip deformation similar to but on a smaller scale than that documented in California (Fig. 2B; Ron et al., 1984; Ron and Eyal, 1985). Domains of left-slip faults (Galilee and Carmel regions) have rotated clockwise by  $23^\circ$  to  $35^\circ$  since the Cretaceous, and domains of right-slip faults (Galilee and Tiberias regions), counterclockwise by  $23^\circ$  to  $53^\circ$  since the Cretaceous and Miocene, respectively. Sites associated with east-striking normal faults in Galilee yield the expected Cretaceous direction. Strike slip on individual faults west of the Dead Sea fault zone (Sea of Galilee, Fig. 2B) is measurable in hundreds of meters, beginning in late Miocene to early Pliocene time (Ron and Eyal, 1985). The normal faults are predominantly of post-middle Pliocene age.

The significance of these paleomagnetic results is that strike-slip basins can no longer be viewed solely in terms of bends, oversteps or junctions along strike-slip faults, to list some popular models; rotations may play an important role in basin formation. In addition, we should expect facies and paleogeographic elements not only to be offset along strike-slip faults, but also to be systematically misaligned. A further implication of block rotations, discussed below, is that the blocks and bounding faults are detached at some level in the crust or upper mantle (Terres and Sylvester, 1981; Dewey and Pindell, 1985).

#### Structural Style of Strike-Slip Faults

There is growing appreciation for the wide range of structural styles along strike-slip faults, both on the conti-

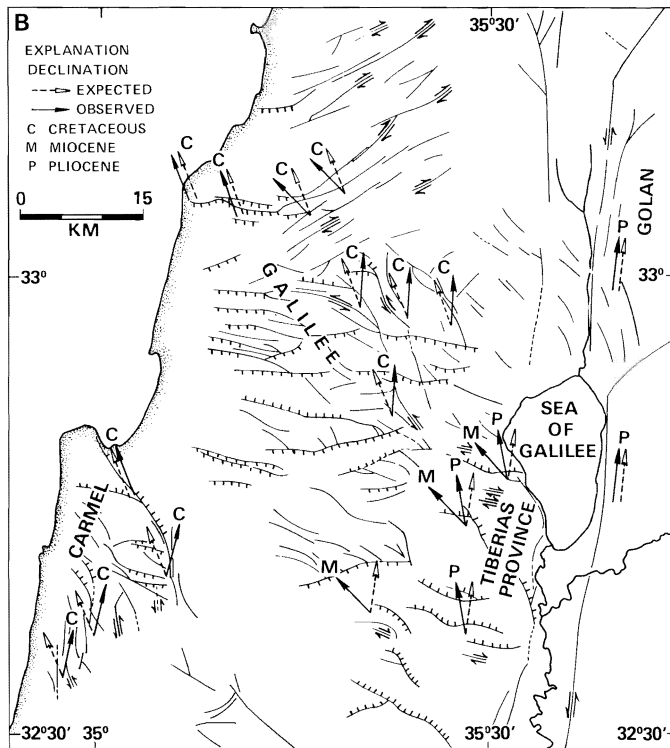
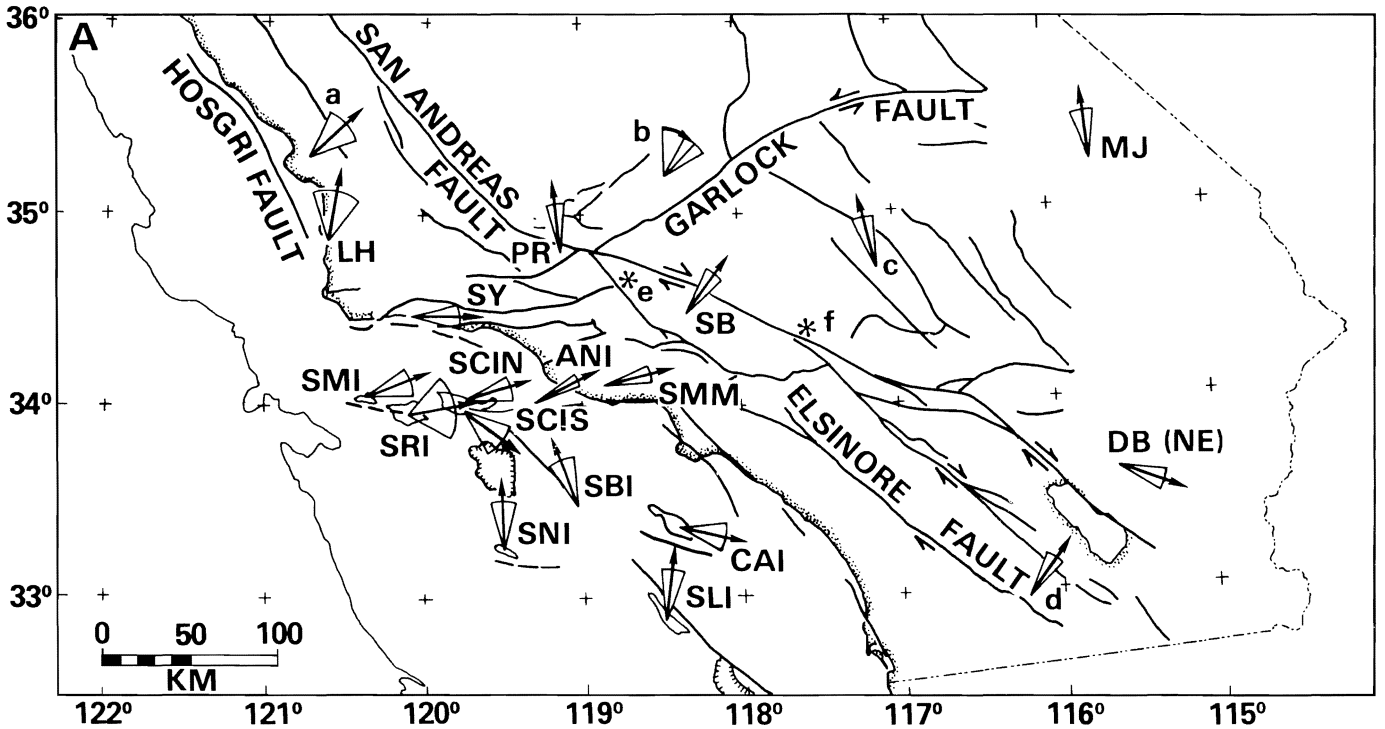


FIG. 2.—A) Paleomagnetic declinations measured in rocks of Neogene age (older than 13 Ma, except where indicated) and Quaternary sediments (site d) in southern California (see Fig. 1 for location). For each site, except e and f, the mean declination is shown along with the 95% confidence limit on the mean. Most of the data and the figure are from Luyendyk et al. (1985). Additional published data; a, Greenhaus and Cox (1979); b, apparent rotation between about 80 and 20 Ma (in comparison with the late Cretaceous pole of Irving, 1979; Kanter and McWilliams, 1981); c, mean counterclockwise rotation of  $15^\circ \pm 11^\circ$  since 6 Ma, but with large variations in declination from one sub-block to another (Morton and Hillhouse, 1985); d,  $35^\circ$  clockwise rotation since 0.9 Ma (Opdyke et al., 1977; Johnson et al., 1983); e, no major rotation since 8.5 Ma; the data can be interpreted in terms of clockwise rotation prior to 7.5 Ma, and counterclockwise rotation after that time (Ensley and Verosub, 1982); f, no significant rotation since 9.5 Ma (Weldon et al., 1984). Abbreviations for sites: ANI, Anacapa Island; CAI, Catalina Island; DB (NE) northeastern Diligencia Basin; LH, Lions Head; MJ, Mojave; PR, Plush Ranch Formation, Lockwood Valley; SB, Soledad Basin; SBI, Santa Barbara Island; SCIN, north Santa Cruz Island; SCIS, south Santa Cruz Island; SLI, San Clemente Island; SMI, San Miguel Island; SMM, Santa Monica Mountains; SNI, San Nicolas Island; SRI, Santa Rosa Island; SY, Santa Ynez Range; a, Morro Rock-Islay Hill Complex; b, southern Sierra Nevada; c, Mojave block; d, Vallecito-Fish Creek Basin; e, Ridge Basin; f, Cajon Pass. Faults are from Jennings (1975).

B) Paleomagnetic declinations measured in rocks of Cretaceous, Miocene, and Pliocene age in northern Israel; strike-slip faults, with sense of slip; and normal faults, indicated by hachures on downthrown side (simplified from Ron et al. 1984; Ron and Eyal, 1985). Domains associated with left-slip faults have undergone clockwise rotation, whereas those associated with right-slip faults have experienced counterclockwise rotation; domains associated predominantly with normal faults yield the expected declination.

nents and in the ocean basins, and for the processes by which those styles arise. Wilcox et al. (1973) recognized that one of the main controls on structural style along continental strike-slip faults cutting appreciable thicknesses of sediment is the degree to which adjacent blocks either converge or diverge during deformation. Little attention has been paid subsequently to the structures associated with divergence (see Harding et al., 1985 this volume). In addition, although experimental models, such as those of Wilcox et al. (1973), constitute a useful point of departure for analyzing the development of structures along strike-slip faults, they cannot account adequately for the considerably larger length scale, lower strain rate, greater strength, complex variations in block motion, syn-deformational sedimentation, or material heterogeneity inherent in natural examples (for further discussion, see Hubbert, 1937).

Since transform faults were first recognized in the ocean basins (Wilson, 1965; Sykes, 1967), a wealth of information has accumulated about their physiographic and structural characteristics, particularly from investigations with deep-towed echo-sounding and side-scan sonar instruments such as Deep Tow, GLORIA, Sea Beam, and Sea MARC 1 (Lonsdale, 1978; Macdonald et al., 1979; Searle, 1979, 1983; Bonatti and Crane, 1984; Fox and Gallo, 1984), the deep-towed ANGUS camera (Karson and Dick, 1983), and the deep-sea submersible ALVIN (Choukroune et al., 1978; Karson and Dick, 1983). The gross crustal structure across fracture zones has been determined by a variety of geophysical means, including seismic refraction experiments (e.g., Detrick and Purdy, 1980; Sinha and Loudon, 1983), and multichannel seismic reflection profiles (e.g., Mutter and Detrick, 1984). Fossil oceanic fracture zones have also been studied in ophiolites such as the Coastal Complex of Newfoundland (Karson, 1984). Sediment accumulations in oceanic transform zones are generally thinner than about 1 km, except in areas such as the northern Gulf of California, where there is an abundant supply of terrigenous detritus (Crowell, 1981a; Kelts, 1981). The main structural features of the transforms are therefore expressed largely in igneous and metaigneous rocks, and they differ in detail from the styles of many continental strike-slip faults.

The morphology of oceanic transforms is in part a function of the offset and spreading rate of associated ridges (Fox and Gallo, 1984), and of changes in spreading direction (Menard and Atwater, 1969; Macdonald et al., 1979; Bonatti and Crane, 1984). For example, slow-slipping transforms (full rate of 1.5 to 5 cm/yr), such as those of the North Atlantic, are characterized by prominent linear topography and aligned closed basins oriented transverse to offset ridges. The relief increases from about 1,500 m for small-offset (<30 km) transforms to several thousand meters for those with large offset (>100 km), and gradients of valley walls are typically 20° to 30°, with scarps locally near vertical (Fox and Gallo, 1984). Ridge-flank topography tends to deviate in the direction of strike-slip within a few kilometers of the transform. In contrast, many fast-slipping transforms (full rate of 12 to 18 cm/yr), such as those of the eastern Pacific, are characterized by broad zones (7 to 150 km wide), composed of numerous small-offset transform segments, linked by oblique extensional do-

mains. Relief ranges from several hundred to a few thousand meters (Fox and Gallo, 1984). The pronounced topography of some transforms may be an expression of changes in plate motion, and of convergent strike-slip deformation. For example, near its eastern intersection with the mid-Atlantic ridge, the Romanche fracture zone shallows to little more than 1,000 m below sea level, and the presence of reefal limestone shows that it was locally emergent at about 5 Ma (Bonatti and Crane, 1984). About 400 km farther east, there is evidence from a seismic reflection profile for reverse faulting and folding on the north flank of the fracture zone (Lehner and Bakker, 1983).

The investigation of oceanic transform faults is currently a major frontier in the geological and geophysical sciences, but in view of the emphasis of this volume on strike-slip deformation and basin formation in continental settings, in the remainder of this paper we focus primarily on continental examples.

### *Models for Strike-Slip Basins*

Models for strike-slip basins are of two kinds: (1) theoretical models, derived from relatively simple assumptions about the thermal and mechanical properties of the lithosphere, but which can be compared with natural basins (Rodgers, 1980; Segall and Pollard, 1980; and in this volume, Aydin and Nur, 1985; Guiraud and Seguret, 1985; Pitman and Andrews, 1985; Royden, 1985); and (2) empirical models, which represent a distillation of geological and geophysical data from several different basins (Crowell, 1974a, b; Aydin and Nur, 1982a; Mann et al., 1983; and in this volume, Nilsen and McLaughlin, 1985; Şengör et al., 1985).

One theoretical approach for investigating the development of a pull-apart basin (Fig. 11A, B below; Burchfiel and Stewart, 1966; Crowell, 1974a, b), the simplest type of basin along a strike-slip fault, is to model the state of stress, secondary fracturing, and vertical displacements produced by infinitesimal lateral displacements on overstepping discontinuities. It is assumed for the sake of simplicity that the discontinuities are planar, vertical, and parallel, that the crust is composed of an isotropic, homogeneous, linear elastic material, and that the far-field stress is spatially uniform (Rodgers, 1980; Segall and Pollard, 1980; based on earlier studies by Chinnery, 1961, 1963; Weertman, 1965). Although such models reproduce the first-order characteristics of pull-apart basins located between overstepping strike-slip faults, they do not account satisfactorily for protracted deformation, material heterogeneity, inelastic behavior, variations in the state of stress with depth, or changes in fault geometry with depth; and currently available models are not applicable to more complex geometry and kinematic history.

Another approach (e.g., Pitman and Andrews, 1985 this volume; Royden, 1985 this volume) is to consider the subsidence and thermal history of pull-apart basins using models similar to the stretching model of McKenzie (1978), but incorporating such effects as finite rifting times, accentuated lateral heat flow from narrow basins, and depth-dependent extension (Jarvis and McKenzie, 1980; Royden and

Keen, 1980; Steckler, 1981; Steckler and Watts, 1982; Cochran, 1983). In the McKenzie model, subsidence results from instantaneous, uniform extension of the lithosphere (rifting), and from subsequent cooling by vertical conduction of heat. Theoretical studies and observations from basins where strike-slip deformation has ceased suggest that many strike-slip basins experience very little thermally driven post-rift subsidence. Basins along strike-slip faults are typically narrow (less than about 50 km wide; Aydin and Nur, 1982a; and in this volume, Aspler and Donaldson, 1985; Cemen et al., 1985; Guiraud and Seguret, 1985; Johnson, 1985; Link et al., 1985; Manspeizer, 1985; Nilsen and McLaughlin, 1985; Royden, 1985; Şengör et al., 1985), and much of the thermal anomaly decays during the rifting stage (Pitman and Andrews, 1985 this volume). In some cases, the absence of evidence for post-rift thermal subsidence may also be due to subsequent deformation of the basin (Reading, 1980; Mann et al., 1983; Nilsen and McLaughlin, 1985 this volume). An example of a large pull-apart basin that may have experienced some thermal subsidence after extension ceased is the Magdalen Basin (Carboniferous), approximately 100 km by 200 km, and situated in the Gulf of St. Lawrence between Newfoundland and New Brunswick (Bradley, 1982; Mann et al., 1983). From an analysis of subsidence and heat-flow data for the Vienna Basin (Miocene), Royden (1985 this volume) has shown that extension was confined to shallow crustal levels, and that the basin formed as a pull-apart between tear faults of the Carpathian nappes (Fig. 10B below). Detached or thin-skinned basins such as the Vienna Basin, also tend to be cooler during post-rift subsidence than those produced by the same amount of extension of the entire lithosphere.

Theoretical models are useful for analyzing processes of basin formation along strike-slip faults, although they are difficult to evaluate owing to the complexities of the real world. Examples of uncertainties are as follows: (1) lithospheric and crustal thickness prior to strike-slip deformation; (2) for predominantly extensional basins, the magnitude of extension, and the variation of extension with depth and from one part of a basin to another; (3) the relative contributions of lithospheric stretching and igneous intrusion in accommodating extension; (4) the manner in which complex fault geometry evolved, and the possible role in basin subsidence of flexural loading due to crustal shortening; (5) for starved basins, poor paleobathymetry; and for non-marine basins, the lack of a suitable datum from which to measure subsidence; (6) age control, particularly in non-marine basins; (7) the degree to which lithification occurred by physical compaction or by cementation of externally derived minerals; (8) the magnitudes of background heat flow and of heat production within the basin sediments; and (9) the relative contributions of vertical and lateral heat conduction, and of fluid motion in thermal history. In spite of these uncertainties, however, modelling studies have great potential when applied to areas for which there is abundant geological and geophysical data.

Recent empirical models for strike-slip basins (e.g., Aydin and Nur, 1982a; Mann et al., 1983) emphasize the mechanisms by which basins evolve along a single strike-slip fault or at fault oversteps, but no attempt has been made

to improve on the qualitative models of Crowell (1974a, b) for more complicated basins involving intersecting strike-slip faults and significant block rotation. In spite of notable improvements in palinspastic reconstructions for orogenic belts, such restorations are commonly difficult to undertake in individual basins, which may have been deformed and dislocated by subsequent strike slip.

#### *Geophysical, Stratigraphic, and Sedimentological Data From Strike-Slip Basins*

Seismic-reflection profiles provide an important tool for basin analysis, augmenting seismic refraction, gravity, and magnetic data, which suggest only the broadest features of basin geometry. Where outcrops and wells or boreholes are few, reflection data, calibrated by available information from wells, are indispensable in developing time-stratigraphy and determining the internal geometry of the basin fill (Vail et al., 1977), and they can provide important insights into the large-scale relations between structural evolution and sedimentation. During the past decade, there has been tremendous progress in techniques for the acquisition and processing of multichannel seismic data, considerable amounts of which have been obtained by industry and academic institutions in areas affected by strike-slip deformation. Unfortunately, much of this information remains relatively inaccessible to a large part of the scientific community. One exception is a series of three volumes edited by A. W. Bally (1983) that includes several seismic profiles across strike-slip zones (D'Onfro and Glagola, 1983; Harding, 1983; Harding et al., 1983; Lehner and Bakker, 1983; Roberts, 1983). In the future, data such as these, when coupled with standard field observations, should significantly improve our understanding of the development of strike-slip basins.

Deep seismic-reflection studies, such as those undertaken in a variety of tectonic settings in the United States by COCORP and CALCRUST, in Canada by LITHOPROBE, and in Europe by BIRPS, should also provide important information about the large-scale characteristics of crustal structure associated with strike slip. By running the seismic recorders for longer than in conventional seismic experiments, it is possible to obtain images of geological structure through the crust to the Moho, and into the upper mantle (White, 1985). Direct sampling of the deep crust may also be possible through programs such as Continental Scientific Drilling.

Continued refinement of the correlation between biostratigraphy, magnetostratigraphy and the numerical timescale (e.g., Armstrong, 1978; Harland et al., 1982; Palmer, 1983; Salvador, 1985) permits improved resolution of times, durations, and rates of sedimentation. Of special significance for the analysis of strike-slip basins in complex continental settings is the more precise correlation obtainable between non-marine, marginal marine and marine successions (e.g., Royden, 1985; Şengör et al., 1985, both in this volume).

Modern facies analysis has drawn attention to the numerous factors influencing basin fill, such as basin geometry and subsidence history, climate, sediment source, drainage patterns, transport and depositional mechanisms, depositional setting, and geological age. Examples of facies

analysis in strike-slip basins are numerous papers in the volumes edited by Ballance and Reading (1980a) and by Crowell and Link (1982); and articles in this volume by Aspler and Donaldson (1985), Johnson (1985), Link et al. (1985), and Nilsen and McLaughlin (1985).

STRUCTURES ALONG STRIKE-SLIP FAULTS

*Major Characteristics*

Strike-slip faults are characterized by a linear or curvilinear principal displacement zone in map view (Fig. 3), because significant lateral displacement cannot be accommodated where there are discontinuities or abrupt changes in fault orientation without pervasive deformation within one or both of the juxtaposed blocks. For example, angular deviations of as little as 3° between 12 to 13 km-long fault segments along the southern San Andreas fault zone may be responsible for young topographic features, as well as for the spatial distribution of aseismic triggered slip (Bilham and Williams, 1985). As viewed in profile, and in places in outcrop, most prominent strike-slip faults involve igneous and metamorphic basement rocks as well as supra-crustal sediments and sedimentary rocks. Such faults are commonly termed "wrench faults," particularly in the literature of petroleum geology (e.g., Kennedy, 1946; Anderson, 1951; Moody and Hill, 1956; Wilcox et al., 1973;

Harding et al., 1985 this volume). Typically, they consist of a relatively narrow, sub-vertical principal displacement zone at depth, and within the sedimentary cover, of braided splays that diverge and rejoin both upwards and laterally (Fig. 4). Arrays of upward-diverging fault splays are known as "flower structures" (attributed to R. F. Gregory by Harding and Lowell, 1979), or less commonly, "palm tree structures" (terminology of A. G. Sylvester and R. R. Smith; Sylvester, 1984). Some strike-slip faults terminate at depth (or upward) against low-angle detachments that may be located entirely within the sedimentary section or involve basement rocks as well. Examples are high-angle to low-angle tear faults and lateral ramps of foreland thrust and fold belts (Dahlstrom, 1970; Butler, 1982; Royden et al., 1982; Royden, 1985 this volume), and tear faults associated with pronounced regional extension, as in the Basin and Range Province of the western United States (Wright and Troxel, 1970; Davis and Burchfiel, 1973; Guth, 1981; Wernicke et al., 1982; Stewart, 1983; Cheadle et al., 1985).

The distinction between wrench faults and tear faults according to whether they are "thick-skinned" or "thin-skinned" (e.g., Sylvester, 1984) is somewhat arbitrary. Several authors have suggested that crustal blocks and associated wrench faults in central and southern California are decoupled near the base of the seismogenic crust (10–15 km) from a deeper aseismic shear zone that accommodates mo-

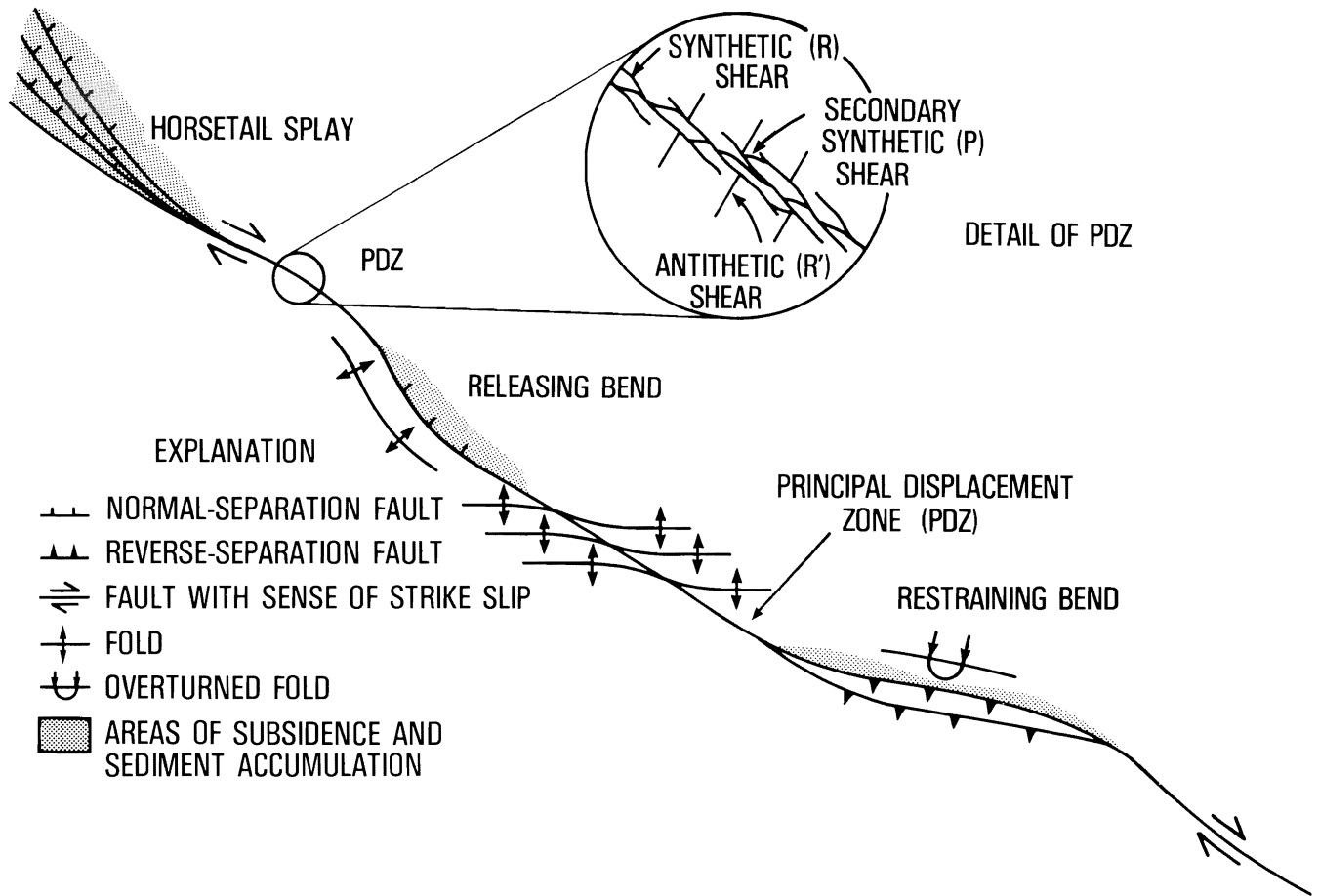
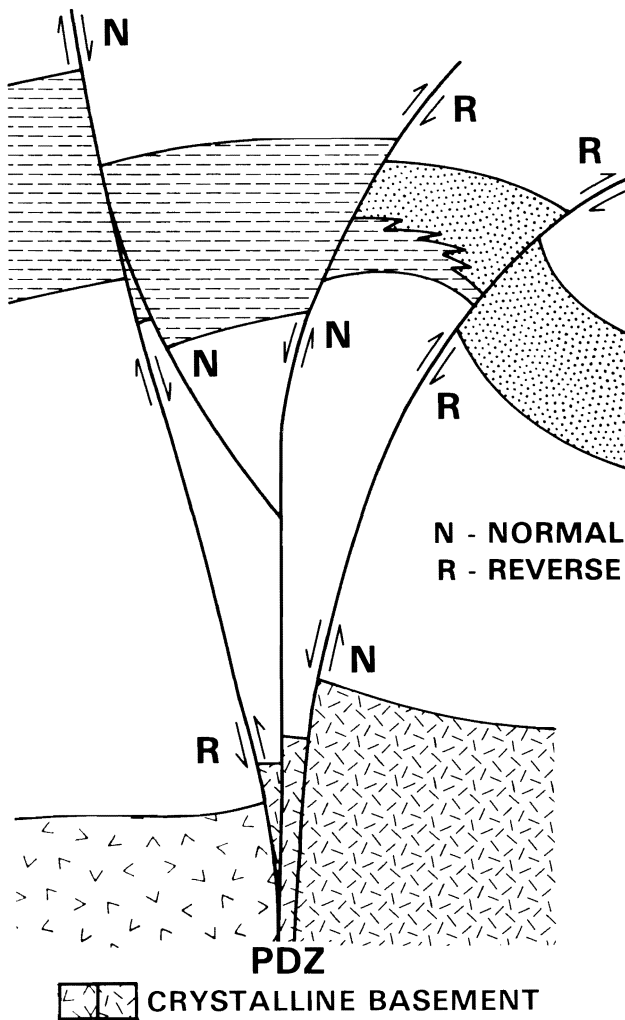


FIG. 3.—The spatial arrangement, in map view, of structures associated with an idealized right-slip fault.



### MAJOR CHARACTERISTICS

- BASEMENT - INVOLVED
- PDZ SUB-VERTICAL AT DEPTH
- UPWARD DIVERGING & REJOINING SPLAYS

### JUXTAPOSED ROCKS

- CONTRASTING BASEMENT TYPE
- ABRUPT VARIATIONS IN THICKNESS & FACIES IN A SINGLE STRATIGRAPHIC UNIT

### SEPARATION IN ONE PROFILE

- NORMAL- & REVERSE-SEPARATION FAULTS IN SAME PROFILE
- VARIABLE MAGNITUDE & SENSE OF SEPARATION FOR DIFFERENT HORIZONS OFFSET BY THE SAME FAULT

### SUCCESSIVE PROFILES

- INCONSISTENT DIP DIRECTION ON A SINGLE FAULT
- VARIABLE MAGNITUDE & SENSE OF SEPARATION FOR A GIVEN HORIZON ON A SINGLE FAULT
- VARIABLE PROPORTIONS OF NORMAL- & REVERSE-SEPARATION FAULTS

 TIME-STRATIGRAPHIC UNIT WITH VARIABLE SEDIMENTARY FACIES

FIG. 4.—The major characteristics, in transverse profile, of an idealized strike-slip fault.

tion between the Pacific and North American plates (e.g., Hadley and Kanamori, 1977; Lachenbruch and Sass, 1980; Yeats, 1981; Hill, 1982; Crouch et al., 1984; Turcotte et al., 1984; Nicholson et al., 1985a, b; Webb and Kanamori, 1985). Recent COCORP seismic reflection profiling in California suggests that the Garlock fault, long recognized as both a major wrench zone (Moody and Hill, 1956) and a tear fault bounding an extensional allochthon (Davis and Burchfiel, 1973; Burchfiel et al., 1983), may terminate downwards against a mid-crustal (9–21 km), low-angle reflecting horizon (Cheadle et al., 1985).

A prominent feature of many strike-slip faults is the occurrence of “en echelon” faults and folds within and adjacent to the principal displacement zone (Figs. 3, 5). The term en echelon refers to a stepped arrangement of relatively short, consistently overlapping or underlapping structural elements that are approximately parallel to each other, but oblique to the linear zone in which they occur (Biddle and Christie-Blick, 1985a this volume; modified from Goguel, 1948, p. 435; Cloos, 1955; Campbell, 1958; Harding and Lowell, 1979). En echelon has also been used for oblique

elements extending tens or even hundreds of kilometers from a principal displacement zone (e.g., Wilcox et al., 1973), where it is doubtful that they have much to do with strike-slip deformation (see Harding et al., 1985 this volume), and in several classic papers on thrust and fold belts for inconsistently overlapping elements arranged parallel to a zone of deformation (e.g., Rodgers, 1963; Gwinn, 1964; Armstrong, 1968; Fitzgerald, 1968; Dahlstrom, 1970). We prefer to describe the latter as a “relay” arrangement (Harding and Lowell, 1979), because it is geometrically different and because it characterizes distinctly different tectonic regimes, those associated with regional extension or with regional shortening rather than with strike slip (Harding and Lowell, 1979; Harding, 1984).

In strike-slip regimes, we also distinguish between en echelon arrangements of structures along a given principal displacement zone, and oversteps between different segments of the principal displacement zone (“en relais” of Harris and Cobbold, 1984; Biddle and Christie-Blick, 1985a this volume). Solitary oversteps (Guiraud and Seguret, 1985 this volume) and many multiple oversteps (Biddle and



Christie-Blick, 1985a this volume) do not constitute a linear zone and by our definition are not en echelon. This is not simply a matter of scale or semantics. Oversteps between different segments and en echelon structures (our usage) appear to have different origins (Aydin and Nur, 1985 this volume; see discussion below). We note, however, that there is no general agreement about such distinctions.

Idealized en echelon arrangements, such as those shown in Figure 3, are reproduced most closely in model studies involving clay, unconsolidated sand, sheets of paraffin wax, Plasticine, or wet tissue paper (Riedel, 1929; Cloos, 1955; Pavoni, 1961; Emmons, 1969; Morgenstern and Tchalenko, 1967; Tchalenko, 1970; Wilson, 1970; Lowell, 1972; Wilcox et al., 1973; Courtillot et al., 1974; Freund, 1974; Mandl et al., 1977; Graham, 1978; Groshong and Rodgers, 1978; Rixon, 1978; Gamond, 1983; Odonne and Vialon, 1983; Harris and Cobbold, 1984), in the experimental deformation of homogeneous rock samples under confining pressure (Logan et al., 1979; Bartlett et al., 1981), and in the deformation of alluvium during large earthquakes (Tchalenko, 1970; Tchalenko and Ambraseys, 1970; Clark, 1972, 1973; Sharp, 1976, 1977; Philip and Megard, 1977). Five sets of fractures are commonly observed (Fig. 5): (1) synthetic strike-slip faults or Riedel (R) shears; (2) antithetic strike-slip faults or conjugate Riedel (R') shears; (3) secondary synthetic faults or P shears; (4) extension or tension fractures (see Biddle and Christie-Blick, 1985a this volume); and (5) faults parallel to the principal displacement zone, or Y shears of Bartlett et al. (1981). In experimental deformation of Indiana limestone, Bartlett et al. (1981) have also described what they call X shears (not shown in Fig. 5). These are symmetrical with R' in relation to the principal displacement zone. The sense of offset for R and P shears is the same as that of the principal displacement zone, whereas R' and X shears have the opposite sense

of offset. The approximate orientation at which faults and associated folds develop under simple conditions is indicated in Figure 5. For the right-slip system illustrated, folds, P shears and X shears are right-handed (Campbell, 1958; Wilcox et al., 1973; Biddle and Christie-Blick, 1985a this volume), whereas R shears, R' shears and extension (tension) fractures are left-handed. As discussed below, however, geological examples tend to be more complicated, and even in the case of Holocene deformation (Fig. 7 below), observed arrangements of structures do not necessarily conform to those predicted by models or experiments. This is because rocks are heterogeneous, because structures develop sequentially rather than instantaneously, and because early-formed structures tend to be rotated during protracted deformation. The structural style is also affected by even a small component of extension or shortening across the principal displacement zone (Wilcox et al., 1973; Harding et al., 1985 this volume), a circumstance that favors fault-parallel folds rather than en echelon ones.

In addition to the occurrence of flower structures, strike-slip faults exhibit several characteristics in profile that result in part from the segmentation of wedge-shaped sediment or rock bodies with laterally variable facies, and in part from a component of convergence or divergence across different fault strands (Fig. 4). Examples are (1) the presence in a given profile across the principal displacement zone of both normal- and reverse-separation faults, and of variable proportions of normal and reverse faults in different profiles; (2) the tendency in a single profile for the magnitude and sense of separation of a given fault splay to vary from one horizon to another; and (3) the tendency in successive profiles for a given fault to dip alternately in one direction and then in the opposite direction, and to display variable separation (both magnitude and sense) for a single horizon.

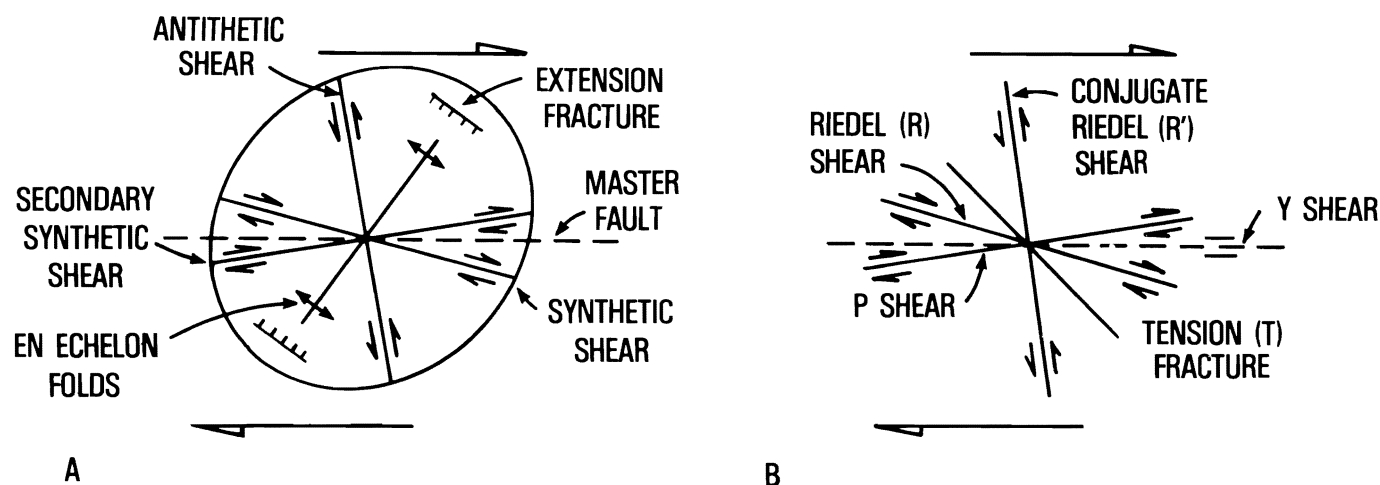


FIG. 5.—The angular relations between structures that tend to form in right-lateral simple shear under ideal conditions, compiled from clay-cake models and from geological examples. Arrangements of structures along left-slip faults may be determined by viewing the figure in reverse image. A) Terminology largely from Wilcox et al. (1973), superimposed on a strain ellipse for the overall deformation. B) Riedel shear terminology, modified from Tchalenko and Ambraseys (1970) and Bartlett et al. (1981). Extension fractures form when effective stresses are tensile (i.e., when pore-fluid pressure exceeds lithostatic pressure); tension fractures form when lithostatic loads become negative (J. T. Engelder, personal commun., 1985). In geological examples, faults with normal separation tend to develop parallel to the orientation of the extension and tension fractures in A and B; faults with reverse separation tend to develop parallel to the orientation of the fold in A. See text for further discussion of structures along strike-slip faults.

On a regional scale, distinctive aspects of strike-slip structural style are evidence for simultaneous shortening and extension, and for random directions of vergence in associated thrusts and nappes (e.g., Heward and Reading, 1980; Miall, 1985 this volume).

#### *Controls on Structural Patterns*

The main controls on the development of structural patterns along individual continental strike-slip faults are (1) the degree to which adjacent blocks either converge or diverge during strike-slip; (2) the magnitude of displacement; (3) the material properties of the sediments and rocks being deformed; and (4) the configuration of pre-existing structures. Each of these factors tends to vary along any given fault, and, except for the last, to change through time. Individual faults are commonly elements of broader regions of deformation, particularly along major intracontinental transform zones, and different faults may be active episodically at different times as a result of block rotation or the reorganization of block boundaries. Structural style may also be influenced directly by the rotation of small blocks that can produce segments of relative convergence and divergence even along straight strike-slip faults (Fig. 6), where adjacent blocks are for the most part neither converging nor diverging. The material properties of the sediments and rocks being deformed tend to change with time where, for example, strike slip is accompanied by sedimentation, so that sediments formerly near the surface become progressively more deeply buried; or where strike slip is accompanied by uplift, so that once buried strata are brought closer to the surface (as in the familiar mechanism of Karig, 1980, for raising "knockers" from great depths in accretionary prisms). Pre-existing structures can markedly influence the location and orientation of strike-slip faults, or simply complicate the overall structural pattern without being reactivated during strike-slip deformation.

*Convergent, Divergent and Simple Strike-Slip Faults.*—The terms "convergent wrench fault" and "divergent wrench fault" were introduced by Wilcox et al. (1973) to describe basement-involved strike-slip fault zones that, judging from the proportions of reverse-separation and normal-separation faults along and adjacent to the principal displacement zone, are thought to involve a significant component of shortening or extension, respectively. Convergent and divergent thus imply much the same as transpressional and transtensional of Harland (1971), although they refer to kinematics rather than to stress, and are for this reason preferred. We note that the use of these terms in the context of deformational process, rather than for description alone, presupposes that strike slip and transverse shortening or extension occurred simultaneously, not sequentially. In some cases, it may be difficult to make such distinctions. Faults along which there is no evidence for preferential convergence or divergence were described by Wilcox et al. (1973) as "simple parallel wrench faults," by Harding and Lowell (1979) as "side-by-side wrench faults," and by Mann et al. (1983) as "slip-parallel faults." Here, we use the term "simple strike-slip fault" in order to avoid confusion with a geometrical pattern in which a number of faults are actually parallel to

each other or arranged side-by-side, and because for any fault, slip is by definition parallel with the fault surface (see Biddle and Christie-Blick, 1985a this volume).

In comparison with simple strike slip, convergent strike slip leads to the development not only of abundant reverse faults, including low-angle thrust faults, but also of folds that are arranged both en echelon and parallel to the principal displacement zone (Fig. 3; Lowell, 1972; Wilcox et al., 1973; Sylvester and Smith, 1976; Lewis, 1980; and in this volume, Anadón et al., 1985; Mann, et al., 1985; Şengör et al., 1985; Steel et al., 1985). In cases of pronounced convergence, as in the western Transverse and Coast Ranges of California (WTR and CR in Fig. 1), the structural style becomes similar to that of a thrust and fold belt (Nardin and Henyey, 1978; Suppe, 1978; Yeats, 1981, 1983; Crouch et al., 1984; Davis and Lagoe, 1984; Wentworth et al., 1984; Namson et al., 1985). In contrast, along divergent strike-slip faults, folds are less well developed, and commonly consist of flexures arranged parallel rather than oblique to the principal displacement zone, although both types are known (Fig. 3; Wilcox et al., 1973; Nelson and Krausse, 1981; Harding et al., 1985 this volume).

Flower structures develop along both convergent strike-slip faults (Allen, 1957, 1965; Wilcox et al., 1973; Sylvester and Smith, 1976; Harding and Lowell, 1979; Harding et al., 1983; Harding, 1985), where they are associated with prominent antiforms and known as positive flower structures, and along divergent faults, where they are associated with synforms and termed negative (D'Onfro and Glagola, 1983; Harding, 1983, 1985; Harding et al., 1985 this volume). The occurrence of upward-diverging fault splays is thus not due essentially to convergence but to the propagation of faults upward through the sedimentary cover toward a free surface (Allen, 1965; Sylvester and Smith, 1976), an interpretation confirmed in experimental studies with unconsolidated sand (Emmons, 1969) and with homogeneous rock samples under confining pressure (Bartlett et al., 1981).

Strike-slip faults may be characterized by convergence or divergence for considerable distances (Namson et al., 1985; Harding et al., 1985 this volume), or only locally at restraining and releasing bends, fault junctions, and oversteps (Crowell, 1974b; Aydin and Nur, 1982b; Mann et al., 1985 this volume). Shortening and extension can occur simultaneously at oversteps associated with block rotation. A small-scale example is provided by a set of left-stepping en echelon R shears formed in a newly seeded carrot field during the Imperial Valley, California, earthquake of 15 October, 1979 (Fig. 6A; Terres and Sylvester, 1981). Elongate blocks of soil separated along furrows, became detached from the less rigid subsoil (a "Riedel flake" of Dewey, 1982), and rotated as much as 70°. The complex pattern of fractures, faults, folds and buckles, and gaps is related to motion on individual blocks. Şengör et al. (1985 this volume) describe an analogous rotated flake more than 70 km long, along the North Anatolian fault. Block rotation may also produce local convergence and divergence along relatively straight fault segments (Fig. 6B; Dibblee, 1977; Nicholson et al., 1985a, b). From an analysis of earthquake hypocentral locations and first-motion studies near the intersection of the San Andreas and San Jacinto faults, California, Ni-

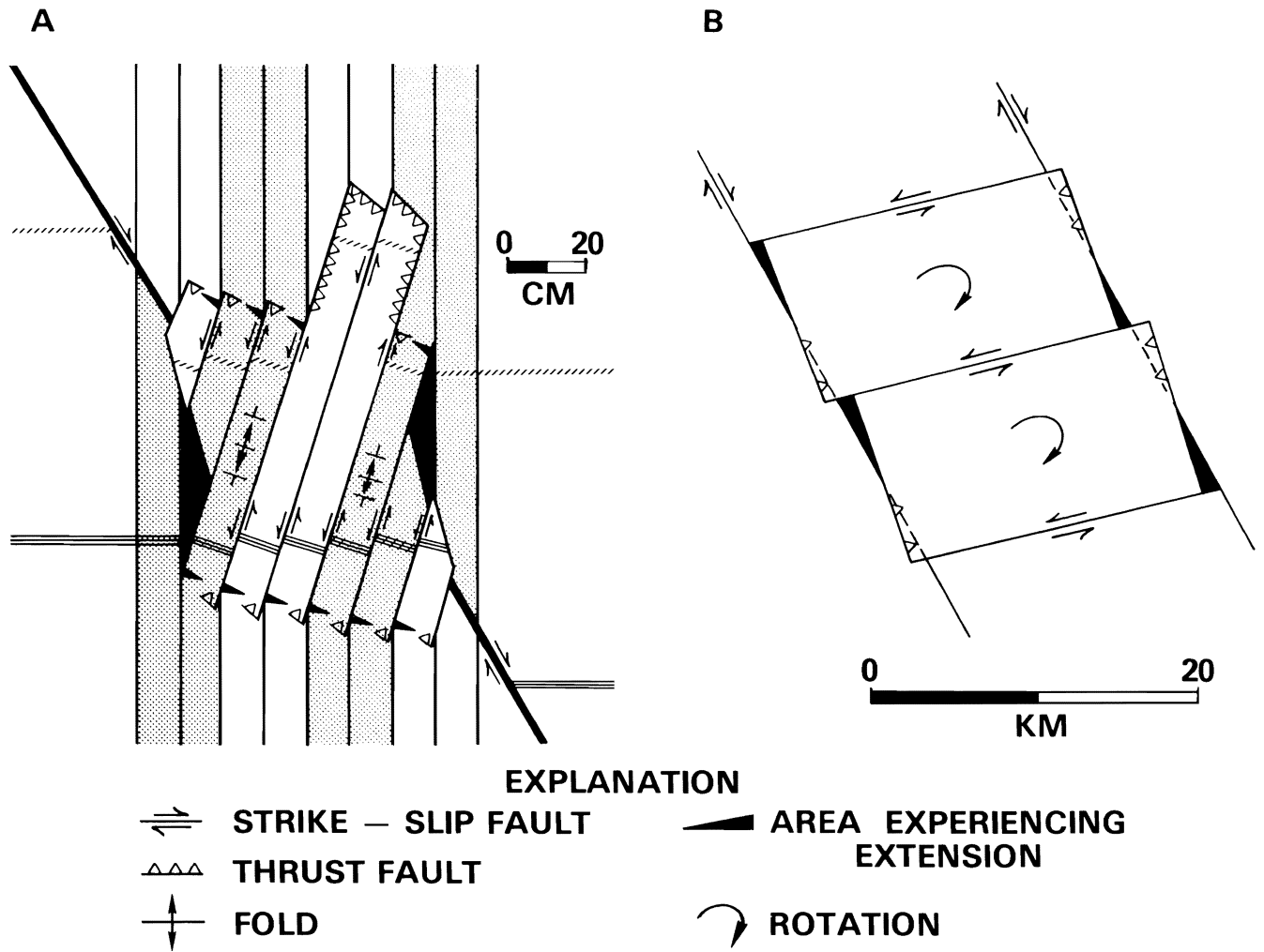


FIG. 6.—A) Geometric model illustrating the development of fractures, faults, folds and buckles, and gaps observed in a surface rupture in the Imperial Valley, California (I in Fig. 1), following the earthquake of 15 October, 1979 (from Terres and Sylvester, 1981). The spaced parallel ruling represents the furrows in a newly seeded carrot field.

B) A model for the rotation of blocks near the intersection of the San Andreas and San Jacinto faults, California, from earthquake hypocentral locations and first-motion studies (from Nicholson et al., 1985a, b; and similar to a figure of Dibblee, 1977). During a large earthquake, there is right slip on one of the major bounding faults; during the inter-seismic interval, the major faults are locked, and right-lateral motion is taken up by clockwise rotation of small blocks, and by minor left-slip faults. Incompatible motion at the corners of blocks leads to the development of normal and reverse fault segments, which alternate along the major right-slip faults. If such a non-elastic process were to continue on a geological timescale, it would produce complex wedge-shaped basins in areas subject to extension, and crustal flakes detached above thrust faults merging with the right-slip faults.

cholson et al. (1985a, b) have shown that incompatible motion at the corners of blocks leads to the development of normal- and reverse-fault segments, which alternate along the major right-slip faults.

*Displacement Magnitude.*—Experimental work, sub-surface studies and field observations indicate that structural style along strike-slip faults is influenced qualitatively by the magnitude of displacement (Wilcox et al., 1973; Harding, 1974; Harding and Lowell, 1979; Bartlett et al., 1981). In experiments, folds and shear fractures tend to form sequentially, although not in the same abundance or in the same order for all materials under all conditions (Wilcox et al., 1973; Bartlett et al., 1981). In clay models, for example, folds and Riedel shears develop first, followed by P shears, and finally by the development of a relatively

narrow through-going principal displacement zone (Morgenstern and Tchalenko, 1967; Tchalenko, 1970; Wilcox et al., 1973). In experiments with rock samples under confining pressure, R and P shears develop concurrently, and R shears tend to propagate toward the orientation of the principal displacement zone with increasing displacement (Bartlett et al., 1981). R' and X shears form at this latter stage. The simultaneous development of both R and P shears cannot be explained by the Coulomb-Mohr failure criterion. The width of the zone of deformation increases rapidly during the initial development of folds and fractures, but tends to stabilize quickly because weakening of the sheared material leads to the concentration of subsequent deformation (Odonne and Vialon, 1983). Early-formed structures tend to rotate during progressive deformation (Tchalenko, 1970;

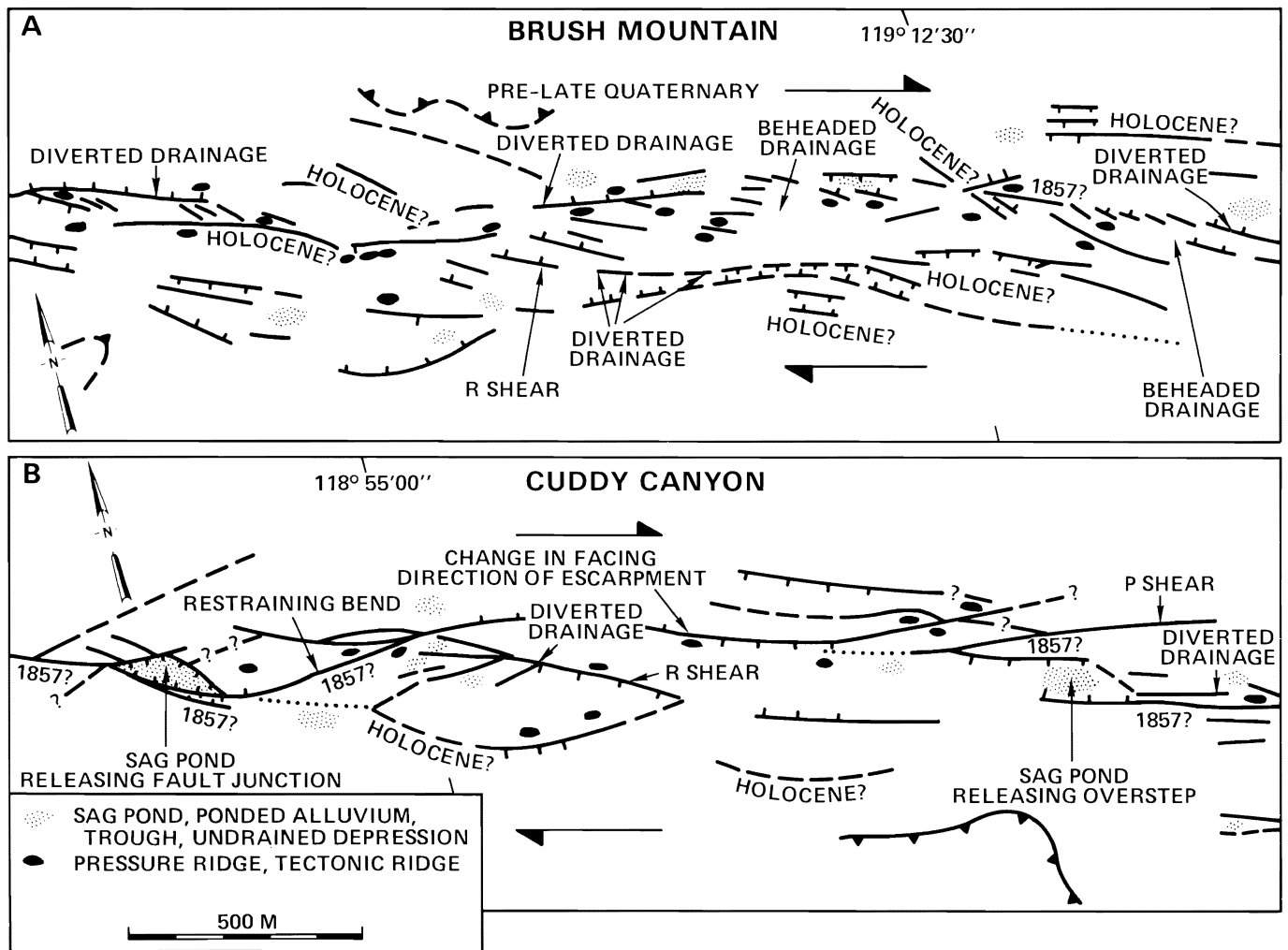


FIG. 7.—Holocene faults and tectonic landforms along the “big bend” of the San Andreas fault, California, A) south of Brush Mountain, and B) in Cuddy Canyon, east of Frazier Park (simplified from T. Davis and E. Duebendorfer, unpublished mapping, 1981, with permission from the authors; see Fig. 1 for location). Hachures on topographically lower side of fault traces; sawteeth on upper plate of thrust faults. The maps indicate the structural complexity that can develop in a few thousand years along a major strike-slip fault. The sediments at the surface are largely of Neogene and Quaternary age (A) and Quaternary age (B). The stipple pattern indicates the sites of basins (sag ponds, ponded alluvium, troughs, undrained depressions); elliptical black dots indicate elevated blocks (pressure ridges, tectonic ridges). Except for a few obvious examples of basins at releasing fault junctions or releasing oversteps, at the scale of these examples, there is no obvious relation between sites of sedimentation and fault geometry.

Wilcox et al., 1973; Rixon, 1978; Odonne and Vialon, 1983), particularly R' shears, initially oriented at a high angle to the principal displacement zone. Thus the final orientations of folds and faults depend on the magnitude of displacement and on the time at which these structures formed during the deformation history.

Many of these features documented in experiments, can also be observed in geological examples. For instance, strike-slip faults with small lateral displacement in the basement are generally expressed within the sedimentary cover by discontinuous faults and folds (e.g., the Newport-Inglewood fault zone of the Los Angeles Basin with right slip of about 200 to 750 m; Harding, 1974). At the same scale, faults with large displacement (such as the San Andreas) consist of a through-going principal displacement zone (see Fig. 5 of Harding and Lowell, 1979). One difference, however, between experiments and natural examples is that in

nature, deformation is commonly accompanied by sedimentation, so that younger sediments record less offset than older ones. Faults within Holocene alluvium along the San Andreas fault zone tend to be discontinuous (Fig. 7; from T. Davis and E. Duebendorfer, unpublished mapping, 1981). Figure 7 also illustrates the rate at which structural complexities can arise. Given an immensely longer geological timescale, and the fact that rocks and sediments are heterogeneous even before deformation, we should not be surprised if the orientation and geometry of observed structures along a given strike-slip fault depart significantly from those predicted by idealized strain ellipse summaries (Fig. 5). This is of considerable importance to the petroleum geologist constructing structure maps for a prospective exploration target from limited data. Idealized models are useful, but they should be applied with caution. That such models commonly “match” geological observations is in part due

to the large number of different fault orientations available (Fig. 5), and interpretations based on such comparisons are not necessarily correct. In addition, the absence of a simple strike-slip structural style does not eliminate the possibility that strike slip played a major role in the deformation.

*Material Properties.*—The character of a given strike-slip fault zone is also a function of the lithology of the juxtaposed rocks, the confining pressure, fluid pressure, and temperature conditions at which deformation occurred, and the rates of strain and recovery (Donath, 1970; Mandl et al., 1977; Sibson, 1977; Logan, 1979; Logan et al., 1979; Ramsay, 1980; Bartlett et al., 1981; Wise et al., 1984). These factors are likely to change during continuing strike-slip deformation as a result of lateral variations in depositional facies, or through sedimentation and burial, uplift and erosion, changes in provenance, or changes in heat flow.

Though difficult to predict, fault-zone characteristics are of significance to petroleum geologists because they influence the tendency of faults to behave as either conduits or barriers to fluid migration. Depending on the conditions of deformation, the accumulation of displacement along a fault may be accompanied by the development of gouge, which can act as a barrier to fluid migration (e.g., Pittman, 1981). However, displacement may also promote fracturing and additional avenues for leakage. As with other faults, strike-slip faults may juxtapose rocks with significantly different permeability, thus forming either seals or conduits for the migration of petroleum across the fault surface (see Downey, 1984, for a general treatment of hydrocarbon seals).

*Pre-existing Structures.*—In regions where strike-slip faults are present, pre-existing structures are of two kinds, “essential” and “incidental.” As applied to the analysis of strike-slip deformation, essential structures are defined here as those which significantly influence the location and orientation of faults and folds during strike slip. Incidental structures are inherited and contribute little to that deformation. Both types of structure are elements of the overall structural pattern of the region, but incidental ones should be excluded from the analysis strike-slip structural style.

Essential structures commonly influence patterns of strike-slip deformation on the continents. The process can be observed today in the Upper Rhine graben (Fig. 8), formed by extension in middle Eocene to early Miocene time in response to Alpine deformation (Illies, 1975; Illies and Greiner, 1978; Şengör et al., 1978). In Pliocene to Holocene time, normal faults parallel to the graben were reactivated as strike-slip faults during uplift of the Alps. Earthquake first-motion studies suggest that much of the current deformation is due to left slip (Fig. 8; Ahorner, 1975), and this sense of displacement is consistent with measurements of in-situ stress (Illies and Greiner, 1978). Other examples of essential structures are discussed in this volume by Cemen et al. (1985), Harding et al. (1985), Mann et al. (1985), Royden (1985), Şengör et al. (1985), and Zalan et al. (1985). In the oceans, the location of some major transform faults is controlled by weaknesses within the continents prior to continental separation (e.g., the Atlantic equatorial megashear zone; Bonatti and Crane, 1984; Zalan et al., 1985 this volume).

Some basement-involved folds in the Death Valley area,

southeastern California, illustrate the concept of incidental pre-existing structures (Figs. 1, 9). Central Death Valley developed in late Miocene through Holocene time as an east-tilted half graben bounded to the north and south by northwest-striking right-slip faults (the northern Death Valley–Furnace Creek fault zone and the southern Death Valley fault zone; Stewart, 1983). Although the type example of a pull-apart basin (Fig. 9B; Burchfiel and Stewart, 1966; Mann et al., 1983), Death Valley is not a particularly good example, because it is only one of many grabens in the Basin and Range Province that formed in response to late Cenozoic regional extension, not as a result of a bend or an overstep in a strike-slip fault (see reviews of the Basin and Range Province by Stewart, 1978; Eaton, 1982; and Fig. 1 of Cemen et al., 1985 this volume). The strike-slip faults appear to be tear faults within an extensional allochthon separating areas that experienced different amounts of extension (Wright and Troxel, 1970; Wernicke et al., 1982; Stewart, 1983). There is no evidence for strike slip within and parallel to the *central* segment of Death Valley (Wright and Troxel, 1970), although the basin undoubtedly opened obliquely, as shown in Figure 9B, and is at present extending in an approximately northwest-southeast direction (Sbar, 1982). Such a strike-slip fault was shown on an earlier map by Hill and Troxel (1966; trend S<sub>1</sub> in the strain ellipse insert of Fig. 9A), and reproduced by Mann et al. (1983). Evidence cited to support that interpretation, in addition to the presence of oblique striae on fault surfaces in the Black Mountains, consists of spectacular northwest-plunging folds within the basement (trend A in Fig. 9A). Hill and Troxel (1966) described the folds as en echelon, but regional arguments, including evidence for the pressure and temperature conditions at the time of folding, suggest that they are probably incidental oblique structures predating strike-slip deformation. Folds with similar orientation are present in metamorphic rocks of both the Panamint Range, west of central Death Valley, and in the Funeral Mountains, north of (and approximately parallel to) the northern Death Valley–Furnace Creek fault zone. Metamorphism in the Funeral Mountains occurred, probably in late Mesozoic time, at a temperature of 600° to 700° C, and at a pressure of 7.2 to 9.6 kb (Labotka, 1980). In the Panamint Range, folding occurred at 70 to 80 Ma during retrograde metamorphism, and at a temperature of about 450° C, following prograde metamorphism at 400° to 700° C (Labotka, 1981; Labotka and Warasila, 1983). These metamorphic conditions and the age of folding inferred in the Panamint Range are incompatible with the Miocene and younger tectonic denudation that accompanied strike-slip deformation in the Death Valley area (see Cemen et al., 1985 this volume). Although low-angle normal faults follow the top of the basement, the folds themselves are incidental structures.

#### BASINS ALONG STRIKE-SLIP FAULTS

Some basins along strike-slip faults developed as a direct response to strike-slip deformation. Others owe their origin to a different tectonic regime, in which strike slip played only a subsidiary role in basin development. Examples of

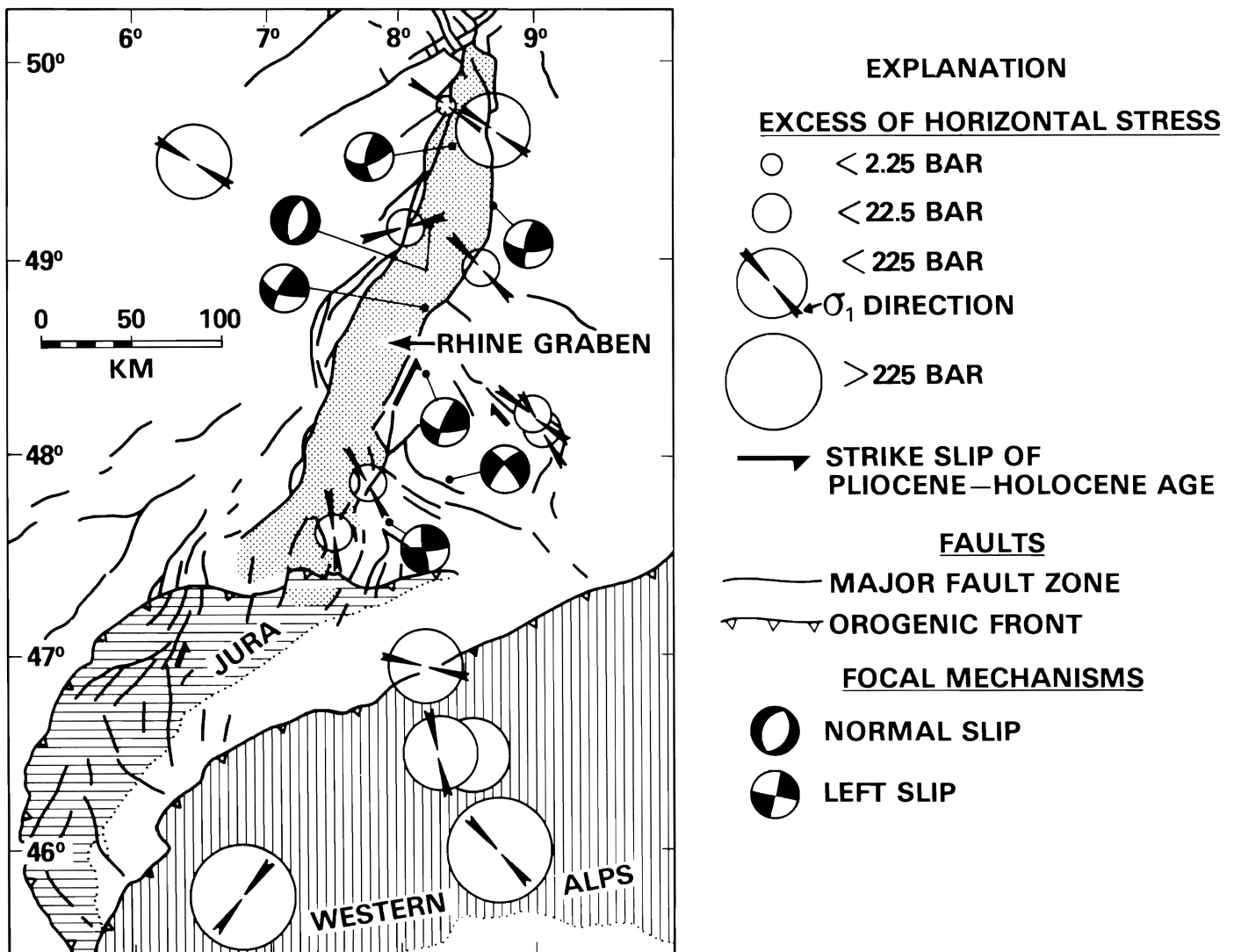


FIG. 8.—The Rhine graben and adjacent segments of the Jura and western Alps; with selected examples of the excess horizontal stress from in situ determinations; and earthquake focal mechanisms (modified from Ahorner, 1975; Illies and Greiner, 1978). Normal faults of Eocene to Miocene age were reactivated as left-slip faults in Pliocene to Holocene time, during uplift of the Alps.

the latter are (1) some intracontinental grabens, foreland basins, and forearc basins associated with strike-slip faults, which though active during sedimentation, had little influence on sedimentation patterns; and (2) various basins that were subsequently reactivated by strike-slip deformation. Basins of these types are considered here only to the extent that the geology is related to strike-slip deformation.

One of the most obvious characteristics of a strike-slip basin is its present geometry in map view, particularly the geometry of any bounding faults, and this has been a useful point of departure for various classification schemes and for most discussions about processes of basin formation (Carey, 1958; Kingma, 1958; Lensen, 1958; Quennell, 1958; Burchfiel and Stewart, 1966; Clayton, 1966; Belt, 1968; Freund, 1971; Crowell, 1974a, b, 1976; Ballance, 1980; Aydin and Nur, 1982a, b; Burke et al., 1982; Crowell and Link, 1982; Fralick, 1982; Mann and Burke, 1982; Mann et al., 1983; Mann and Bradley, 1984; see glossary in Bid-

dle and Christie-Blick, 1985a this volume). However, with the exception of very young sedimentary accumulations, the geometry, location, and perhaps orientation of a given basin have undoubtedly changed with time, and some prominent faults may be younger than much of the sedimentary fill. In many cases, present geometry may thus provide only limited information about either the kinematic history or the ultimate controls on basin evolution, and the utility of some of the classification schemes is questionable.

In this section, we focus on processes of basin formation, on the tectonic setting of depositional sites, and on certain distinctive characteristics of the stratigraphic record along strike-slip faults. We emphasize as others have before (Bally and Snelson, 1980) that every basin has a unique history, and that simple models may provide only a superficial summary of basin development. No strike-slip basin can be considered thoroughly understood unless its history can be reconstructed by a series of well-constrained palinspastic

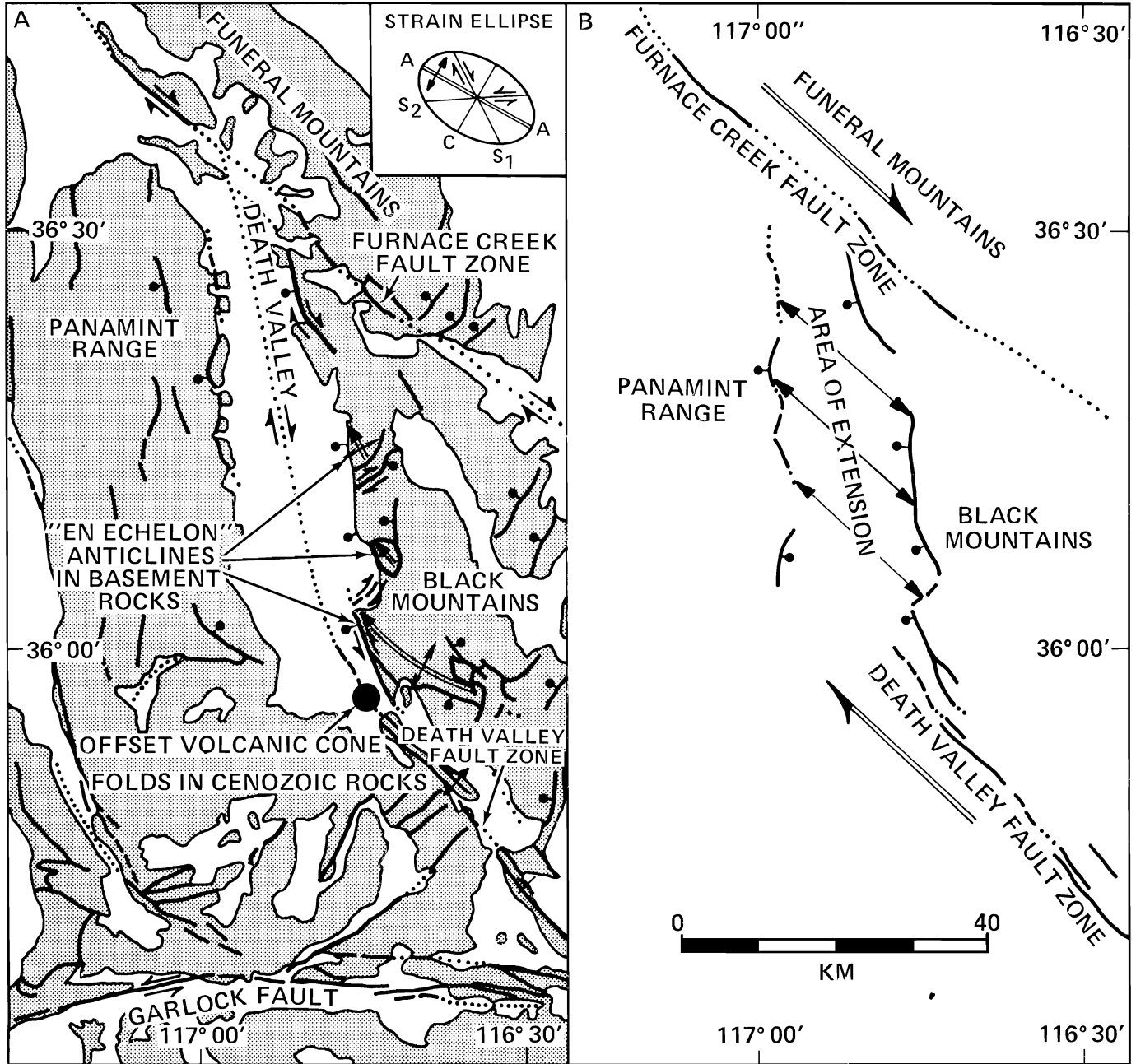


FIG. 9.—Major fault zones in the Death Valley area, California (modified from Stewart, 1983; see Fig. 1 for location), showing the interpretations of A) Hill and Troxel (1966), and B) Burchfiel and Stewart (1966). Shading (A) indicates outcrops of Proterozoic to Tertiary sedimentary and volcanic rocks; unshaded area represents Quaternary alluvial deposits. Evidence for strike slip along the northern Death Valley-Furnace Creek and southern Death Valley fault zones includes en echelon folds in Cenozoic rocks and an offset volcanic cone, together with regional stratigraphic arguments.

A) A buried strike-slip fault is inferred in the central north-trending segment of Death Valley on the basis of oblique striae on fault surfaces in the Black Mountains, and of "en echelon" anticlines in basement rocks (Hill and Troxel, 1966). The insert compares the orientations of observed structures with an idealized strain ellipse for the overall deformation; right slip inferred parallel to direction C is incompatible with orientations summarized in Figure 5.

B) Death Valley interpreted as a pull-apart along an oblique segment of a strike-slip fault system (Burchfiel and Stewart, 1966). Indicators of crustal stress and regional seismicity indicate continued extension in an approximately northwest-southeast direction parallel with the Furnace Creek and southern Death Valley fault zones (Sbar, 1982). See text for further explanation.

maps. Many of the examples discussed are taken from California, a classic region with which we are most familiar, but where appropriate, we draw attention to articles in this volume dealing with strike-slip basins in other parts of the world.

### *Processes of Basin Formation*

Sedimentary basins form along strike-slip faults as a result of localized crustal extension, and especially in zones of continental convergence, of localized crustal shortening. Individual basins vary greatly in size (e.g., Aydin and Nur, 1982a; Mann et al., 1983), but they tend to be smaller than those produced by regional extension (many intracontinental grabens) or regional shortening (foreland and forearc basins). In addition, a given basin may alternately experience both extension and shortening on a timescale of thousands to millions of years, through variations in the motion of adjacent crustal blocks (e.g., the Ventura Basin, California, Fig. 15 below; and in this volume, Miall, 1985; Nilsen and McLaughlin, 1985; Steel et al., 1985); or extension in one direction (or in one part of the basin) may be accompanied by shortening in another direction (or in another part of the basin). A possible example of the latter is the occurrence of positive flower structures in the Mecca Hills on the northeastern side of the Salton Trough, southern California, a basin that can be related to extension in a northwestward direction in the overstep between the San Andreas and Imperial faults (Fig. 1; Sylvester and Smith, 1976; Crowell, 1981b; Fig. 6 of Harding et al., 1985 this volume). The directions of extension and shortening also tend to vary within a given basin and to change through time, especially where crustal blocks rotate (as seen on a small scale in Fig. 6 A), where there are significant differences in the rate of internal strain of adjacent crustal blocks (see Figs. 10, 11, and 12 of Şengör et al., 1985 this volume), or when there are changes in lithospheric plate motion.

In the simplest strike-slip basins (e.g., the "pull-apart hole" and "sharp pull-apart basin" of Crowell, 1974a, b), the bounding blocks are torsionally rigid and deform only at their edges, and subsidence is due to extension only in a direction parallel to the regional strike of the fault(s). Examples include many small pull-apart basins (Aydin and Nur, 1982a; Mann et al., 1983), and some detached or thin-skinned basins such as the Vienna Basin (Fig. 10B; Royden, 1985 this volume). In the vicinity of the Salton Trough, however, the crustal structure inferred from seismic refraction and gravity data (Fuis et al., 1984) suggests that significant crustal thinning has occurred outside the overstep between the San Andreas and Imperial faults (Fig. 10A). The interpretation of the gravity data is not unique, but the relatively flat gravity profile across the Salton Trough requires that the upper surface of the subbasement (lined area) largely mirror the contact between the sedimentary rocks and basement (Fuis et al., 1984). The subbasement was modelled with a density of  $3.1 \text{ gm/cm}^3$ , and refraction data suggest a P-wave velocity of greater than or equal to  $7.2 \text{ km/s}$ , consistent with significant amounts of mafic igneous rocks intruded into the lower part of the crust. The depth to the top of the subbasement decreases abruptly from 16

km at the Salton Sea to approximately 10 km at the U.S.-Mexico border, about 30 km to the south, indicating that the gross crustal structure is for the most part related to the opening of the Gulf of California, and is not simply inherited from an earlier phase of regional extension.

The profile shown in Figure 10A is probably characteristic of junctures between continental transform systems and divergent plate boundaries, but some strike-slip basins are detached. Examples are known from (1) areas of pronounced regional shortening, such as the Vienna Basin, which formed adjacent to tear faults of the Carpathian nappes (Fig. 10B; Royden, 1985 this volume), and the St. George Basin, located in the forearc of the Bering Sea, Alaska (Marlow and Cooper, 1980); and (2) areas subject to marked regional extension, such as the West Anatolian extensional province of Turkey (Fig. 18 of Şengör et al., 1985 this volume), and the Basin and Range Province of the western United States (Cemen et al., 1985; Link et al., 1985, both in this volume). Detached strike-slip basins may also prove to be relatively common in intracontinental transform zones, particularly those located along former convergent plate boundaries. In central and southern California, for example, the presence of mid-crustal detachments is suggested by (1) earthquakes with low-angle nodal planes, and the alignment of earthquake hypocenters parallel to the gently dipping base of the seismogenic crust (Nicholson et al., 1985a, b; Webb and Kanamori, 1985); (2) a pronounced change, near the junction of the San Andreas and San Jacinto faults (SJ in Fig. 1), in the patterns of seismicity with depth, with shallow seismicity suggesting the rotation of small crustal blocks (Fig. 6B; Nicholson et al., 1985a, b); (3) regional patterns of earthquake travel-time residuals (Hadley and Kanamori, 1977); (4) the distribution of upper crustal seismic velocities (Hearn and Clayton, 1984; Nicholson et al., 1985a); (5) geodetic constraints on the effective elastic thickness of the upper crust (Turcotte et al., 1984); (6) paleomagnetically determined patterns of large-scale block rotation (Fig. 2A; Luyendyk et al., 1985); (7) deep seismic-reflection studies that reveal the presence of low-angle reflecting horizons (Cheadle et al., 1985); and (8) evidence from surface geology and shallow seismic reflection profiles for numerous Neogene and Quaternary low-angle faults (Yeats, 1981, 1983; Crouch et al., 1984; Wentworth et al., 1984; Namson et al., 1985). We note that it is not yet clear how much of the California margin is detached, although the scale probably exceeds that of individual basins. Manspeizer (1985 this volume) interprets the Dead Sea Basin as detached on a smaller scale above listric normal faults bounded by the overstepping strands of the Dead Sea fault zone (his Fig. 13, after a concept of K. Arbenz, and reproduced on the cover of this book).

As discussed above, lithospheric or crustal extension produces a thermal anomaly, and subsequent cooling leads to additional subsidence (McKenzie, 1978; Royden and Keen, 1980; Steckler and Watts, 1982). In narrow basins, and in those for which extension is not "instantaneous," but occurs over an interval of more than about 10 m.y., a significant fraction of the thermal anomaly decays during the rifting stage, increasing the amount of syn-rift subsidence at the expense of the post-rift (Jarvis and McKenzie, 1980;



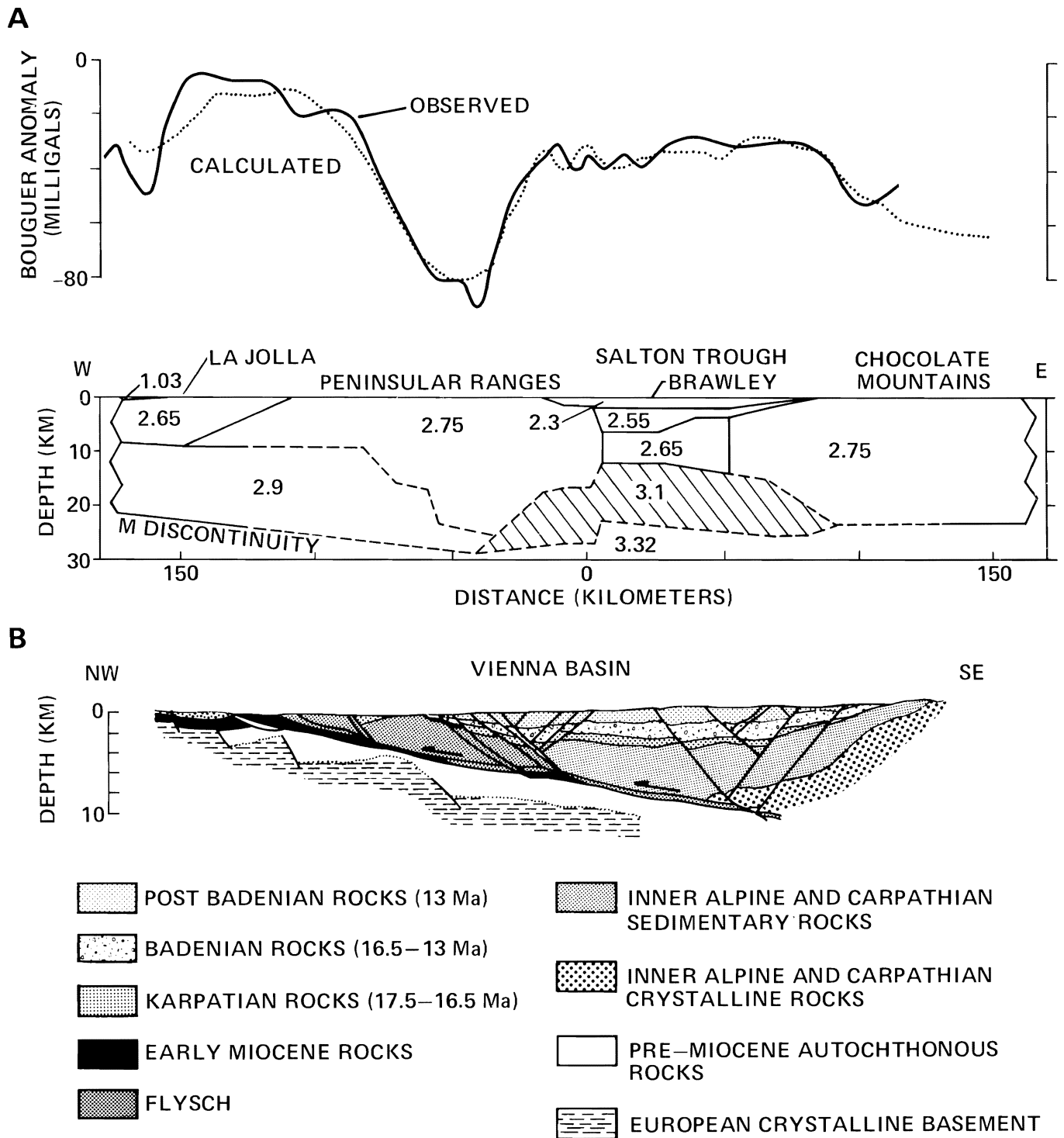


FIG. 10.—A comparison of thick-skinned and thin-skinned pull-apart basins: the Salton Trough, California (A), and the Vienna Basin, Austria and Czechoslovakia (B).

A) Cross section of southern California from La Jolla to the Chocolate Mountains (from Fuis et al., 1984, see Fig. 1 for location). The observed gravity anomaly is compared with the anomaly calculated from the model (densities in  $\text{gm}/\text{cm}^3$ ). Solid boundaries are those controlled by seismic refraction data; dashed lines indicate boundaries adjusted to fit the gravity data. Sub-basement (lined area, density  $3.1 \text{ gm}/\text{cm}^3$ ) beneath the Salton Trough provides most of the gravitational compensation for sedimentary rocks (densities  $2.3$  and  $2.55 \text{ gm}/\text{cm}^3$ ) and inferred metasedimentary rocks (density  $2.65 \text{ gm}/\text{cm}^3$ ). The San Andreas and Imperial faults are located near the east and west edges of the block with density of  $2.65 \text{ gm}/\text{cm}^3$ .

B) Cross section of the Miocene Vienna Basin, a detached pull-apart basin superimposed partly on nappes of the outer Carpathian flysch belt, and partly on nappes of the inner Carpathians (section 3 from Fig. 6 of Royden, 1985 this volume). Tertiary thrusts are indicated by arrows; Miocene normal faults displace Miocene rocks at the surface; normal faults confined to the autochthon are mainly Jurassic syn-sedimentary faults associated with Mesozoic rifting.

Steckler, 1981; Cochran, 1983; Pitman and Andrews, 1985 this volume). The post-rift thermal anomaly is also less for detached basins than for those produced by an equivalent amount of extension of the entire lithosphere, and detached basins thus experience reduced post-rift subsidence. Extensional strike-slip basins tend to be short-lived, but the rifting stage is preferentially represented in the stratigraphic record because most such basins are small, in some cases detached, and in many cases subject to subsequent uplift and erosion.

Another potentially important mechanism for basin subsidence along strike-slip faults, in addition to crustal extension, is loading due to the local convergence of crustal blocks (e.g., the Ventura Basin of southern California; Fig. 15 below; Burke et al., 1982; Yeats, 1983). Similar basins bounded by major thrust faults are discussed by Şengör et al. (1985 this volume) and by Steel et al. (1985 this volume). As in foreland basins (Beaumont, 1981), patterns of subsidence are likely to be influenced by lithospheric flexure, a mechanism of regional isostatic compensation in which loads are supported by broad deflection of the lithosphere as a result of lithospheric rigidity (Forsyth, 1979; Watts, 1983), but to our knowledge, no attempt has yet been made to model flexure in strike-slip basins. Some strike-slip basins experience both extension and shortening (e.g., a pull-apart basin adjacent to a convergent strike-slip fault). Flexural effects are reduced when the lithosphere is warm and relatively weak, and we presume that in such circumstances subsidence induced by loading would be localized.

Subsidence due to extension, thermal contraction or crustal loads is amplified by sediment loading (Steckler and Watts, 1982). Simple isostatic considerations indicate that for typical crustal and sediment densities, sediment loading accounts for about half the total subsidence. Loading by water is important only to the extent that sea level or lake level varies.

Although it is relatively simple to construct models of the processes outlined above for hypothetical basins, it is clearly difficult to unravel the processes from the geological record of a given strike-slip basin. A fruitful avenue of research will be to use a combination of geological and geophysical data from well studied basins to test these theoretical concepts.

#### *Tectonic Setting of Depositional Sites*

The location of depositional sites along strike-slip faults is controlled by several factors operating at a variety of length scales and time scales. Along intracontinental transform zones, such as in California, and the Dead Sea fault zone, these include (1) crustal type and the configuration of pre-existing crustal structures, especially the distribution, orientation and dimensions of zones of weakness, such as faults; (2) variations in the overall motion of adjacent lithospheric plates; and (3) the kinematic behavior of crustal blocks within the transform zone.

*Pre-existing Crustal Structure.*—Pre-existing crustal structure influences the location of major strike-slip faults, and the way in which crustal blocks between those faults move and deform to produce basins. The San Andreas fault, for

example, originated in southern California by about 8 Ma, with most of the 320 km of right slip accumulating since about 5.5 Ma during oblique opening of the Gulf of California (Fig. 1; Karig and Jansky, 1972; Crowell, 1981b; Curray and Moore, 1984). Strike-slip faulting is only the latest event in a long geologic history, extending back into Proterozoic time, and including the assembly, amalgamation and accretion of various suspect terranes to North America during Mesozoic time; Mesozoic to Miocene arc magmatism, associated forearc and perhaps back-arc sedimentation; and Oligocene to Miocene Basin and Range extension (Hamilton, 1978; Crowell, 1981a, b; Dokka and Merriam, 1982; Champion et al., 1984). The present Gulf is the result of seafloor spreading beginning in latest Miocene time, but possibly along the line of a somewhat older Miocene tectonic depression that has been termed the proto-Gulf of California (Karig and Jansky, 1972; Terres and Crowell, 1979; Crowell, 1981a, b; but see Curray and Moore, 1984, for a modified interpretation). With local exceptions, such as in the vicinity of the Transverse Ranges (ETR and WTR in Fig. 1), the San Andreas fault and related transform faults to the south in the Gulf of California approximately parallel the boundaries of older tectonic elements, and occupy a position along the eastern edge of the Cretaceous Peninsular Ranges and Baja California Batholith (Hamilton, 1978).

Vink et al. (1984) suggested that the Gulf of California and the associated transform system formed just within the continent because continental lithosphere is weaker than oceanic lithosphere, and that rifting close to a continent-ocean boundary invariably follows a continental pathway. A similar model has been proposed by Steckler and ten Brink (1985) to explain the development in mid-Miocene time of the Dead Sea fault, at the expense of extension in the Gulf of Suez. According to their model, the hinge zone along the Mediterranean margin between thinned and unthinned continental crust acted as a barrier to continued northward propagation of the Red Sea rift.

The distribution of Neogene basins in California is probably influenced to a certain extent by crustal composition and thickness, and by the location of older basins (Blake et al., 1978). For example, the San Joaquin Basin (SJB in Fig. 1), located along the eastern side of the San Andreas Fault in central California, is superimposed on a forearc basin of late Jurassic to mid-Miocene age, which in turn overlies the boundary between Sierran-arc and ophiolitic basement (Blake et al., 1978; Hamilton, 1978; Bartow, 1984; Bartow and McDougall, 1984; Namson et al., 1985).

*Relative Plate Motion.*—Another factor that influences the development of sedimentary basins along a continental transform zone is the overall relative motion of the adjacent lithospheric plates (Crowell, 1974b; Mann et al., 1983). Such motion may be strictly transform, or it may involve a component of either convergence or divergence (Harland, 1971). In practice, however, it is generally difficult to predict the behavior of individual blocks solely on this basis, owing to the complex jostling that takes place in a broad transform plate boundary, even when there is little variation in relative plate movement. The character of the plate boundary is also sensitive to the migration of any unstable

triple junctions, and to changes in the position of the instantaneous relative-motion pole. Either of these eventualities may be associated with the reorganization of plate-boundary geometry, and with the transfer of slices of one plate to another (Crowell, 1979).

Mann et al. (1983) argued that the gross distribution of regions of extension and shortening along the San Andreas and Dead Sea transform systems can be explained by comparing the orientation of the major faults with the interplate slip lines suggested by Minster et al. (1974). According to Mann et al., active strike-slip basins occur preferentially where the principal displacement zone is divergent with respect to overall plate motion (e.g., the Wagner and Delfin Basins of the northern Gulf of California; and the Dead Sea Basin); push-up blocks (the Transverse Ranges, California, and the Lebanon Ranges) occur where the principal displacement zone is convergent. Such statements are probably valid where most of the plate motion is taken up along a single fault (e.g., Dead Sea fault zone), but less certain where the motion is accommodated by a number of major faults, and especially where crustal blocks rotate or deform internally.

Estimates of relative motion between the Pacific and North American plates, and of Quaternary slip rates for strike-slip faults in California suggest that there should be pronounced shortening across the Transverse Ranges of California (ETR and WTR in Fig. 1), an observation qualitatively supported by abundant geological observations (Nardin and Henyey, 1978; Jackson and Yeats, 1982; Yeats, 1983; Crouch et al., 1984). About half to two thirds of the Pacific-North American relative motion (56 mm/yr; Minster et al., 1974; Minster and Jordan, 1978, 1984) occurs on the San Andreas fault (Sieh and Jahns, 1984; Weldon and Humphreys, 1985). Recent estimates of the slip rate are 25 mm/yr during the Quaternary in southern California (Weldon and Sieh, 1985), and 34 mm/yr during the Holocene in central California (Sieh and Jahns, 1984). The remaining displacement is thought to be taken up by the San Jacinto and Elsinore faults (about 10 and 1 mm/yr, respectively, for Quaternary time; Sharp, 1981; Ziony and Yerkes, 1984), and by offshore faults such as the San Gregorio-Hosgri (6 to 13 mm/yr for the late Pleistocene and Holocene; Weber and Lajoie, 1977; fault not shown in Fig. 1), assuming that Quaternary plate motion has been much the same as the average motion for the past 5 to 10 m.y. In assessing these data, Weldon and Humphreys (1985) have concluded that the magnitude of Quaternary shortening across the Transverse Ranges is considerably smaller than expected. They suggest that plate motion may have been taken up not merely by translation on the major faults but in part by counterclockwise rotation about a pole 650 km southwest of the "big bend" of the San Andreas fault (Fig. 1), an interpretation that contrasts with paleomagnetic evidence for predominantly clockwise rotations west of the San Andreas on a longer time scale (Fig. 2A; Luyendyk et al., 1985). The rates of both strike-slip faulting and regional shortening are uncertain (see Bird and Rosenstock, 1984, for a different model), but available data indicate only a qualitative relation between the orientation of interplate slip lines and patterns of uplift and subsidence, even for tectonic features as prominent as the

Transverse Ranges. In zones of continental convergence, such as Turkey (Şengör et al., 1985 this volume), patterns of deformation are even more complicated.

The Neogene history of California also provides examples of plate boundary reorganization resulting from the migration of unstable triple junctions and changes in the position of the instantaneous relative-motion pole. The transform plate margin originated in California at about 30 Ma, following the impingement of the Pacific plate against the North American plate (Atwater, 1970; Atwater and Molnar, 1973). The transform system lengthened, probably intermittently, by northward motion of the northern trench-transform-triple junction and by generally southward motion of the southern transform-trench-ridge triple junction. At the same time, perhaps because of irregularities along the plate boundary (Crowell, 1979), slices of the North American plate were incorporated within the evolving transform system and offset differentially along the plate margin (Crouch, 1979). By about 5.5 Ma, the southern triple junction reached the mouth of the proto-Gulf of California, and much of the transform motion was taken up by the San Andreas fault, effectively transferring the Peninsular Ranges and Baja California to the Pacific plate (Curry and Moore, 1984). This reorganization of the plate boundary may be related to a small change in the relative plate motion determined by Page and Engebretson (1984). Another consequence of the inferred change in plate motion may be the onset in Pliocene and Pleistocene time of the shortening, described above, across the Transverse Ranges and Coast Ranges of California, and an acceleration in the subsidence rates for several basins (Yeats, 1978). In the Ventura Basin, for example, where considerable geochronological precision is possible, subsidence rates are estimated to have increased from approximately 250 m/m.y. at about 4 Ma to between 2,000 and 4,000 m/m.y. in the past million years. Note that these figures of Yeats (1978) incorporate inferred changes in water depth, but no corrections for compaction, loading or eustatic changes in sea level.

A modern analogue for the geological complexities at a trench-transform-triple junction, and a mechanism by which tectonic slices are transferred from one plate to another, has been described by Herd (1978). In the vicinity of the Cape Mendocino triple junction, the San Andreas fault appears to split into two subparallel fault zones, approximately 70 to 100 km apart. The western zone, the San Andreas proper, terminates at Cape Mendocino (CM in Fig. 1). The eastern zone, consisting of the right-stepping Hayward, Healdsburg-Rodgers Creek, Maacama and Lake Mountain fault zones (H, RC, M, and LM in Fig. 1), appears to be very youthful, and extends northward onto the continental shelf about 150 km north of Cape Mendocino.

*Kinematic Behavior of Crustal Blocks.*—The kinematic behavior of fault-bounded crustal blocks within a transform zone has long been considered the principal control on the development of strike-slip basins (Lensen, 1958; Kingma, 1958; Quennell, 1958; Crowell, 1974a, b, 1976; Reading, 1980; Aydin and Nur, 1982a; Mann et al., 1983). The general idea is that subsidence tends to occur where strike slip is accompanied by a component of divergence, as a result,

for example, of a bend or an overstep in the fault trace ("pull-apart basin" of Burchfiel and Stewart, 1966) or through extension near a fault junction ("fault-wedge basin" of Crowell, 1974b). Uplift occurs where there is a component of convergence, although an overriden block may be depressed by the overriding one. Examples of strike-slip basins with different geometry and of different size are illustrated in Figure 11. Crowell (1974b) described bends associated predominantly with stretching and subsidence or with shortening and uplift as "releasing" and "restraining" bends, respectively, and here, we extend the use of the terms

"releasing" and "restraining" to kinematically equivalent oversteps and fault junctions. As recognized by Crowell (1974a, b), the kinematic behavior of individual crustal blocks is superimposed on broader patterns of plate interaction, and in complexly braided fault systems, many basins experience episodic subsidence as a result of changes in fault geometry and/or block motion. In some cases, therefore, the stratigraphic record of a strike-slip basin may not be related in a simple way to the present configuration of faults, and it may even be difficult to predict contemporary patterns of subsidence and uplift from fault geometry.

Consider, for example, basins and horsts associated with fault junctions. Some of the numerous possible patterns of subsidence and uplift at a simple fault junction are illustrated in Figure 12, modified from Figure 11 of Crowell (1974b). As originally discussed by Crowell, and reproduced in subsequent summary articles (e.g., Bally and Snelson, 1980; Reading, 1980, 1982; Freund, 1982), the wedge between the faults shown in Figure 12A is said to be "compressed and elevated" where the "faults converge," and "extended" where the "faults diverge." It is clear, however, that faults that converge in one direction diverge in the opposite direction. Assuming deformation only along the edges of blocks, and rotation only to the extent required by fault curvature, uplift and subsidence are related to the orientation (dip and strike) of the faults with respect to the overall slip vectors of the blocks (horizontal in Fig. 12B), and to the amount of extension or shortening associated with each fault. Although distinct in cross section, note that similar map-view configurations of basins and horsts can arise at either releasing or restraining junctions. More complicated arrangements can be envisaged at junctions with both releasing and restraining characteristics, and where blocks are internally deformed or rotated (see Fig. 6). Segmentation of the block between the branching faults may produce grabens within a horst or horsts within a graben, as in the Dead Sea fault zone (Garfunkel, 1981). In general, we expect basins and horsts to evolve continuously by a combination of fault slip and rotations about both vertical and horizontal axes, but there are also discontinuous changes due to episodic slip and rotation, and to the propagation of new faults. These are needed from time to time to eliminate tectonic knots, or complexities that inhibit further strike-slip deformation. Thus basins with thick sedimentary accumulations may at times experience uplift, and horsts with little preserved sediment may subside. Many strike-slip basins are actually slices of basins offset along one or more younger strike-slip faults.

An important aspect of strike-slip deformation and basin formation, but one for which documentation is acquired with difficulty, is the manner in which complexities such as fault bends, oversteps and junctions arise or are subsequently removed. Some curvature in strike-slip faults is attributable to a near pole of plate rotation (e.g., the Dead Sea fault zone; Garfunkel, 1981), and some bends may result from crustal heterogeneity or from local variations in the distribution of stress influencing the path of fault propagation. In eastern Jamaica, for example, a prominent right-hand restraining bend between the left-lateral east-striking Plan-

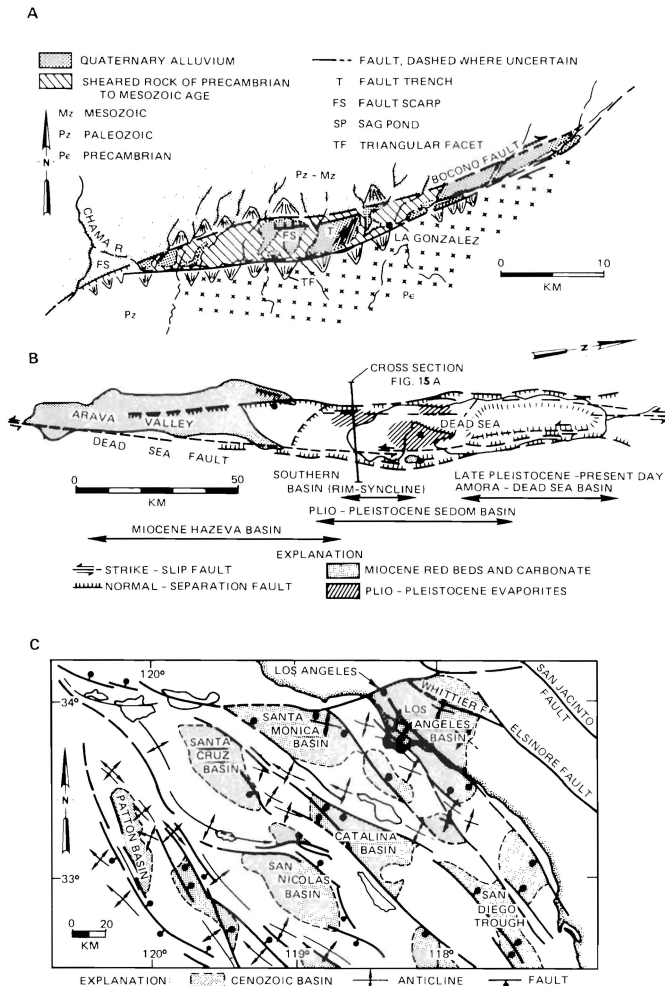


FIG. 11.—A comparison in map view of strike-slip basins of different ages, geometry, and scale.

A) The La González Basin, a lazy Z-shaped pull-apart basin of Pliocene (?) to Quaternary age along the Boconó fault zone, Venezuela (from Schubert, 1980).

B) The Dead Sea Rift, a rhomboidal pull-apart basin of Miocene to Holocene age (from Zak and Freund, 1981). Since Miocene time, the depocenter has migrated northward from the Arava Valley to the site of the present Dead Sea. For cross section, see Figure 15 A.

C) Selected faults, anticlines, and Cenozoic strike-slip basins in the southern California borderland (modified from Moore, 1969; Junger, 1976; Howell et al., 1980; see Figs. 1 and 2A for location). Many of these basins differ from the La González and Dead Sea Basins in being bounded by strike-slip faults of different orientations in a broad transform zone.

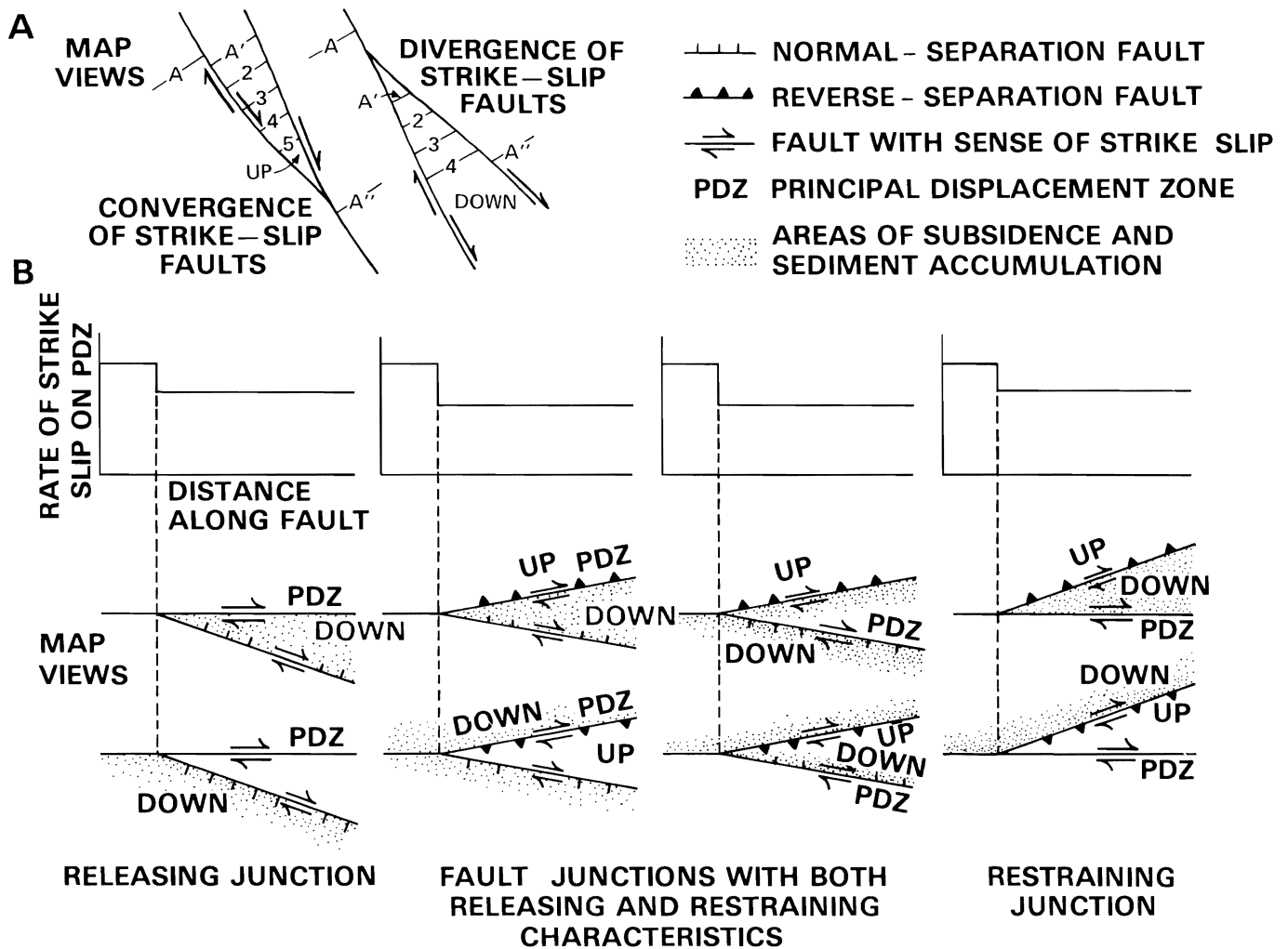


FIG. 12.—A) Sketch maps showing uplift of the tip of a fault wedge with convergence of right-slip faults, and subsidence of the tip with divergence (from Crowell, 1974b).

B) An alternative version of A, showing a range of possible results of slip along a bifurcating right-slip fault. See text for further explanation.

tain Garden and Duanvale fault zones appears to have nucleated in Miocene time on northwest-striking normal faults that bounded a Paleogene graben (Fig. 2 of Mann et al., 1985 this volume). Other bends in strike-slip faults are due to the deformation of initially straight faults, as a result of (1) incompatible slip at a fault junction; (2) rotations within one or more adjacent blocks; or (3) intersection of a strike-slip fault with a zone of greater extensional or convergent strain. The development of the “big bend” of the San Andreas fault (Fig. 1) was attributed by Bohannon and Howell (1982) to incompatible displacement in late Cenozoic time on the San Andreas and Garlock faults (320 km of right slip, and 65 km of left slip, respectively; Smith, 1962; Smith and Ketner, 1970; Crowell, 1981b). As discussed above, this deformation was accompanied by counterclockwise rotation of the Mojave block between the faults (Fig. 2A; Morton and Hillhouse, 1985). Bends may also develop as a result of small-scale block rotation of the sort documented by Nicholson et al. (1985a, b; Fig. 6B). Şengör et al. (1985 this volume) describe fault bends due to the intersection of

strike-slip faults with zones of accentuated extensional strain at the western end of the North Anatolian fault, and of accentuated convergent strain near the junction of the North Anatolian and East Anatolian faults in eastern Turkey (see their Figs. 10, 11, and 12).

Oversteps, and branching and braiding are fundamental features of many strike-slip fault zones and fault systems, and they develop by a number of mechanisms: (1) bending of initially straight faults; (2) direct and indirect interaction between faults; (3) segmentation of curved faults; (4) faulting within a weak zone oblique to possible failure planes; and (5) reactivation of pre-existing extension fractures (Crowell, 1974b; Freund, 1974; Segall and Pollard, 1980; 1983; Mann et al., 1983; Aydin and Nur, 1982a, 1985 this volume). Mann et al. (1983) proposed that pull-aparts evolve from incipient to mature (“extremely developed”) basins through a sequence of closely related states. According to them, basins tend to form at releasing bends, and develop by way of spindle-shaped and “lazy S” (or “lazy Z”) basins such as the La González Basin, Venezuela (Fig. 11A;

Schubert, 1980), to rhomboidal basins such as the Dead Sea Rift (Fig. 11B; Zak and Freund, 1981; Manspeizer, 1985 this volume), and eventually, in some cases, to long narrow troughs floored by oceanic crust (e.g., the Cayman Trough of the northern Caribbean). In this model, oversteps arise by the propagation of secondary strike-slip faults in the vicinity of the releasing bend. A possible example of this branching process is the junction between the San Gabriel and San Andreas faults, the site of the Ridge Basin (Fig. 1). Between about 12 and 5 Ma, 13 km of sediment was deposited at a right bend along the San Gabriel fault, which at the same time experienced as much as 60 km of right slip (Crowell, 1982; Nilsen and McLaughlin, 1985 this volume). Beginning between 5 and 6 Ma, the San Gabriel fault ceased to be active, and strike-slip deformation was taken over by the San Andreas fault.

The interaction of parallel faults propagating from opposite directions is inherent in the models of Rodgers (1980) and Segall and Pollard (1980) for the evolution of pull-apart basins, and a possible example, the Soria Basin of northern Spain, is discussed in this volume by Guiraud and Seguret (1985). Theoretical studies indicate that significant interaction should occur between strike-slip faults if they are separated by less than twice the depth of faulting (Segall and Pollard, 1980). For strike-slip faults in California, where seismicity is observed to depths of 10 to 15 km, interaction is expected if faults are closer than 20 to 30 km. The other mechanisms listed above for generating oversteps are discussed elsewhere in this volume by Aydin and Nur (1985), and only briefly mentioned here. A consistent sense of overstepping along some curved fault zones suggests a relation between the sense of step and curvature (Aydin and Nur, 1985 this volume). An example of overstepping faults within an inappropriately oriented weak zone is the series of transform faults in the Gulf of California (Fig. 2 of Mann et al., 1983). The reactivation of pre-existing extension fractures appears to be largely a small-scale phenomenon (Segall and Pollard, 1983).

In the light of the foregoing discussion of the kinematics of strike-slip basins, we here consider the interpretation of one of the key geological elements needed to undertake a palinspastic reconstruction: piercing points of known age with which to derive the displacement history of the major strike-slip faults (e.g., Crowell, 1962, 1982, for offsets across the San Andreas and San Gabriel faults, California; and Freund et al., 1970, for offsets across the Dead Sea fault). Owing to the paucity of suitable geologic "lines," such reconstructions are commonly difficult. In the simplest case, the time of earliest movement on a strike-slip fault is given approximately by the age of the youngest rocks offset by the maximum amount ( $T_1$  in Fig. 13A). For the San Andreas fault in southern California, this is late Miocene (Crowell, 1981a). The timing for the Dead Sea fault is less well constrained as approximately mid-Miocene, but definitely younger than basaltic dikes dated as about 20 Ma (Garfunkel, 1981), and the nature of this sort of uncertainty is shown diagrammatically in Figure 13A. Another difficulty arises if displacement varies along the fault as well as increasing with time. Strike-slip faults commonly branch and intersect, and in places accommodate differential ex-

ension or shortening. Figure 13 illustrates alternative interpretations of the displacement history of a hypothetical fault, given piercing points of known age ( $T_0, T_1, T_2, T_3$ ), known displacement magnitude ( $D_0, D_1, D_2, D_3$ ), and known location along one side of the fault ( $L_0, L_1, L_2, L_3$ ). In one interpretation (Fig. 13A), the fault moves episodically over a long interval, beginning between times  $T_1$  and  $T_2$ , and ending at  $T_4$ . For rocks of a given age, the magnitude of displacement is the same at all points along the fault. In the second interpretation (Fig. 13B), faulting is confined to a short interval between times  $T_3$  and  $T_4$ . Piercing points of different ages indicate different magnitudes of displacement, because displacement varies along the fault. In reality, a given fault may exhibit displacement that is both variable and episodic, with the rate of movement changing from one interval of geologic time to another (not shown in Fig. 13). A given data set may also be incompatible with the extreme interpretations presented. The point of the illustration, however, is that in areas of complex deformation it is important to interpret the offset history of a given strike-slip fault in a regional plate-tectonic context, incorporating other stratigraphic clues to the timing of faulting.

#### *Distinctive Aspects of the Stratigraphic Record*

Strike-slip basins are present in many different plate tectonic settings, and they are filled with sediments deposited in a variety of marine and non-marine environments, subject to a range of climatic conditions. In spite of these obvious differences, however, certain aspects of the stratigraphic record appear to be distinctive. These are (1) geological mismatches within and at the boundaries of basins, that is, features which document the occurrence of strike slip; (2) a tendency for longitudinal as well as lateral basin asymmetry, owing to the migration of depocenters with time; (3) evidence for episodic rapid subsidence, recorded by thick stratigraphic sections, and in some marine basins by rapid deepening; (4) the development of pronounced topographic relief, which is associated with abrupt lateral facies changes and local unconformities at basin margins; and (5) marked differences in stratigraphic thickness, facies geometry, and the occurrence of unconformities from one basin to another in the same region.

*Geological Mismatches.*—A geological mismatch occurs where rocks juxtaposed by a fault require a considerable amount of displacement to have taken place on the fault or on another structure cut by the fault. Such mismatches occur at sutures, in thrust and fold belts, and across some low-angle normal faults in extensional allochthons. They are also common in regions deformed by major strike-slip faults. The segment of the San Gabriel fault between the southern part of the Ridge Basin and the eastern Ventura Basin is a good example (Fig. 14; Crowell, 1982). Lateral facies relations, clast-size trends, and paleocurrent data suggest that conglomerate of the upper Miocene Modelo Formation of the Ventura Basin was derived from a nearby source to the northeast. The conglomerate consists of distinctive clasts of gabbro, norite, anorthosite, and gneiss as large as 1.5 m in diameter for which no nearby source is known across the

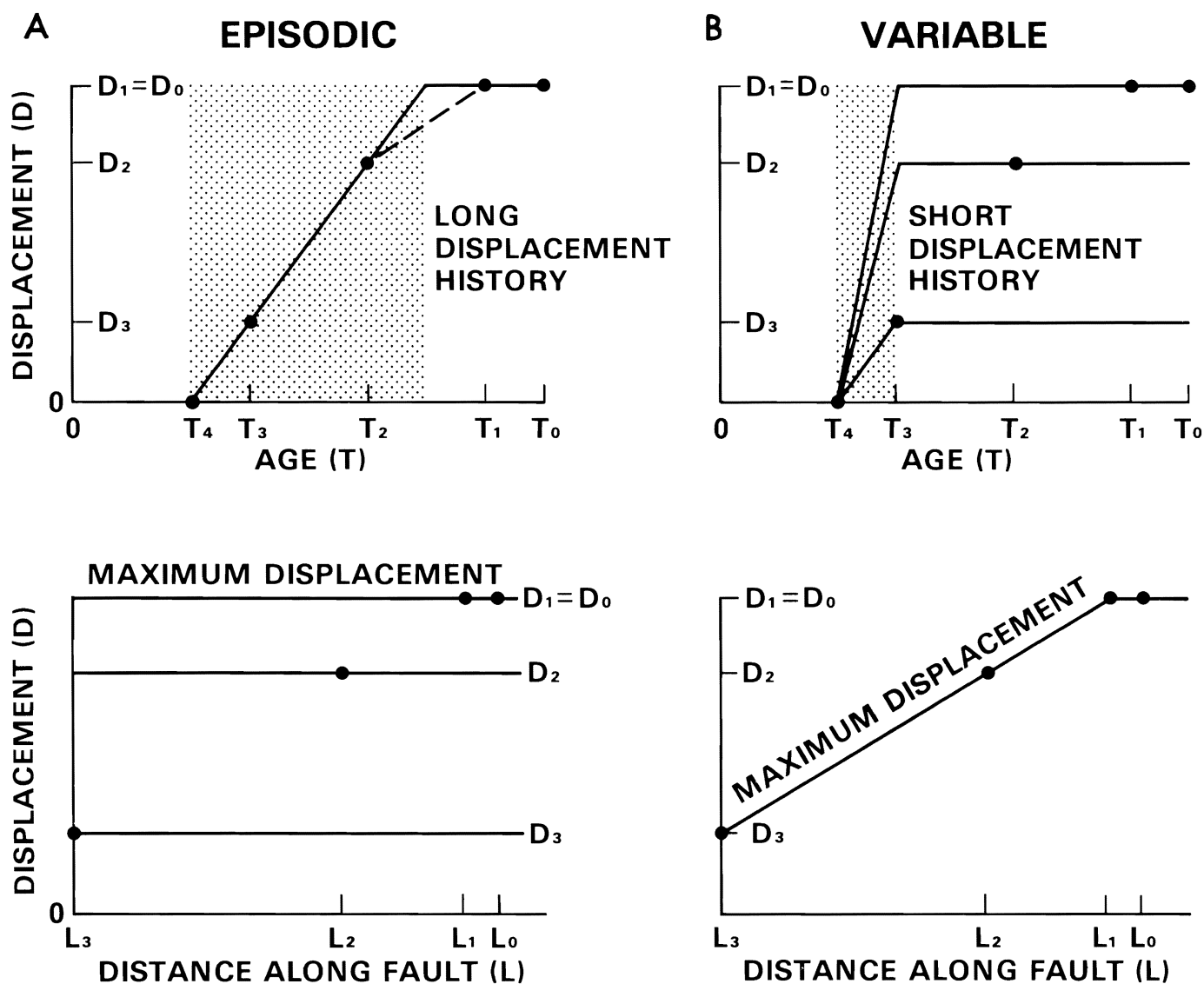


FIG. 13.—Alternative interpretations of the displacement history of a hypothetical fault, given piercing points of known age ( $T_0, T_1, T_2, T_3$ ), known displacement magnitude ( $D_0, D_1, D_2, D_3$ ), and known location along one side of the fault ( $L_0, L_1, L_2, L_3$ ).

A) The fault moves episodically over a long interval, beginning between time  $T_1$  and time  $T_2$ , and ending at  $T_4$ . For rocks of a given age, the magnitude of displacement is the same at all points along the fault. Note that by connecting displacement-age pairs, one might infer (perhaps incorrectly) that faulting began as early as time  $T_1$ , and that the slip rate increased after  $T_2$  (dashed line).

B) Movement along the fault is confined to a short interval between times  $T_1$  and  $T_4$ . Piercing points of different ages indicate different magnitudes of displacement, because displacement varies along the fault.

San Gabriel fault. Instead, that region is underlain by thick Miocene and older sedimentary rocks overlying a basement terrane quite different from that represented in the Modelo clasts. Immediately across the fault from the Modelo conglomerate, the Violin Breccia, also upper Miocene and also marine, was derived from a predominantly gneissic source to the southwest, where the basement is still largely covered by Miocene and older sedimentary rocks. The mismatch between sediments and suitable source rocks on both sides of the San Gabriel fault can be resolved, however, by removing between 35 and 60 km of right slip (Crowell, 1982). We emphasize that although evidence of this sort is important for documenting the magnitude or even occur-

rence of lateral offsets, the presence of similar geology on opposite sides of a fault at a particular locality does not necessarily preclude strike-slip deformation, a phenomenon that Crowell (1962) termed regional trace slip.

*Basin Asymmetry.*—Many sedimentary basins are asymmetrical, especially if faults occur preferentially along one side, and as in the case of grabens formed by regional extension (Harding, 1984), the sense of asymmetry in strike-slip basins may change from one profile to another (e.g., the Gulf of Elat; Ben-Avraham et al., 1979). The faults bounding strike-slip basins may be characterized by either normal separation (e.g., the Dead Sea Rift; Fig. 15A) or reverse separation (e.g., the Ventura Basin, California; Fig.

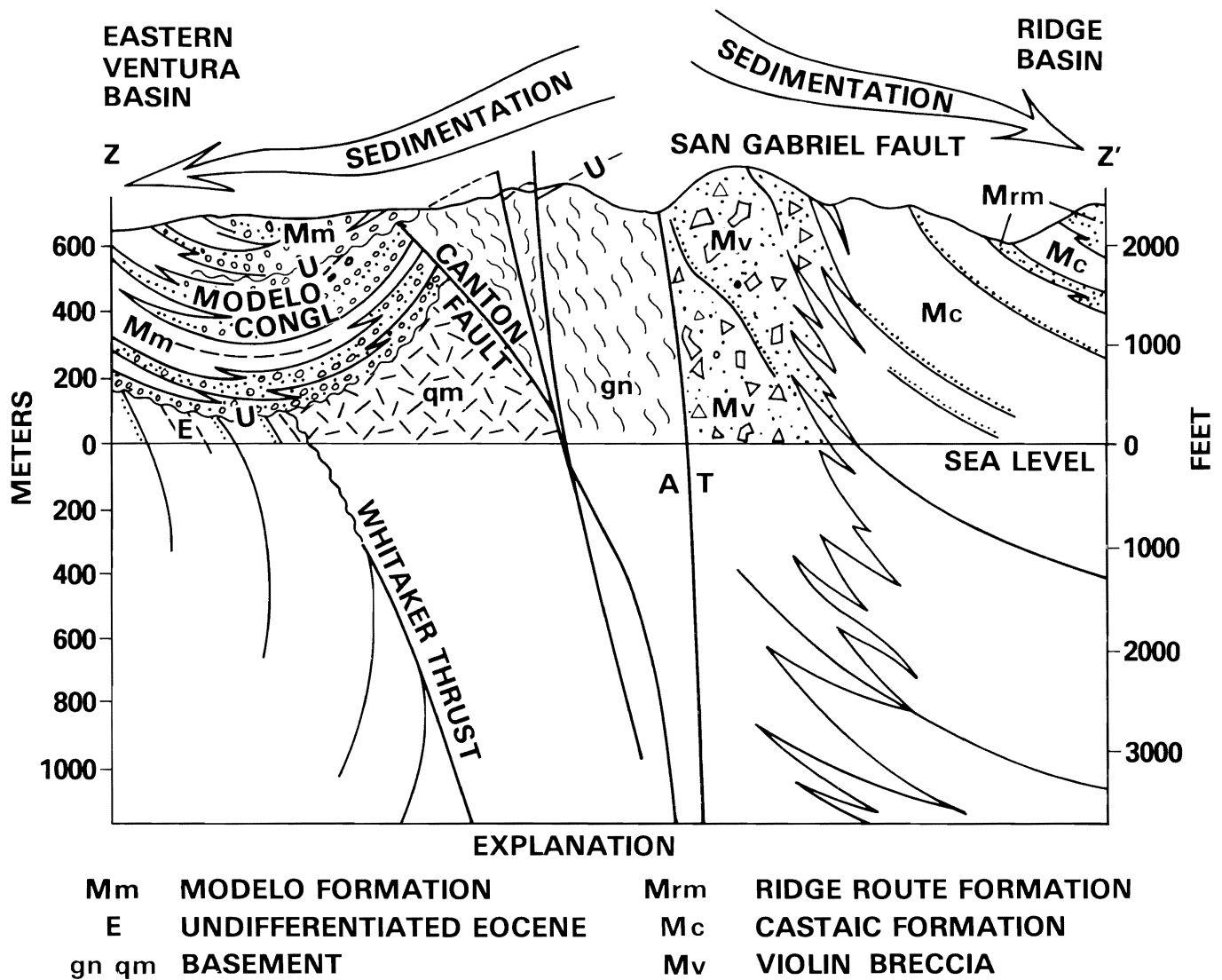


FIG. 14.—An example of a geological mismatch across a strike-slip fault: the San Gabriel fault, California (from Crowell, 1982; see Fig. 1 for location). The Modelo Conglomerate, derived from the northeast, is faulted against the Violin Breccia, derived from the southwest. T, displacement toward the observer; A, displacement away from the observer.

15B), or as discussed in the section on structural style, both normal and reverse faults may be present in the same basin (see Nilsen and McLaughlin, 1985 this volume). A particularly distinctive feature of strike-slip basins is the tendency for longitudinal as well as lateral asymmetry. The depocenter of the Dead Sea Basin, for example, has migrated northward more than 100 km from the site of the Arava Valley in the Miocene to the present Dead Sea (Fig. 11B; Zak and Freund, 1981; Manspeizer, 1985 this volume). Other basins with longitudinal asymmetry described in this volume are the Ridge Basin, California, and Hornelen Basin, Norway (Nilsen and McLaughlin, 1985), the Soria Basin of northern Spain (Guiraud and Seguret, 1985), the Nonacho Basin of Canada (Aspler and Donaldson, 1985), and possibly the Central Basin of Spitsbergen (Steel et al., 1985). *Episodic Rapid Subsidence.*—Strike-slip basins are characterized by extremely rapid rates of subsidence (Fig. 16).

even more rapid than many grabens and foreland basins, and where there is an abundant sediment supply, by very thick stratigraphic sections in comparison with lateral basin dimensions (S. Y. Johnson, 1985; Nilsen and McLaughlin, 1985, both in this volume). For example, about 13 km of sediment accumulated in the Ridge Basin in only 7 m.y. (Crowell and Link, 1982), and 5 km of sediment was deposited in the Vallecito-Fish Creek Basin in about 3.4 m.y. (N. M. Johnson et al., 1983). The Ventura Basin subsided nearly 4 km in the past 1 m.y. (Fig. 15B; Yeats, 1978). Marine basins and some deep lakes tend to become temporarily starved of sediment, a situation that promotes the accumulation of fine-grained organic-rich sediments suitable for the generation of petroleum (Graham et al., 1985; Link et al., 1985 this volume). Depending on local patterns of deformation, however, the subsidence in strike-slip basins is also episodic, and may end abruptly. The Vallecito-



Fish Creek Basin has been uplifted more than 5 km in the past 0.9 m.y. (N. M. Johnson et al., 1983).

*Local Facies Changes and Unconformities.*—Although not diagnostic of a strike-slip setting, many strike-slip basins form adjacent to uplifted blocks with pronounced topographic relief. As described in many of the papers that follow, this leads to very coarse sedimentary facies along some basin margins and to abrupt lateral facies changes. Local

vertical movements of blocks result in localized unconformities.

*Contrasts Between Basins.*—Again, because of local tectonic controls, patterns of sedimentation vary markedly from one basin to another within the same region (see the description of Eocene sedimentation in Washington by Johnson, 1985 this volume). In the case of basins for which original geometry has been obscured by subsequent defor-

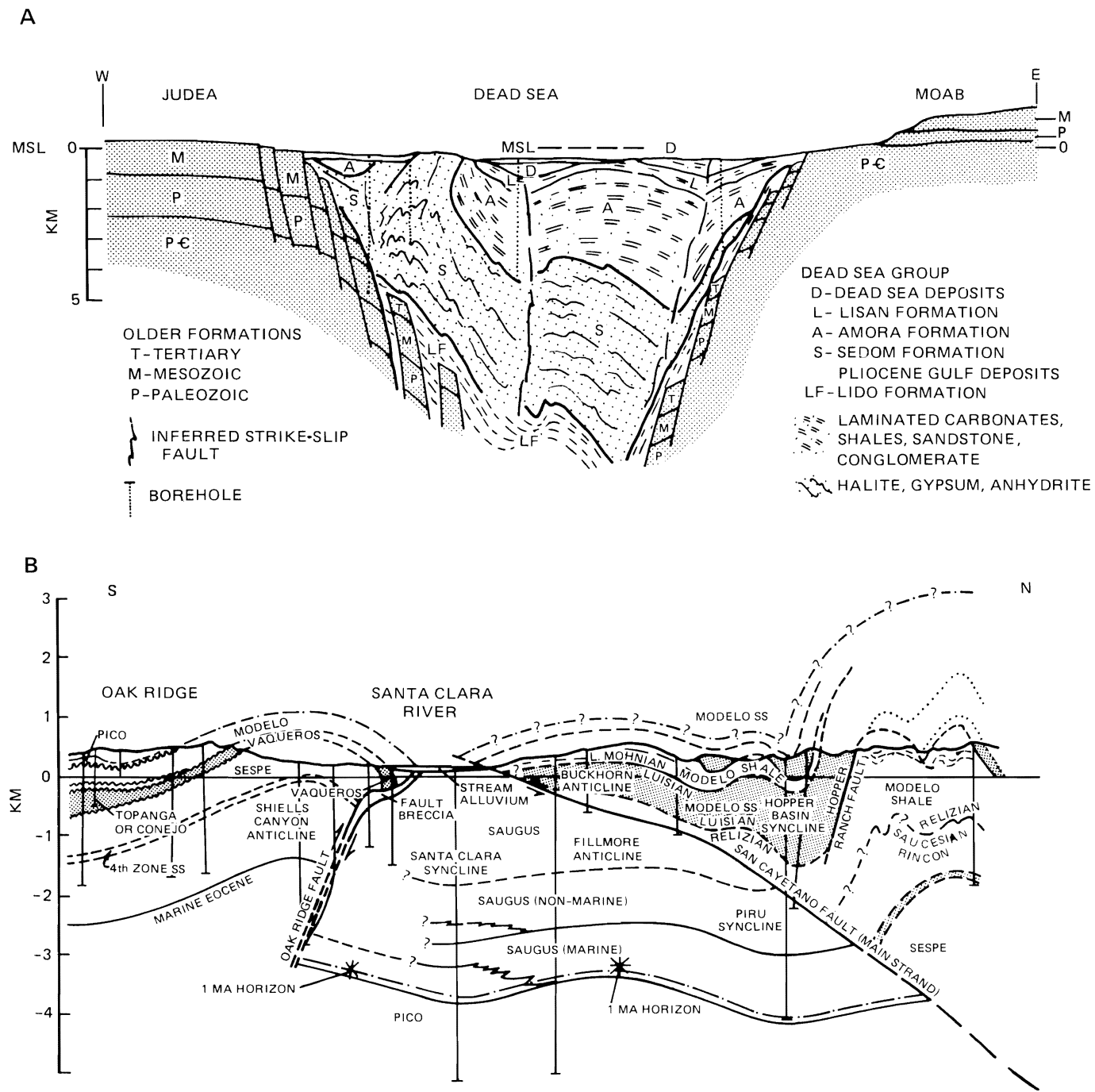


Fig. 15.—A comparison of strike-slip basins in profile. A) The Dead Sea Rift, bounded by faults with normal separation (from Zak and Freund, 1981; see Fig. 11B for location). B) The Ventura Basin, California, bounded by faults with reverse separation (from Yeats, 1983; see Fig. 1 for location).

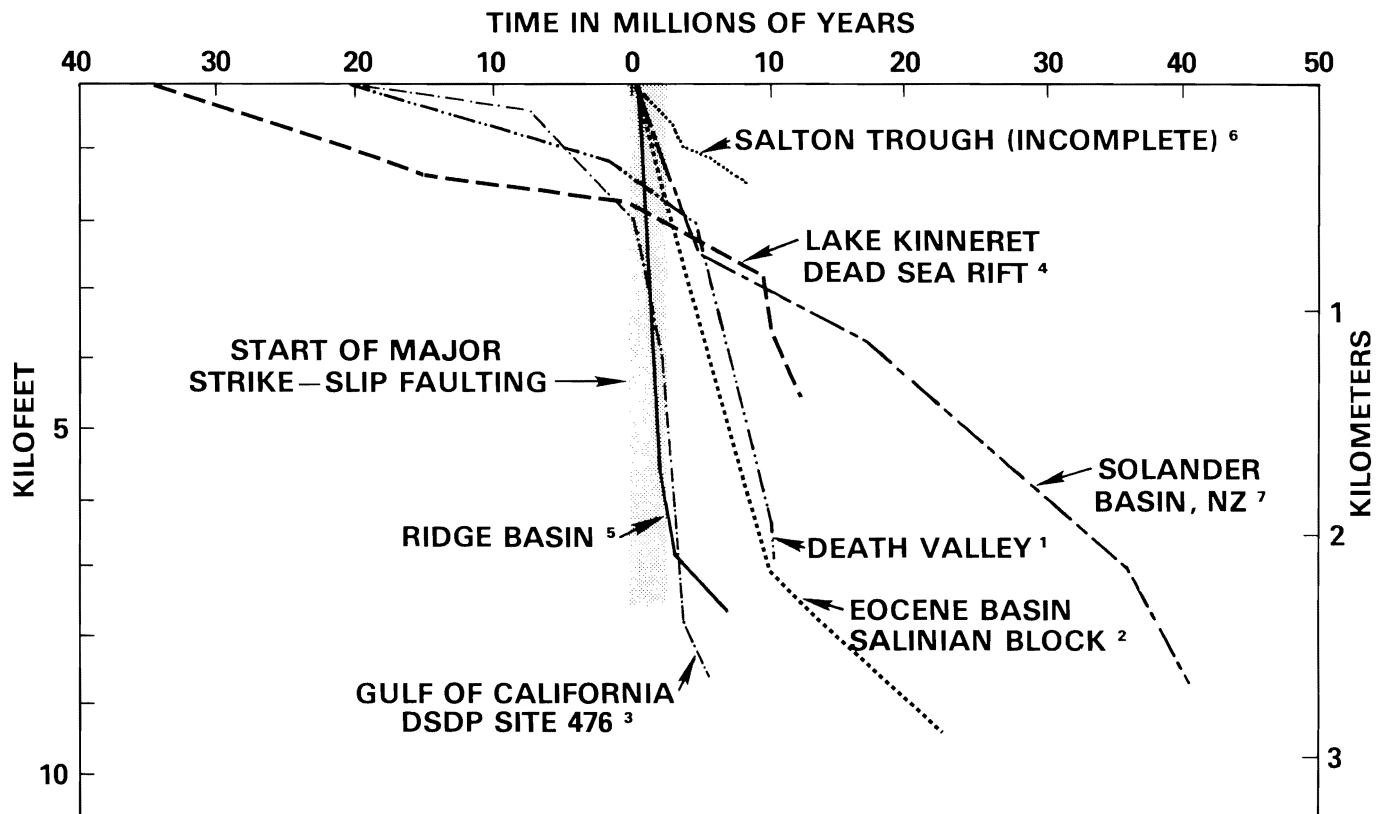


Fig. 16.—Tectonic subsidence curves for selected strike-slip basins. Tectonic subsidence is calculated by correcting cumulative stratigraphic thicknesses for the effects of sediment loading, sediment compaction, and water-depth changes. For each basin, the patterned area in the vicinity of the origin represents the approximate time of onset of major strike-slip deformation. Sources of data: 1, Hunt and Mabey (1966); 2, Graham (1976); 3, Curray et al. (1982); 4, Ben-Avraham et al. (1981); 5, Crowell and Link (1982), and papers therein; 6, Kerr et al. (1979); 7, Norris and Carter (1980).

mation (e.g., many Paleozoic and Precambrian examples), regional stratigraphic comparisons may provide some of the more important clues that sedimentation was accompanied by strike-slip deformation (Heward and Reading, 1980; Aspler and Donaldson, 1985 this volume).

#### ECONOMIC IMPORTANCE

Oil and gas are the most important resource exploited in strike-slip basins. Huff (1980) estimated that as much as 133 billion barrels has been discovered, but we regard this figure as rather high. The Dead Sea, known as Lake Asphaltitis to the ancient Greeks, was mentioned by the Latin authors Josephus and Tacitus as a source of asphalt used for caulking ships and preparing medicines (Nissenbaum, 1978). In modern times, some of the most productive basins have been in California. About 10 billion barrels of oil have been produced from the Los Angeles, Ventura, Santa Maria, Salinas, and Cuyama Basins alone (Taylor, 1976). Current production from these basins is about 600,000 barrels of oil per day, and the ultimate recovery is estimated to be as much as 20 billion barrels of oil equivalent. The Point Arguello oil field, discovered in the offshore Santa Maria Basin in 1981, is the largest U.S. discovery since Prudhoe Bay, Alaska, with between 2.2 and 2.5 billion barrels of oil in place, and total recoverable reserves of be-

tween 300 and 500 million barrels (Crain et al., 1985). Other northern California offshore basins are promising (Crouch and Bachman, 1985). Additional reserves may be discovered in provinces affected by strike-slip deformation elsewhere in the world, such as the southern Caribbean region (Leonard, 1983), the Bering Sea (Marlow, 1979; Fisher, 1982), the Dead Sea (Wilson et al., 1983), and parts of China (Li Desheng, 1984). Individual strike-slip basins range from very rich to non-productive, depending on such factors as the presence of source rocks, thermal history and maturation, migration potential, reservoir quality and distribution, occurrence of traps and seal, and preservation of trapped hydrocarbons (Kingston et al., 1983b; Biddle and Christie-Blick, 1985b). Most important is the timing of maturation, migration and trap formation, because strike-slip basins tend to be short-lived. Examples of commodities, other than petroleum, found in strike-slip basins are ground water, coal and lignite, geothermal energy, and subsurface brines.

#### CONCLUSIONS

(1) Structural patterns along strike-slip faults differ in detail from simple predictions. The geometry of faults and folds along a given fault zone is generally a result of protracted, episodic deformation of heterogeneous sediments

and sedimentary rocks, involving the rotation of crustal blocks at a variety of scales, as well as strike slip, together with varying degrees of convergence and divergence, all superimposed on a pre-existing structural grain. On a regional scale, distinctive aspects of the structural style are the occurrence of en echelon structures, evidence for simultaneous shortening and extension, and random directions of vergence in associated thrusts and nappes.

(2) Basins form along strike-slip faults as a result of localized crustal extension and/or localized crustal shortening. The main processes leading to subsidence are mechanical thinning, thermal contraction, and loading due to the convergence of crustal blocks, amplified by the effects of sediment loading. The rifting stage is preferentially represented in the stratigraphic record of many strike-slip basins, because they are typically narrow, and in some cases detached (and therefore subject to less post-rift thermal subsidence), and many basins experience subsequent uplift and erosion. The importance of lithospheric flexure depends on the amount of extension involved in basin formation. Flexural effects are probably reduced during rifting.

(3) The location of depositional sites along strike-slip faults is controlled by crustal type and the configuration of pre-existing crustal structures, variations in the motion of adjacent lithospheric plates, and the kinematic behavior of crustal blocks. A key factor influencing the development of sedimentary basins is the manner in which overall plate motion is accommodated by discrete slip on major faults, and by the rotation and internal deformation of blocks. Subsidence history cannot be determined with confidence from present fault geometry, which therefore provides a poor basis for basin classification.

(4) Distinctive aspects of the stratigraphic record along strike-slip faults are geological mismatches; a tendency for basins to be asymmetrical both longitudinally and laterally; thick stratigraphic sections representing short intervals of time; the occurrence of abrupt lateral facies changes and local unconformities; and marked differences in stratigraphic thickness, facies geometry, and occurrence of unconformities from one basin to another in the same region.

(5) A major frontier for research in strike-slip basins is that of integrated geophysical, geological and modelling studies in a variety of plate-tectonic settings. The geophysical work should include standard seismic reflection, seismic refraction, gravity, magnetic and heat flow measurements, together with paleomagnetic studies to establish the magnitude and timing of rotations, and the boundaries of the blocks experiencing rotation. The geological work should include structural, stratigraphic and sedimentological studies using outcrop, borehole and seismic reflection data, together with investigations of diagenesis and paleothermometry. Modelling should be directed at using all available data in quantitative tests of our notions of the processes that control the development of strike-slip basins.

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