Busby + Ingersoll (1995) p. 425 - 456

Strike-Slip Basins

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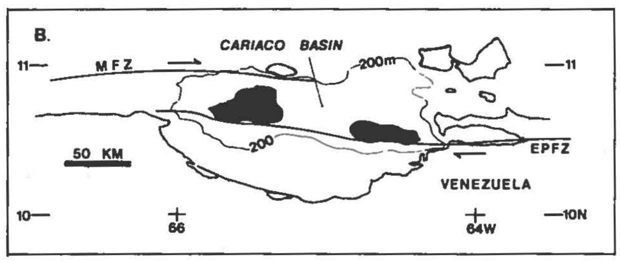
Introduction

Various types of basins, previously grouped together by many workers as "pull-apart basins" (Burchfiel and Stewart, 1966), commonly form along major strike-slip faults. We refer to these basins herein as *strike-slip basins,* following Mann et al. (1983), because "pull-apart basins" are only one of a range of basin types that may develop as a result of strike-slip faulting.

Strike-slip basins range in size from small sag ponds to rhombochasms as wide as 50 km (Fig. 12.1). Basin length-to-width ratios are typically 4:1, with a range from 1 to 10:1 (Aydin and Nur, 1982). The surface dimensions of a strike-slip basin may be measured either structurally by bounding faults or flexures, or physiographically by the extent of the area of subsidence. The basin margins, however, are commonly deformed by younger folds and faults, or have been displaced from the basin depocenter by continued strike slip. Bounding faults may dip in varying directions and be of various types. As a result, definition of basin size and shape may be a difficult task because bounding faults may dip either more steeply or more gently with depth, and may merge at depth into a single master fault; some basin margins may sag rather than be faulted. Some strike-slip basins



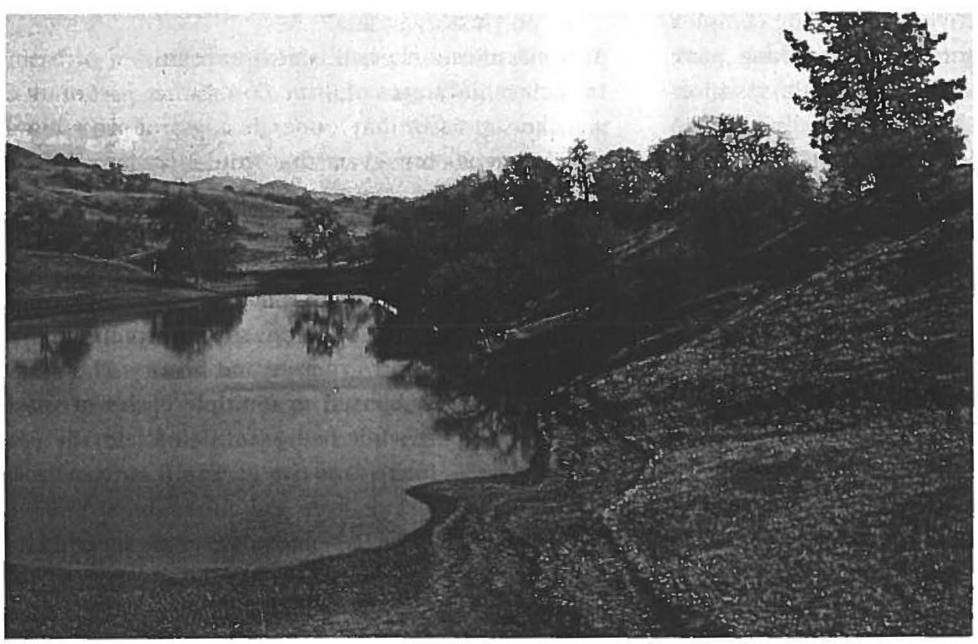


Fig. 12.1 Examples of range of scales of stepover basins. A) Photograph of sag pond on central San Andreas fault, California. B) Fault and bathymetric map of Cariaco basin at right step between dextral Moron (MFZ) and El Pilar (EPFZ) fault zones, Venezuelan borderland (Reproduced with permission from Mann, et al., 1983).

in: Busby + Ingersoll (1995) Tectonics of Sedimentary Basins may also be bounded at their bases by sub-horizontal faults.

Strike-slip basins are common and have been best studied along continental transform boundaries, such as the San Andreas fault of California, where extensive hydrocarbon exploration has yielded abundant three-dimensional control. The processes leading to crustal extension and subsidence in strike-slip settings are generally not as well understood as they are in other tectonic settings, however, and they may be extremely complex at different scales. Basins that develop along and adjacent to strike-slip faults are generally small and short-lived compared to those that develop in other plate-tectonic settings. Thermomechanical models for their formation, as well as structural and stratigraphic models of their evolution, are generally poorly developed. Each basin must be examined in its temporal, spatial, and tectonic setting within a very broad framework of paradigms that have been collected from strike-slip basins around the world. New concepts are required to model their behavior and resolve their sedimentary and tectonic histories.

Classification of Faults and Basins

The plate-tectonic settings of strike-slip faults and strike-slip basins, as developed by Woodcock (1986) and Reading (1980), respectively, reflect the complex settings and histories of horizontal slip along plate margins and provide a framework for fault classification (Fig. 12.2). Strike-slip faults may be divided into transform and transcurrent faults (Table 12.1; Sylvester, 1988).

The term "transform fault" is used for strike-slip faults that define plate boundaries, penetrate the crust, and include: (1) ridge transforms that displace spreading ridges; (2) boundary transforms that accommodate horizontal displacement between plates; and (3) trench-linked transforms sub-parallel to trenches in convergent settings.

The term "transcurrent fault" is restricted to intraplate faults that penetrate only the upper crust and include: (1) indent-linked strike-slip faults in zones of convergence and tectonic escape; (2) intracontinental strike-slip faults that are unrelated to

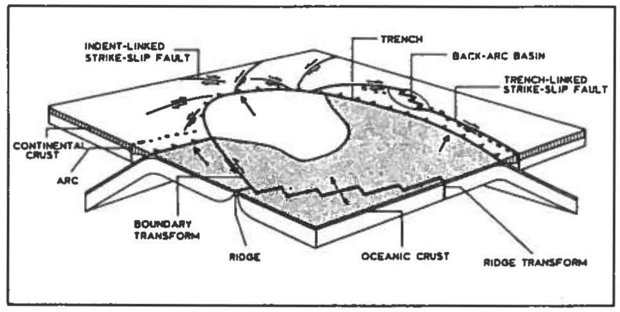


Fig. 12.2 Plate-tectonic settings of strike-slip faults (Reproduced with permission from Woodcock, 1986). Diagram is a schematic representation of India-Asia collision, looking northeast from southern Indian Ocean (compare with Fig. 10.5).

indentation and typically separate distinct regional tectonic domains; (3) tear faults that accommodate the differential displacement within or at the margin of an allochthon; and (4) transfer faults connecting overstepping segments of parallel or en-echelon strike-slip faults (Allen and Allen, 1990).

We adopt a six-fold classification of strike-slip basins based on the geometry of bounding faults and the kinematic setting of the basin (Fig. 12.3):

1. fault-bend basins
2. stepover basins
3. transrotational basins
4. transpressional basins
5. polygenetic basins
6. polyhistory basins

This classification permits easy recognition of basins in their initial stages of formation. Either part or all of a strike-slip basin may undergo stages of growth and development, however, that could fit into any of these basin types. As strike-slip basins evolve through time and space, they may undergo significant translation along principal strike-slip zones. Changes in their plate-tectonic framework, even including minor changes in direction and rate of plate motion, may radically affect basin structure and history. The basins are commonly subjected to multiple cycles of subsidence and uplift while being translated laterally (the "porpoising" effect), and in general, are partly to wholly uplifted.

Fault-bend basins generally develop at releasing bends along strike-slip faults (Figs. 12.3, 12.4). Extension results as one fault block slides past and away

Table 12.1 Classification of strike-slip faults by Sylvester (1988), based partly on Woodcock's (1986) relation of strike-slip faults to plate-tectonic setting (see Fig. 12.2)

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

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

from the other. Sharply defined clefts may form at the mesoscopic scale along releasing fault bends. At larger and longer scales, the sliding block may sag into the extended zone. More commonly, parts of both blocks sag toward the fault to form an elongate zone of subsidence along the fault bend, referred to as a lens-shaped basin (Crowell, 1974b) or a "lazy pull apart" (Mann et al., 1983). How the walls of a fault-bend basin converge at depth and merge with the master fault is largely unknown. Fault-bend basins may also form near restraining bends, where one of the fault blocks extends differentially as it slides around the bend, out of the zone of restraint.

Fault-bend basins are strongly asymmetric, have prominent coarse-grained aprons along their principal displacement zones, are commonly lens-shaped in map view, and generally develop in transtensional settings, subsequently undergoing inversion in transpressional settings. These basins may resemble other rift basins (see Chapter 3), especially in two-dimensional seismic-reflection profiles.

Stepover basins (Fig. 12.3B) form between the ends of two parallel to sub-parallel strike-slip faults that are not connected (Aydin and Nur, 1985; Schubert, 1986; Sarewitz and Lewis, 1991). Because the zone between two fault segments in en-echelon arrangement is termed a "stepover," basins that form as a result of extension between the two faults are referred to as "stepover basins." The bounding faults may merge at depth into a single master fault. At basement level, the extended domain between the two en-echelon faults generally forms a mesh-like arrangement of normal and strike-slip faults that are steep at depth. Stepover basins typically

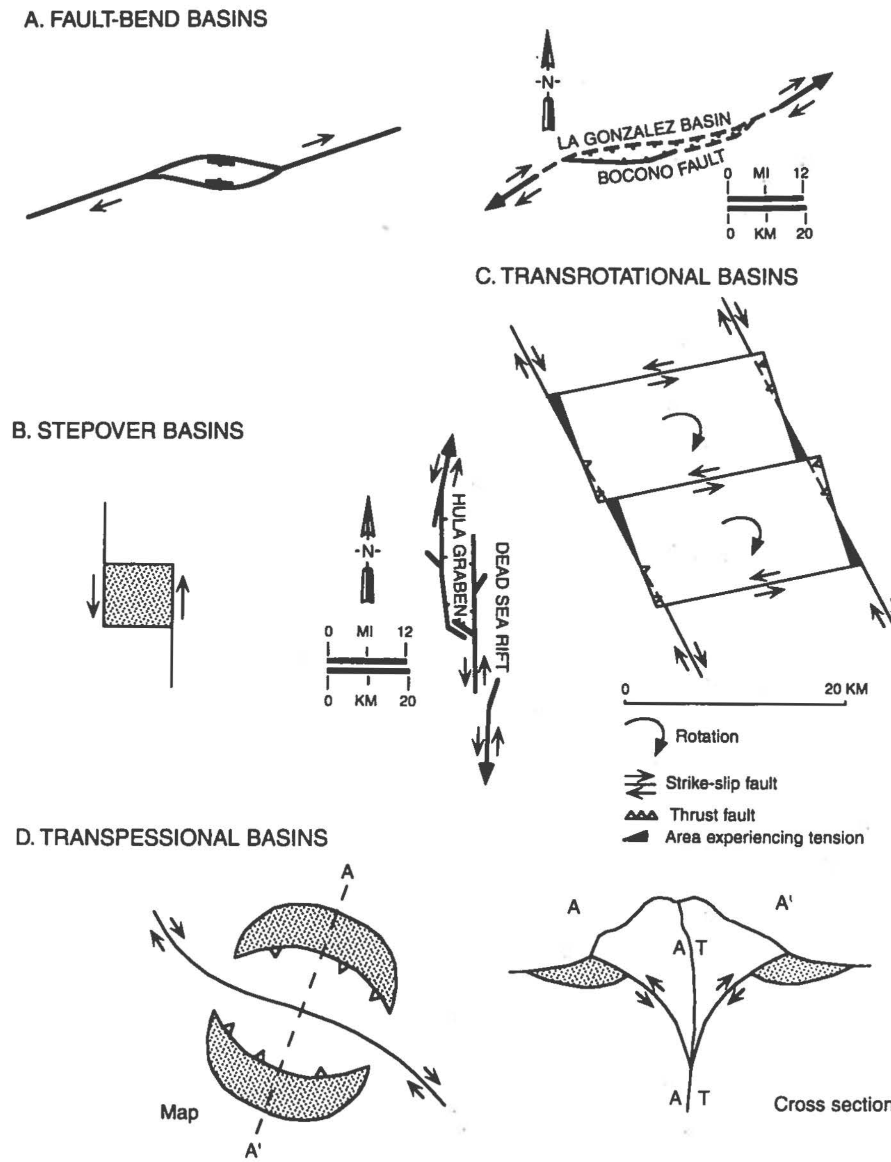


Fig. 12.3 Diagrammatic maps of six strike-slip basin types: A) fault-bend basin (left) with map of La Gonzalez basin, Venezuela (right); B) stepover basin (left) with map of part of Dead Sea rift (right); C) transrotational basins (black areas); D) transpressional basins (dot pattern) in map view (left) and cross section (right); E) polygenetic basins (dot pattern) in regional extension (left) and in regional shortening (right); and F) polyhistory basins initiated as rift basin.

form between left-stepping faults in left slip and between right-stepping faults in right slip (Fig. 12.3B).

Stepover basins may be more symmetric than fault-bend basins, with coarse-grained aprons shed basinward from all faulted margins. Depocenters may not lie preferentially adjacent to one of the marginal faults. Transverse structures may be common, segmenting the basin into separate subbasins. Continued transtension may extend and rupture the crust, producing magmatic activity, high heat flow, and in extreme cases, generation of new crust that may be younger than the overlying sedimentary succession. The floors of other stepover basins may consist of gently dipping faults at the basement-cover interface or deeper in the basement.

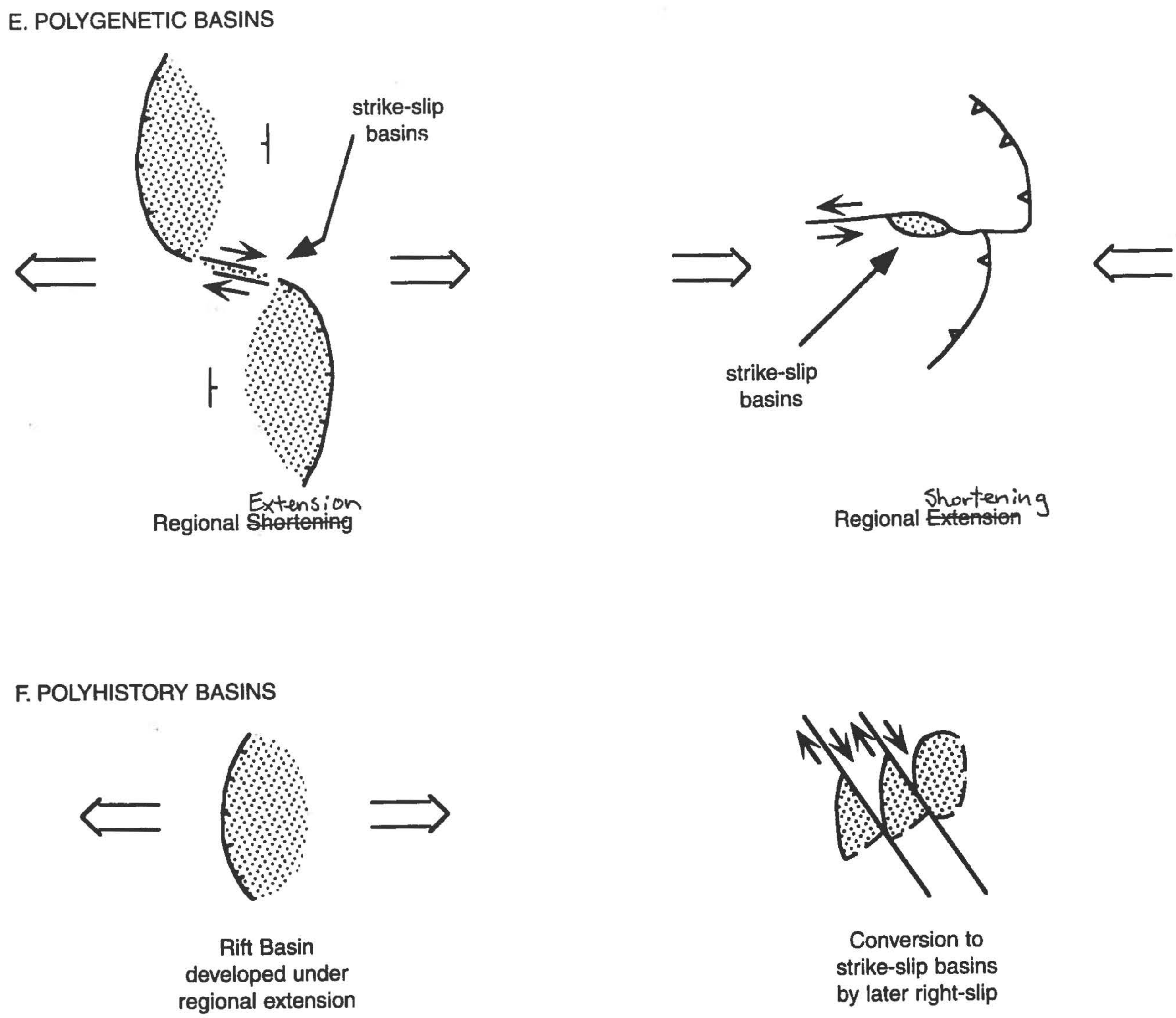


Fig. 12.3 Continued.

Transrotational basins (Ingersoll, 1988b) develop as a result of continued shear strain that causes the extension fractures and the blocks between them to rotate about a subvertical axis in the same direction as the direction of principal shear strain, clockwise in right simple shear and counterclockwise in left simple shear (Fig. 12.3C). The rate and magnitude of rotation depend on the rate of shear strain. Triangular gaps or transrotational basins form among irregularly shaped, rotated blocks (Fig. 12.5). Detachment faults within the crust may floor the basins and separate upper rotated blocks from underlying unrotated blocks. The upper block may undergo rotation and strike-slip deformation during and following deposition. Major amounts of slip along basin-bounding faults are not necessary.

Fig. 12.4 Small fault-bend basin from Superstition Hills earthquake zone. Fault strikes away from viewer, from bottom of photograph to tower on horizon. Axis of basin is about 25° to fault strike. Net right-lateral strike-slip at this site was 1.5 m (photograph by A. G. Sylvester).

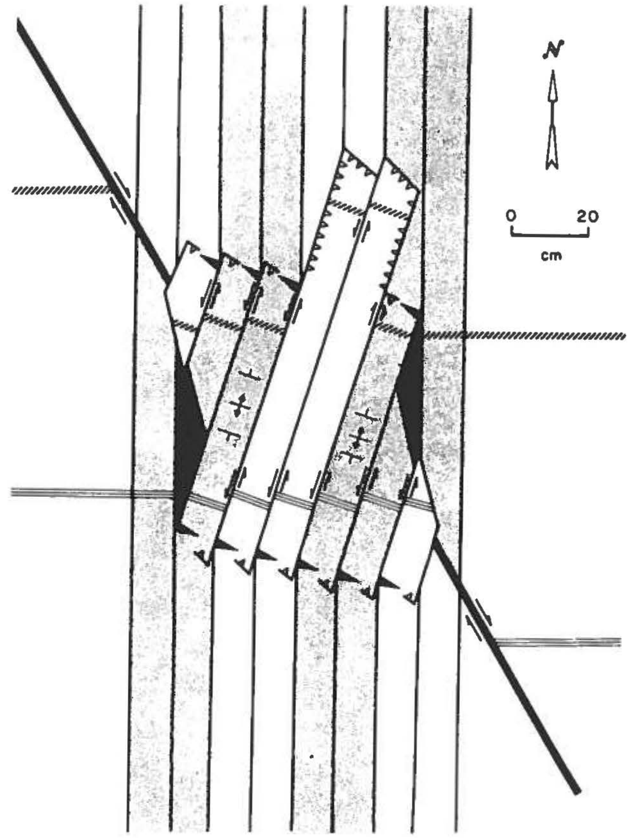


Fig. 12.5 Idealized diagram of gaps or basins (black areas) among rotated blocks in a right-simple-shear couple (Reproduced with permission from Terres and Sylvester, 1981). Hachures are on overthrust parts of blocks.

Transpressional basins (Fig. 12.3D) are generally long and narrow structural depressions, parallel to regional faults and folds, and are commonly bounded by underlying thrust faults and flanking strike-slip, or reverse faults. These basins may be dominated by axial transport of sediments parallel to structural depressions that develop in several ways, generally in response to flexural subsidence. They form next to uplifted and overthrust fault blocks that dip back into the principal displacement zone or other strike-slip faults (Fig. 12.6), typically adjacent to positive "flower" or "palm tree" structures in zones of transpression along strike-slip faults. Subsidence results from flexural loading of the marginal crust, forming mini-foreland basins adjacent to the uplifted blocks. Piggyback basins, forebulges, and related structures may develop in the transpressive regions. Paired transpressional basins may form on opposite sides of transpressive uplifts. With continued strike-slip along the principal displacement zone, patterns of sedimentation and subsidence can be very complex. Other transpressive basins may develop in response to synclinal downfolding or paired reverse faulting within transpressive regions.

Polygenetic basins (Fig. 12.3E) as defined herein develop in local regimes of strike slip in generally convergent or divergent tectonic settings. In rift settings, transfer faults or accommodation zones may link border faults of normal displacement and bound small strike-slip basins. In convergent settings, strike-slip faults and strike-slip basins may be confined to

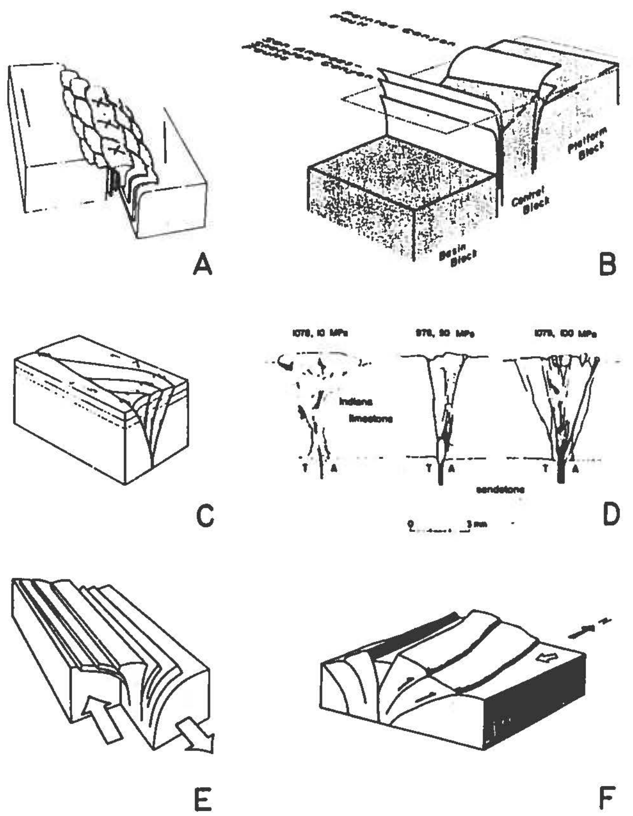


Fig. 12.6 Conceptual diagrams of flower or palm-tree structures in right simple shear (from Sylvester, 1988): A) from Lowell (1972, p. 3099); B) from Sylvester and Smith (1976); C) from Woodcock and Fisher (1986); D) from Bartlett et al. (1981); E) adapted with modifications from Ramsay and Huber (1987, p. 529); and F) with axial graben from Steel et al. (1985).

the upper plate of allochthonous thrust sheets, yielding a type of piggyback basin. Polygenetic strike-slip basins are also common in accommodation zones, where they may be confined to upper structural plates or hanging walls.

Polyhistory basins (Fig. 12.3F) are those in which episodes of pure extensional rifting or pure compressional thrusting alternate with episodes of strike slip, generating complex and commonly multicyclic basins. They can be of almost any size and shape, and they generally record pulses of subsidence caused by varying mechanisms. Many "successor basins" in complex orogenic belts are of this type; they may be long-lasting, with multiple histories of uplift and subsidence related to shifting tectonic settings. Polyhistory strike-slip basins may also have the characteristics of many of the other types of strike-slip basins, but they tend to be even more complicated because of their varying tectonic framework.

Structural Framework

Fault Pattern

The structural patterns that develop along strike-slip fault systems depend on four principal factors (Christie-Blick and Biddle, 1985): 1) the kinematics (convergent, divergent, or parallel) of the fault system; 2) the magnitude of the displacement; 3) the material properties of the rocks and sedimentary infills in the deforming zone; and 4) the configuration of pre-existing structures. These factors may yield a curviplanar principal displacement zone with subsidiary faults having widely divergent and changing strikes. Extension occurs at releasing bends (Fig. 12.3B), and shortening takes place at restraining bends (Fig. 12.6); thus, basins and uplifts may be produced by curviplanar faults as well as overstepping faults.

Because basins along and adjacent to strike-slip fault systems may change both gradually and abruptly through time and space from divergence (transtension) to convergence (transpression), their tectonic and sedimentary histories may be extraordinarily complex. In simple strike-slip fault systems, the strike of the fault relative to the block or plate-motion vectors determines whether displacement along the fault has a component of transpression or transtension.

The upward branching "flower structures," characteristic in cross sections of strike-slip faults, range from simple, single-strand features with uniform dips toward the principal displacement zone to complex, upward convex, multi-strand faults that dip in variable directions (Harding et al., 1983; Harding, 1985; Fig. 12.6). In transpressional zones, most faults that make up "positive flower structures" have the cross-sectional appearance of reverse faults; in transtensional zones, most faults that make up "negative flower structures" appear to be normal faults (Fig. 12.7). Positive flower structures have an overall antiformal character with abundant folds as a result of net shortening, whereas negative flower structures have an overall synformal character with folds that result from net extension.

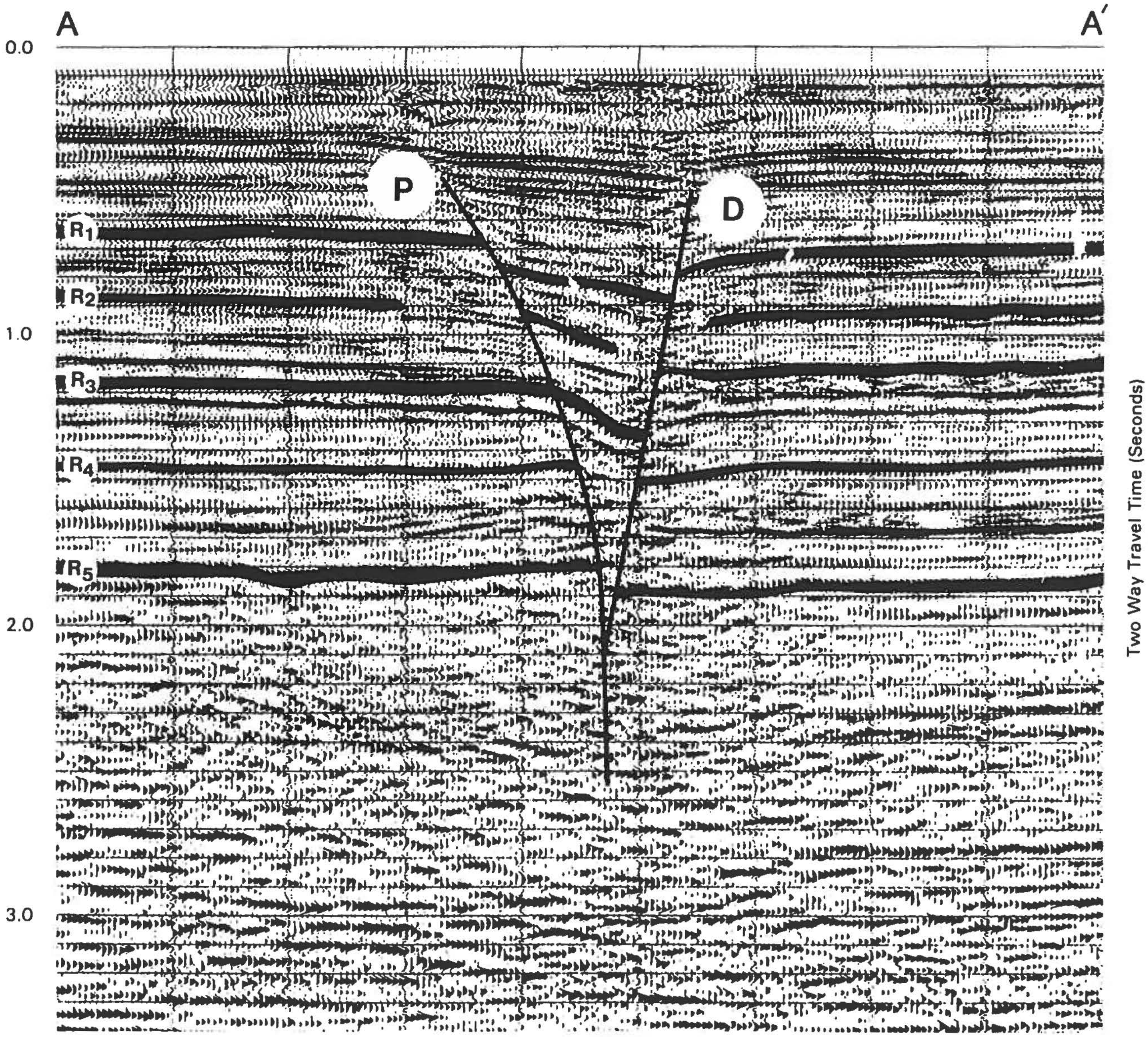


Fig. 12.7 Seismic-reflection image of negative flower structure (Reproduced with permission from D'Onfro and Glagola, 1983). P and D indicate fault names.

Where subparallel to the principal displacement zone, individual faults may have various dip directions, dip amounts, senses of displacement, and separation of rocks of different ages. Lateral changes from transpressional to transtensional settings cause gradual to abrupt changes in senses of displacement of either parts of or all of the flower structure. Because the overall strike-slip displacements include regional to local oblique as well as dip-slip displacements, individual faults must be studied in four dimensions. Useful criteria for the correct interpretation of positive and negative flower structures were described and discussed by Harding (1985, 1990).

Rotation of blocks about vertical to subvertical axes within strike-slip fault zones may also produce areas of shortening and extension (e.g., Luyendyk and Hornafius, 1987). Rotation of large blocks may result in formation of basins of regional extent. Some strike-slip basins have subhorizontal detachment surfaces that separate structurally higher, rotated crustal fragments characterized by brittle deformation from structurally lower crustal units characterized by more ductile deformation.

Basin Geometry

Strike-slip basins develop by various processes related to local transtensional and transpressive regimes. Oversteps and bent fault segments (releasing bends) commonly yield strike-slip basins that become more elongate parallel to the principal displacement zone through time. Some of the parameters that exert control on the length, width and depth of strike-slip basins are:

1. the degree of lateral and vertical curvature of the principal displacement zone and other fault surfaces
2. the depth to the brittle-ductile transition in the crust
3. the amount of displacement along the principal displacement zone and other faults
4. the age of the basin relative to the age of the principal displacement zone
5. the spacing between the overstepping faults
6. the length that the overstepping faults extend beyond one another

None of these parameters has been sufficiently quantified, unfortunately, to yield a predictive model that fits all strike-slip basins.

The type of and depth to basement rocks exert strong controlling influences on strike-slip-basin geometry. Where basement rocks are shallow, consist of massive and rigid plutono-metamorphic rocks, and are overlain by a thin cover of sedimentary strata, strike-slip basins are generally sharply defined and bounded by steeply dipping faults (Ben Avraham, 1985). Where basement consists of thick, easily deformed sedimentary or layered metamorphic rocks, overlain by a thick cover of sedimentary strata, strike-slip basins are generally poorly defined and bounded by low-angle listric transpressional or transtensional faults. As an example, the sedimentary cover on the Bering Shelf is more than 5,000 m thick and the basement rocks consist of a succession of sedimentary and layered metamorphic rocks (Worrall, 1991). Because the sedimentary cover is so thick, the strike-slip basins are less well-defined at the surface than they are at depth. Faults do not propagate from the basement all the way to the surface, and folds die out upward in the stratigraphic succession. Thus, recognition of the basin form and genesis requires three-dimensional definition and analyses at several different structural levels.

Most basins in their embryonic stages appear to be related to either releasing-bend geometries or non-parallel master faults. The development of basins along releasing bends of strike-slip faults yields s-shaped or z-shaped basins that, with continued displacement, form rhomb-shaped basins, locally referred to as "rhombochasms" (e.g., Mann et al., 1983). With continued elongation, continental crust may thin and extend sufficiently to rupture, yielding mantle uplift and generation of an oceanic spreading ridge, such as the Cayman trough (Perfit and Heezen, 1978; Rosencrantz et al., 1988). In these types of basins, length is commonly more than three times greater than width.

At the continental scale, strike-slip basins may develop between tectonic plates as they diverge in transtension or converge in transpression. Individual basins may develop by one or several mechanisms; because of their scale, the basins may be quite complex and involve diverse mixtures of normal, lateral and reverse faults. In continental interiors, equally complex basin types and fault patterns are possible (e.g., Zolnai, 1991). Basins along intraoceanic transform faults appear to be complex structurally, but simpler depositionally, as a result of lower sedimentation rates and less diverse source areas (e.g., Barany and Karson, 1989).

Extension, Subsidence and Thermal History

The formation of strike-slip basins depends largely on the orientation of the principal direction of extension relative to the direction of bulk shear strain, on the overstepping arrangement of discontinuous and discrete fault segments, and on the bending geometry of the fault. In instantaneous, homogeneous, bulk simple-shear (Fig. 12.8), the principal direction of extension is 45° to the mean shear direction. The brittle manifestations of bulk simple shear in rocks are extension fractures (designated T), shear fractures (R, R′ and P), and faults that are parallel to the principal displacement zone. The T and R fractures

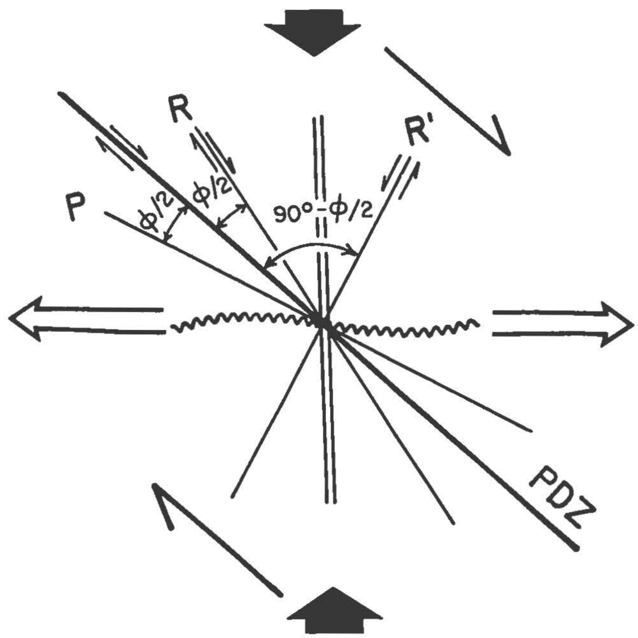


Fig. 12.8 Geometric relations among associated structures formed in right simple shear (Reproduced with permission from Sylvester, 1988). Plan view of Riedel model with orientations of structures formed in incipient right simple shear along a southeast-striking vertical fault. Double parallel line represents orientation of extension (T) fractures; wavy line represents orientation of fold axes. Abbreviations: P, P fractures; R and R′, synthetic and antithetic shears, respectively; PDZ, principal displacement zone; ø, angle of internal friction. Stubby black arrows = axis of principal shortening; open arrows = principal axis of lengthening.

develop commonly in en-echelon arrays with a self similarity from the scale of micro- and mesoscopic laboratory experiments, to surficial fault ruptures with dimensions measurable in meters, to fault zones from 1 to 2 km wide (Tchalenko, 1970). The development of R and T fractures at larger scales is problematic and depends on the width of the zone of strike-slip shearing.

Extension may form obliquely across T or R fractures, parallel to the entire shear zone, yielding a sag within the zone, or between pairs or sets of overstepping R fractures, or as an array of en-echelon, external faults (Fig. 12.8). In the latter instance, the sense of overstepping is opposite to the arrangement of stepover basins.

Two major strike-slip basin types may be distinguished on the basis of mantle involvement (Allen and Allen, 1990), although basin histories may be difficult to summarize in simple terms or to model effectively because source areas and depocenters migrate laterally through time: 1) Those in which the mantle has been involved, that is, "hot basins;" and 2) those in which the mantle has not been involved, and are thus, generally thin-skinned, that is, "cold basins."

Because strike-slip basins evolve so rapidly and complexly, most writers have had limited success in modeling their thermal evolution and subsidence/uplift histories. In "hot" basins, uniform-extension models with modifications for lateral loss of heat have been applied with some success. In "cold" basins, post-extensional thermal subsidence is insignificant. Transtension during formation of some strike-slip basins may yield extraordinarily high rates of subsidence and sedimentation; equally high rates of uplift and erosion may occur during transpression. Many strike-slip basins pass through multiple phases of subsidence and uplift during the complex evolution of the basin, and different parts of the same basin may undergo rapid uplift and rapid subsidence at the same time.

Continental-margin and intracontinental strike-slip basins may receive huge volumes of sediment, whereas intraoceanic basins may be starved of sediment. Because narrow basins cool rapidly by lateral heat conduction, they appear to undergo rapid early subsidence to form deep basins with associated sediment starvation during early stages of development (Allen and Allen, 1990). As a result of greater lateral heat loss, however, narrower basins subside at faster rates and to greater depths than do wider basins; this relation has profound implications for hydrocarbon generation and trapping in strike-slip basins. As the initial subsidence rate decreases and pathways are established for sediment to enter the basin, sedimentation rates will ultimately exceed subsidence rates, leading to basin filling.

Depositional Framework

The sedimentary fill of strike-slip basins may be extraordinarily variable and complex, depending on whether the basins are submarine, sublacustrine, subaerial, combinations of these, or variable through time and space. Nilsen and McLaughlin (1985) summarized the stratigraphic and sedimentological aspects of three well exposed strike-slip basins, the

Little Sulphur Creek (McLaughlin and Nilsen, 1982) and Ridge (Link, 1982) basins of California, and the Hornelen basin (Steel and Gloppen, 1980) of western Norway, which are principally of nonmarine origin and of varying size and character, as follows (Fig. 12.9):

1. The basins are asymmetric, with their structurally deepest parts close to and subparallel to the syndepositionally most active strike-slip margins.
2. The basins are characterized by diverse depositional facies, including talus, landslide, alluvial-fan, braided- and meandering-fluvial, deltaic, fan-delta, shoreline, shallow- and deep-lacustrine, turbidite, chemical-precipitate and algal-limestone deposits.
3. The basin fill is characterized dominantly by axial infilling, subparallel to the principal displacement zones.
4. The basin-margin deposits are distinctive. Small debris-flow-dominated alluvial fans containing coarse sedimentary breccia and conglomerate formed along the syndepositionally active principal displacement zones. Along the inactive or less active margins are larger streamflow-dominated alluvial fans and fluvial deposits that contain finer-grained conglomerate and little breccia (compare with rifts; see Chapter 3).
5. The basin fill is characterized by abrupt facies changes.

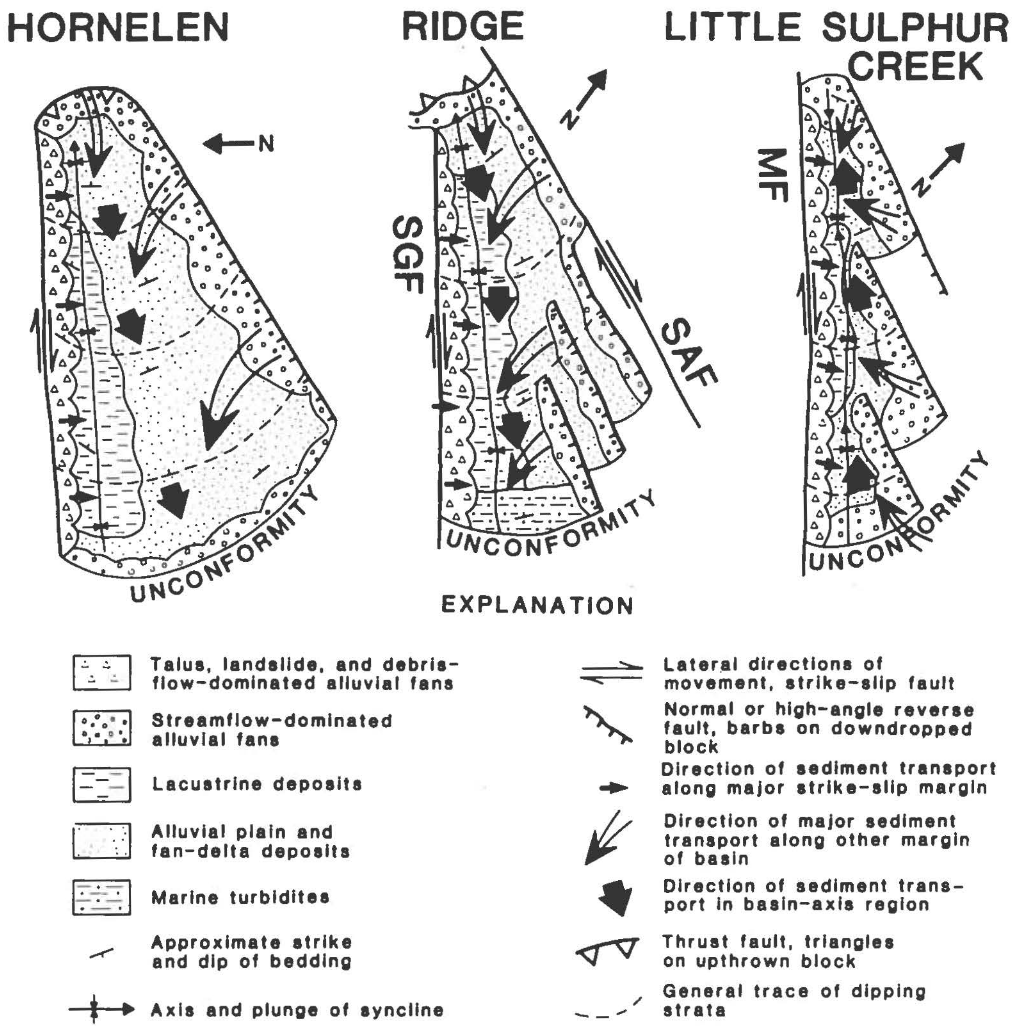


Fig. 12.9 Tectonic and sedimentary comparison of Hornelen basin, Norway, Ridge basin, southern California, and Little Sulphur Creek basin, northern California (Reproduced with permission from Nilsen and McLaughlin, 1985). Orientations and scales of basins vary considerably but are shown here at approximately the same orientation and size for comparison. Length of each basin is approximately: Hornelen, 50 km; Ridge, 25 km; Sulphur, 10 km. Abbreviations: MF, Maacama fault; SAF, San Andreas fault; SGF, San Gabriel fault.

6. The basin fill was derived from multiple basin-margin sources that changed through time as a result of continued lateral movement along basin-margin faults.
7. The basin fill may be petrographically diverse and complex as a result of the multiple sources that changed through time and space.
8. The basins contain very thick sedimentary sections compared to their areas.
9. The basin fill is characterized by high rates of sedimentation, roughly 2.5–3.0 mm/y.
10. The basins are characterized by depocenters that migrated in the same directions as source terranes along the principal displacement zones; these migration directions are generally opposite to the directions of axial sediment transport.
11. The basin fill is characterized by abundant synsedimentary slumping and deformation, possibly in response to basinwide shaking from earthquakes along basin-margin faults.

Many of these attributes are also characteristic of other generally elongate and restricted types of basins, especially rift basins, foreland basins and trench-slope basins. Strike-slip faults may form significant components, in fact, of these three basin types, as well as of many others. Recognition of strike-slip basins on the basis of structural criteria is probably a more straightforward process than is their recognition on the basis of stratigraphic and sedimentologic criteria. The most useful criterion may simply be the recognition of source areas that have been laterally displaced from sedimentary successions that contain unique detritus, especially if coarse-grained; however, even this criterion may be difficult to apply in areas of excellent exposure because: 1) the uniqueness of the source area must be proven convincingly; 2) the possible removal of similar rocks from areas that have undergone subsequent uplift and erosion must be convincingly disproved; 3) the transport paths of the unique detritus into the basin must be demonstrable; and 4) fault displacement contemporaneous with basin filling, rather than postdepositional displacement, must be demonstrable. These relations have been difficult to prove in most ancient basins, especially where strike-slip faults are no longer active or where erosion has stripped away most of the source area(s) or basin(s).

The sedimentary fill of many strike-slip basins is, in general, dominated by repetitive, basinwide, upward coarsening sequences that cut across all facies boundaries. These sequences seem to be tectonically controlled by basin-wide changes in base level that induce progradation of marginal and axial coarse-grained facies over axial finer-grained marine, lacustrine and related facies. Upward coarsening sequences are especially well developed in the Little Sulphur Creek basins, Hornelen basin and Ridge basin (Steel and others, 1977; Steel and Gloppen, 1980; Nilsen and McLaughlin, 1985). The sequences may result from periodic major tectonic subsidence of the basin, uplift of the basin margins, lowering of sea level and (or) lake level, climatic changes, or combinations of all these factors.

Recognition of Ancient Strike-Slip Basins

Several factors make recognition of ancient strike-slip basins difficult:

1. Their depositional and structural histories are complex.
2. Lateral movements along principal displacement zones and other faults commonly detach, rotate and translate basins or parts of basins from their places of origin, making paleogeographic reconstructions tenuous.
3. Polycyclic episodes of subsidence and uplift commonly remove major parts of the stratigraphic and structural record.
4. Most easily studied late Cenozoic strike-slip basins are nonmarine, resulting in poor intrabasinal and interbasinal age control and stratigraphic correlations.
5. Many cross-sectional and plan-view features of strike-slip basins resemble those of rift basins and forelands, making their recognition problematic; detailed three- and four-dimensional studies are required, and reliance upon two-dimensional seismic-reflection lines is commonly misleading.
6. Strike-slip faults are commonly reactivated as normal, reverse, oblique, thrust and more complex zones of faulting; changes in tectonic framework may yield preservation of only parts of the original strike-slip basins.

7. Few ancient deep-sea strike-slip basins survive subsequent episodes of subduction intact enough to provide suitable models for their structural and depositional evolution.

Reliable criteria for recognition of strike-slip basins need to be developed from the sedimentary pattern, stratal succession and filling mechanisms of the basins. Of the depositional criteria outlined by Nilsen and McLaughlin (1985), lateral migration of depocenters parallel to principal displacement zones appears to us to be the most useful. Depocenters in most other basins tend to migrate either transversely away from or toward the principal bounding faults, or to remain in more or less the same location through time. In strike-slip basins, however, depocenters migrate laterally, parallel to the principal bounding faults. By this process, basins lengthen through time without excessive widening, accumulate extraordinary thicknesses of fill, and contain abrupt facies changes and petrographic variations throughout their history. Polycyclic histories, changing structural settings, and repeated episodes of rapid subsidence and uplift characterize long-lived strike-slip zones, such as many of those along the San Andreas fault (Crowell, 1974a,b), even though the basins themselves are commonly short-lived.

Other useful structural and stratigraphic criteria for recognizing strike-slip basins include:

1. lateral displacement of related depocenters
2. lateral offset of matched source areas and deposits
3. presence of coarse-grained alluvial-fan, fan-delta, submarine/and sublacustrine-apron, and related deposits along the flanks of basins that contain principal displacement zones (these depocenters typically alternate along strike in rift basins from one flank to the other, reflecting activity of border faults [see Chapter 3])
4. presence of thick, but laterally restricted sedimentary sequences characterized by high sedimentation rates
5. localized uplift and erosion, resulting in unconformities of the same age as thick nearby sedimentary fill
6. abrupt lateral facies variations
7. presence of strike-slip faults on one or more sides of the basin
8. development of basinwide upward coarsening sequences that develop in response to tectonically induced basin deepening

Examples of Strike-Slip Basins

Fault-Bend Basins

Most modern and ancient strike-slip basins are probably fault-bend basins (Fig. 12.3A) because bends in strike-slip faults are common. Although modern Death Valley in California (Burchfiel and Stewart, 1966) is generally regarded as a classic example of a "pull-apart basin," we consider it and several adjacent basins as polygenetic basins (Fig. 12.3E), because they have formed within a large region dominated principally by extension.

Ridge Basin, Southern California

Ridge basin is one of the best studied fault-bend basins in the world (Fig. 12.10). It developed during Miocene and Pliocene time adjacent to the right-lateral San Gabriel fault (Crowell and Link, 1982). The basin is 30 to 40 km in length, 6 to 15 km wide, and about 400 km^2 in areal extent; its cumulative fill of 7,000 to 13,000 m was deposited at a rate of about 3 mm/y (Nilsen and McLaughlin, 1985).

Ridge basin developed as a stretched and sagged crustal wedge northeast of the San Gabriel fault in the area where the fault had a curvilinear trace. Crowell (1982a) inferred formation of the basin along a restraining bend, whereas May et al. (1993) inferred formation of the basin along a releasing bend of the San Gabriel fault (Fig 12.11A). The San Gabriel fault was the principal zone of right slip during most of the history of the basin, and its curved trace may have been the reason for uplift of a source area along the southwestern margin of the basin that migrated northwestward relative to stable North America through time, shedding detritus that formed the Violin Breccia (Crowell, 1974a,b, 1982b). The Clearwater and Liebre fault zones became active strike-slip faults as the basin widened and lengthened.

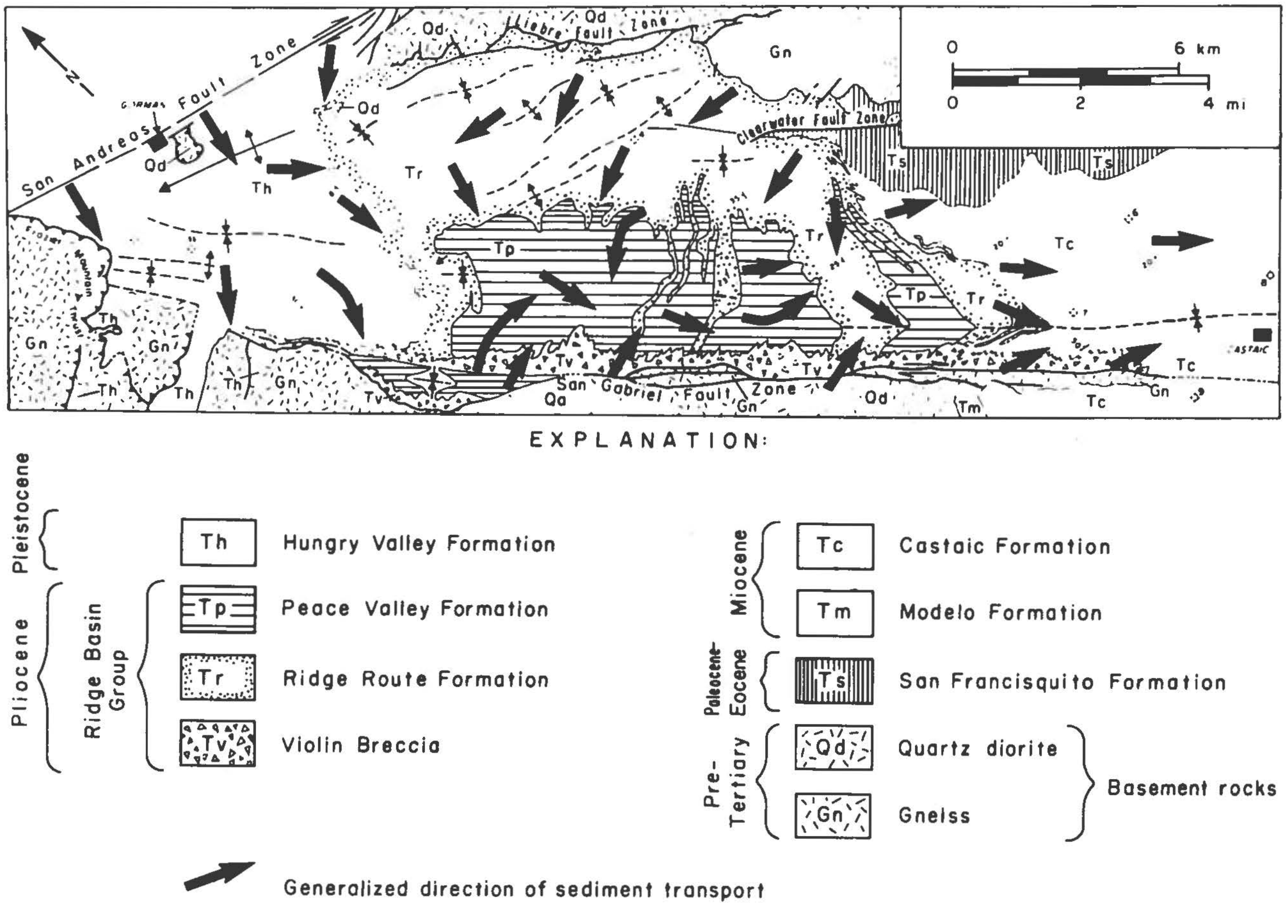


Fig. 12.10 Generalized map (left) and schematic composite stratigraphy (right) of Ridge basin (Reproduced with permission from Crowell and Link, 1982). Note that strata are shingled, such that no one vertical section contains all units (see Fig. 12.11C).

Most of the sediment filling the basin was carried by rivers draining source areas located to the northeast (Fig. 12.10). The basin depocenter migrated northwestward as the source of the Violin Breccia moved northwestward along the southwest side of the San Gabriel fault (Fig. 12.11B); as a result, the Violin Breccia is progressively younger northwestward, forming shingled sediment lobes that record northwestward migration of the depocenter. The Violin Breccia formed as a coarse-grained apron of sediment deposited transverse to the basin axis. It consists of fan-shaped bodies that were moderately deep marine during the early history of the basin, and were fan deltas and alluvial fans during the later history of the basin. These fan-delta and alluvial-fan deposits interfinger basinward with fluvial and lacustrine strata that record longitudinal basin filling, southward and southeastward down the axis of the basin concurrently with northwestward migration of the depocenter and deposition of the Violin Breccia (Fig. 12.10). Deep- to shallow-lacustrine facies were deposited in the depocenter area adjacent to the San Gabriel fault (Fig. 12.10) (Link and Osborne, 1978). The total stratigraphic thickness is much greater than the accumulated thickness at any particular locality in the basin (Fig. 12.11C). Abundant synsedimentary deformational structures in the basin fill may record penecontemporaneous seismicity. When the principal locus of right slip transferred to the San Andreas fault about 5 Ma, the basin-margin faults became inactive, and the basin was folded, uplifted, and transported northwestward about 220 to 240 km relative to stable North America.

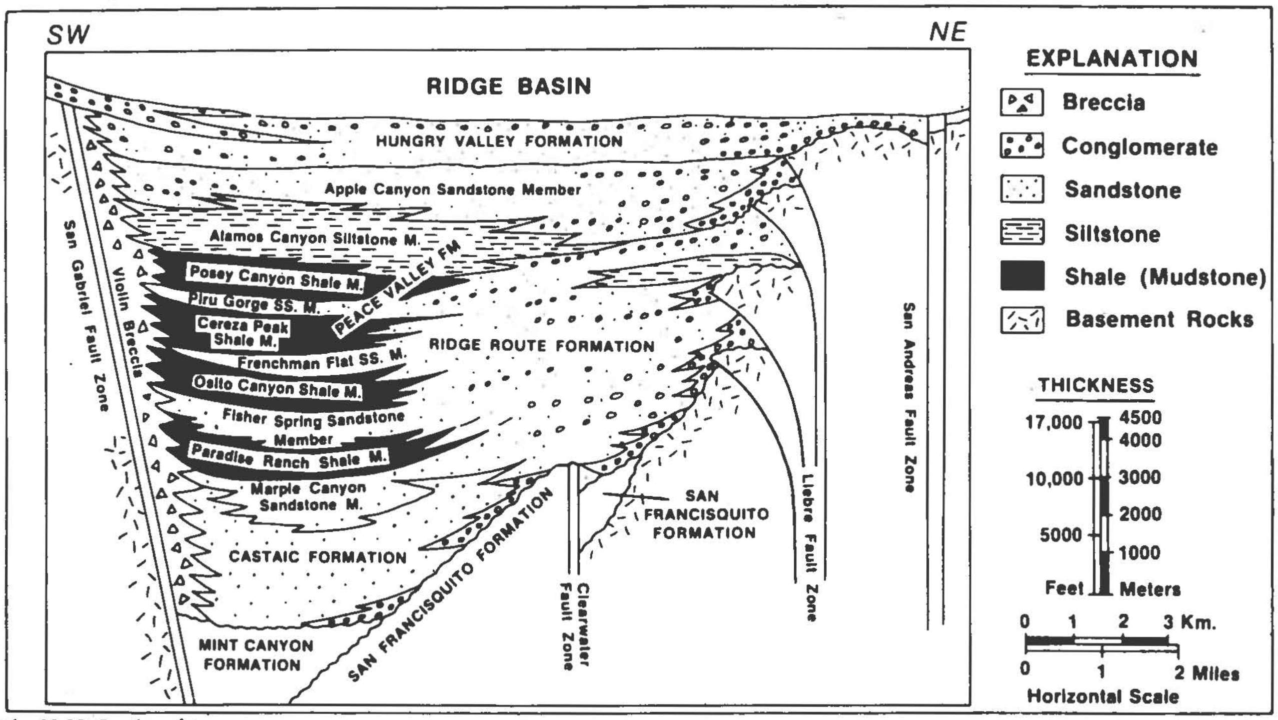
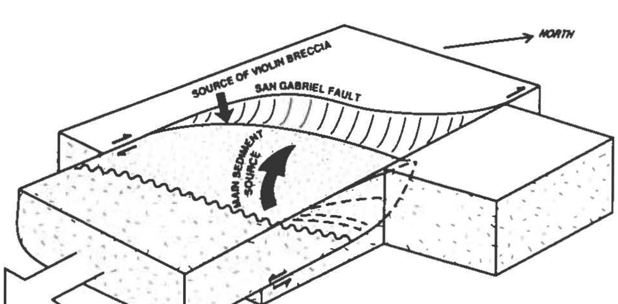
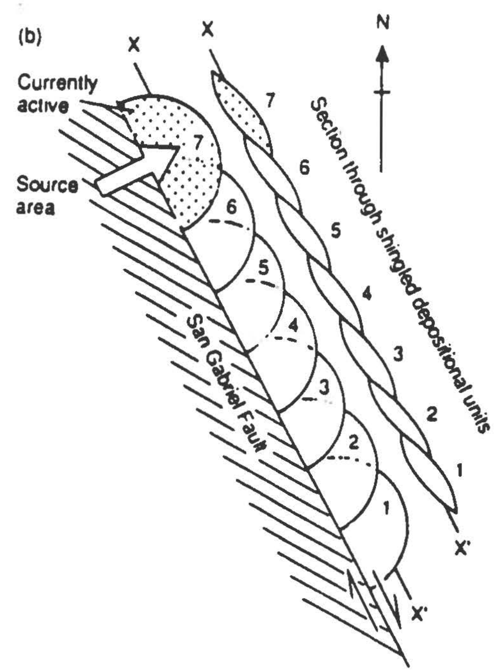


Fig. 12.10 Continued.

Fig. 12.11 Diagrammatic sketches of Ridge basin showing: A) its tectonic origin (after May et al., 1993); B) sedimentation model that gives rise to shingled arrangement of basin-margin strata (Violin Breccia)(Reproduced with permission from Crowell, 1982b); and C) arrangement of basin axial strata, parallel to depositional trough. Depocenter remained fixed relative to source area(s) across from San Gabriel fault as previously deposited strata were rotated and carried relatively south. Upper and center cross sections depict scheme at early and late stages, whereas lower cross section is diagrammatic depiction of present pattern of exposures after uplift and erosion. Total stratigraphic thickness (A+B+C . . .) is much greater than maximum thickness in any single well (from Crowell, 1982a).





La González Basin, Venezuela

La González basin is located along the Boconó fault, a late Tertiary (probably Pliocene) to recently active right-slip fault that is 500 km long and has a displacement of less than 100 km (Schubert, 1980, 1984, 1988). The fault-bend basin is about 40 km long and 5 km wide; 7 to 9 km of right-lateral slip and at least 1 km of vertical separation in middle to late Pleistocene time caused the basin to form (Fig. 12.12). The upper Pliocene to upper Pleistocene

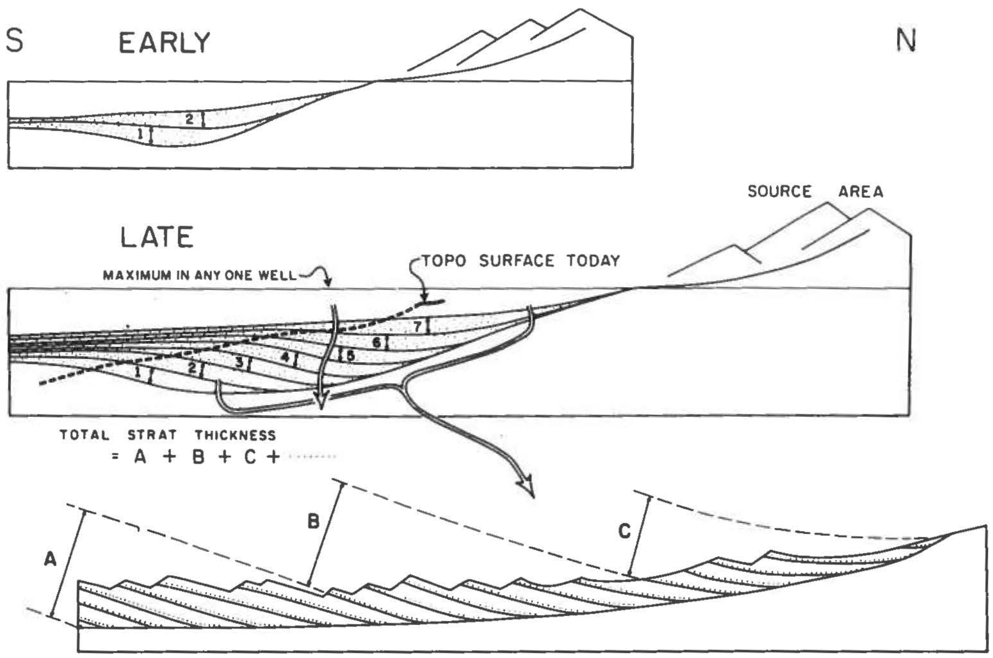


Fig. 12.11 Continued.

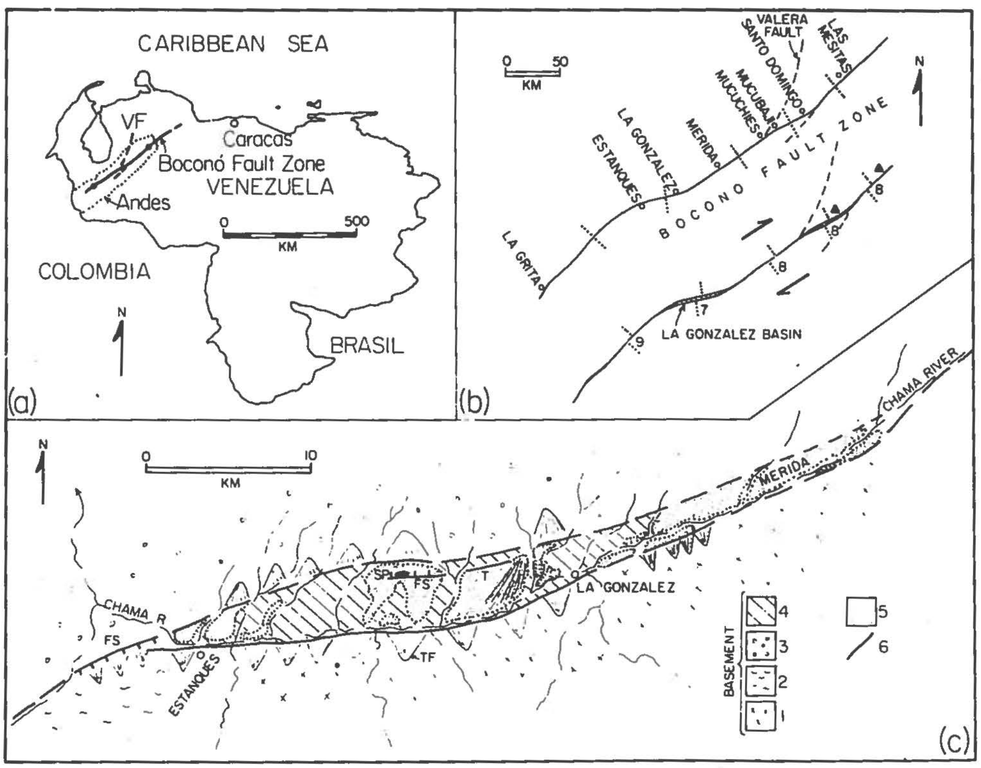


Fig. 12.12 Tectonic framework of La González basin, Venezuela (Reproduced with permission from Schubert, 1980). a) Index map. b) Fault-bend basins along Boconó— fault trace. c) Geologic sketch map of basin. 1: Precambrian; 2: Paleozoic; 3: Paleozoic-Mesozoic; 4: sheared fault zone; 5: Quaternary alluvial sediments; 6: fault trace, broken where uncertain. Abbreviations: T, fault trench; FS, fault scarp; SP, sag pond; TF, triangular facet.

fluvial basin fill overlies pre-Tertiary rocks, and the modern basin floor lies more than 2 km below the surrounding mountain ranges (Schubert, 1984). The fluvial basin fill can be divided into four climatic-tectonic sequences and is at least several hundred meters thick (Schubert, 1992); it has been transported into the basin from both flanks. The modern drainage is primarily axial to the west-southwest.

Stepover Basins

The geometric patterns, structural axes, and depocenters of stepover basins depend on the spacing between overstepping or en-echelon strike-slip faults, the amount of overlap between the faults, and the depth to basement (Rodgers, 1980). Where basement is shallow, small subbasins develop. Where basement is deep, a single, central basin may form where the overlap is slight; with increasing overlap, two or more subbasins may form. Several individual basins between overstepping strike-slip faults may coalesce into a single, larger basin, such as Binchuan basin in Yunnan province of southern China (Deng et al., 1986).

North China Basin

The North China basin is a large (~2 x 10^5 km^2) strike-slip basin that formed transtensionally in an intracratonic setting at an overstep in a large-scale right-slip system of nonparallel, right-stepping, northeast-striking faults (Nábelek et al., 1987; Chen and Nábelek, 1988). Localized and rapid post-mid-Pliocene subsidence resulted from right slip on the master bounding faults, as well as interplay of numerous other types of faults of different scales in the interior of the basin (Fig. 12.13).

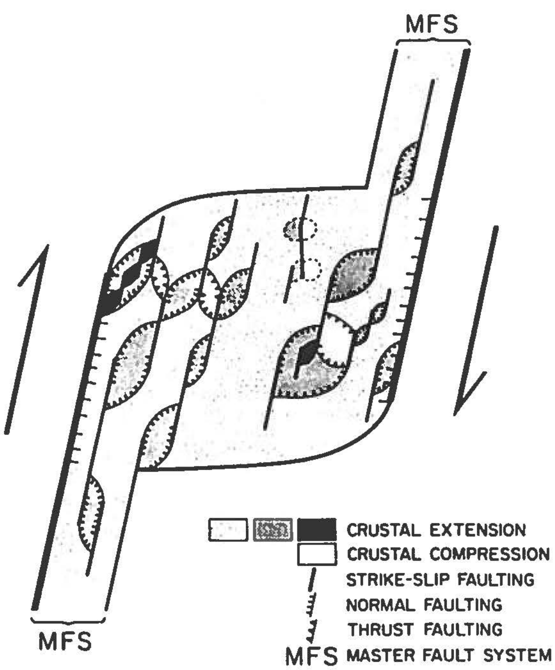


Fig. 12.13 Idealized sketch of postulated process of formation of North China basin (Reproduced with permission from Nábelek et al., 1987). Basic geometry of basin is repeated on several scales within basin. Regions of uplift are present locally, either as a result of small bends along strike-slip faults or due to interaction between neighboring strike-slip faults. Complex block rotations may have also occurred. East-west width of North China basin is approximately 600 km.

The basin began to form in early to middle Eocene time. Several fault sets divide the extensively fractured underlying basement rocks into rhombic and triangular blocks; these pre-existing faults strongly influenced the locus and sequence of numerous large earthquakes along the north margin of the basin in 1978 (Nábelek et al., 1987).

Three major earthquakes occurred in the 1978 Tangshan sequence. The epicenter of the main earthquake was at the junction of two strike-slip faults, a northnortheast-striking right-slip fault and an east-northeast-striking left-slip fault having a significant component of thrusting. One of the two strongest aftershocks occurred at the northeast end of the main fault in a right stepover in the northnortheast-striking fault system; its nodal planes strike east-west, its focal mechanism was almost purely normal, and almost 1 m of subsidence occurred here during the earthquake. The earthquake sequence demonstrated that pull-apart subsidence caused by large faults can overprint displacements on smaller ones, producing subsidence on a regional scale.

Dead Sea Basin

The Dead Sea basin is a large (132 x 16 km) and deep (8–10 km) symmetric graben (Fig. 2.14A) formed by divergent left slip along an active fault system that extends from the Gulf of Aqaba to southeastern Turkey (ten Brink and Ben-Avraham, 1989). The basin is located along the transform plate boundary that separates Arabia from Africa, and it opened in response to Miocene differential extension between the Gulf of Suez and the northern Red Sea. The total of about 105 km of left slip along the Dead Sea transform probably started in middle Miocene time (15.5–11.5 Ma); slip of only 30 km has occurred in the last 5 my, implying a low relative plate velocity (6 mm/y) between Arabia and Africa.

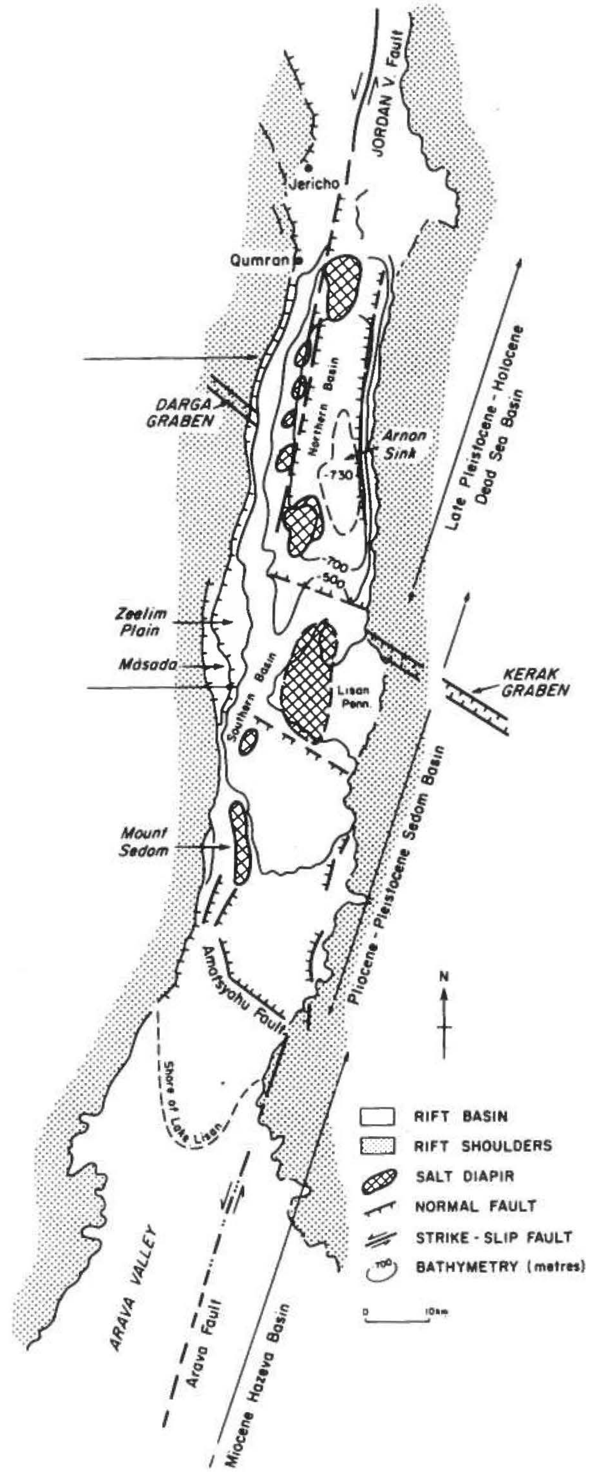


Fig. 12.14 Geologic setting of Dead Sea basin. A) Geologic map of Dead Sea-Arava depression, illustrating tectonic elements, physiographic features, northward migration of depocenter, and basin asymmetry expressed by bathymetry (Reproduced with permission from Manspeizer, 1985). Early Miocene (25–14 Ma) strike slip of about 60–65 km opened Arava basin that filled with about 2 km of red beds during a pause in strike-slip displacement. Later displacement in last 4.5 my allowed deposition of more than 4 km of marine to lacustrine rock salt of the Sedom Formation, overlain by 3.5 km of lacustrine evaporitic carbonate and clastic sedimentary strata. B) Schematic block diagram of Dead Sea region, viewed to north, with generalized rift physiography, tectonic framework and depositional domains (Reproduced with permission from Manspeizer, 1985). Bedrock of different ages on rift shoulders, basin bathymetry and eastward thickening of strata illustrate tectonic asymmetry.

The shallow basement rocks consist of massive and rigid crystalline plutonic and metamorphic rocks that are overlain by a thin cover of sedimentary strata (Fig. 12.14B). The basin is bounded longitudinally by an overstepping pair of vertical left-slip faults, and transversely by several sharply defined faults that structurally separate the basin into several 20 to 30 km segments (Fig. 12.14A). Deformation occurs mostly along the strike-slip faults, whose surface traces extend discontinuously along the length of the basin, and to a lesser degree on transverse faults. The intervening sedimentary strata are relatively undeformed due partly to the presence of a salt layer beneath much of the basin that decouples the sedimentary fill from the basement (ten Brink and Ben-Avraham, 1989).

Sediment accumulation and subsidence rates of the Dead Sea basin are high, with more than 7 km of upper Cenozoic marine, lacustrine, evaporite and fluvial-deltaic deposits (Manspeizer, 1985; ten Brink and Ben-Avraham, 1989). Subsidence of the Dead Sea basin probably began in Pliocene time, when much of the evaporitic succession was deposited, together with local marine strata. Subsidence accelerated in Pleistocene time when several kilometers of lacustrine, fluvial and other terrestrial sediments accumulated. The depocenter has migrated northward, accompanied by diapiric intrusion of evaporites, generating an asymmetric basin with extraordinarily abrupt facies changes.

The basin is narrowest and shallowest toward its northern and southern ends, and widest and deepest in the central part. Gravity data suggest that the basin sags toward its center and lacks northern and southern boundary faults (ten Brink et al., in press). The basin formed and continues to form by crustal extension along its long axis (Fig. 12.14B); maximum extension is concentrated in the central part, where it widens by passive collapse of its flanks between the two overstepping strike-slip faults that bound the

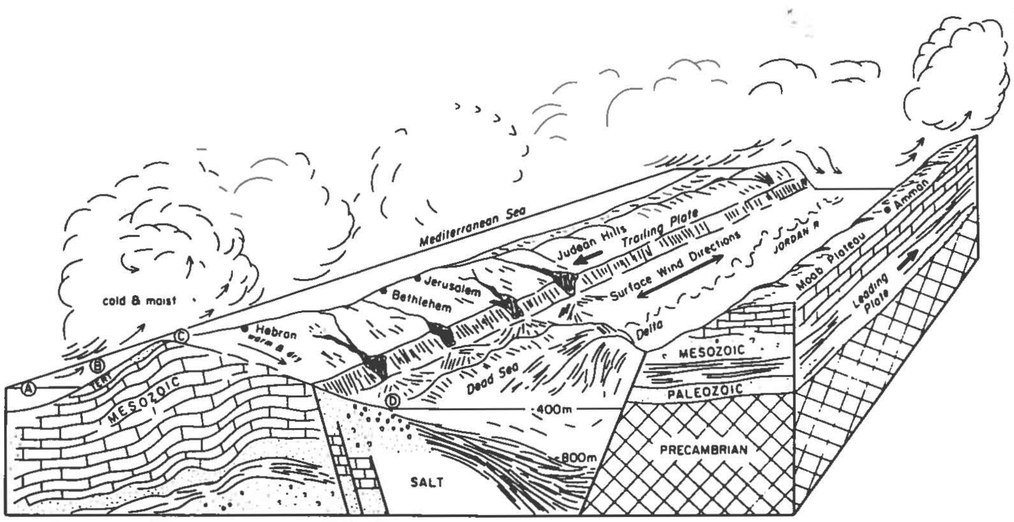
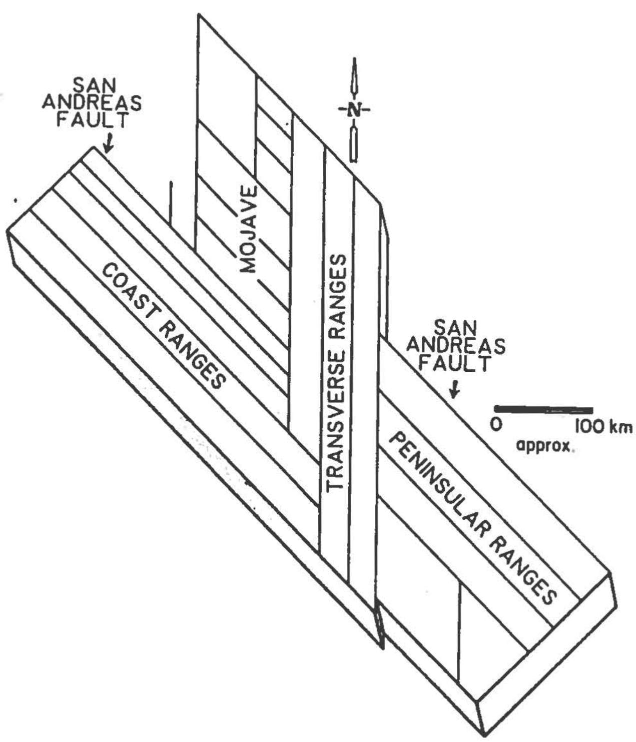


Fig. 12.14 Continued.

basin. The basin lengthened northward by simultaneous northward propagation of the southern strand of the Dead Sea fault and a retardation of the northern strand (ten Brink and Ben-Avraham, 1989; ten Brink et al., in press).



Transrotational Basins

Fault-bounded transrotational basins develop between blocks that rotate differentially in a shear zone rather than along a single fault strand (Fig. 12.3C). Rotation of blocks about vertical or subvertical axes yields prominent triangular or rhomb-shaped basins that may be quite large in very wide zones of distributed

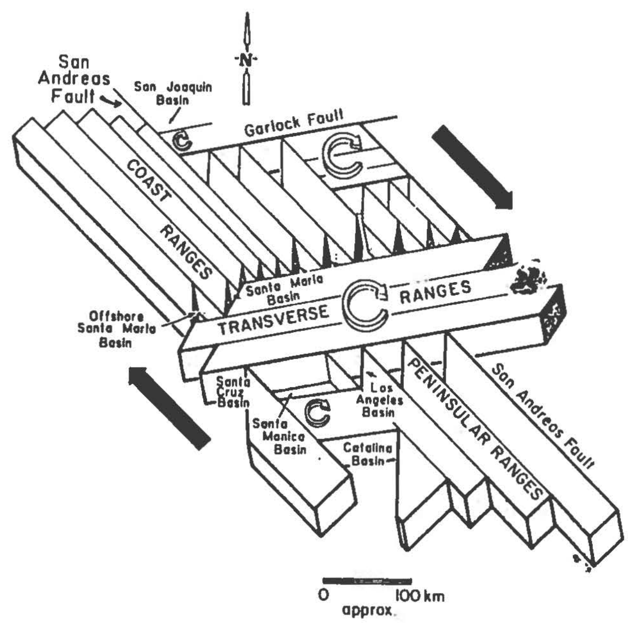


Fig. 12.15 Rotation model for southern California. See Luyendyk and Hornafius (1987) for discussion and details of faults used in model. A) Inferred initial fracture pattern and prerotation geometry during Oligocene. B) Late Miocene geometry after clockwise rotation, which occurred mainly during Middle Miocene. Sinistral and dextral slip has occurred, and deltoid basins have opened at joins of rotated and unrotated blocks. (Reproduced with permission from Luyendyk and Hornafius, 1987.)

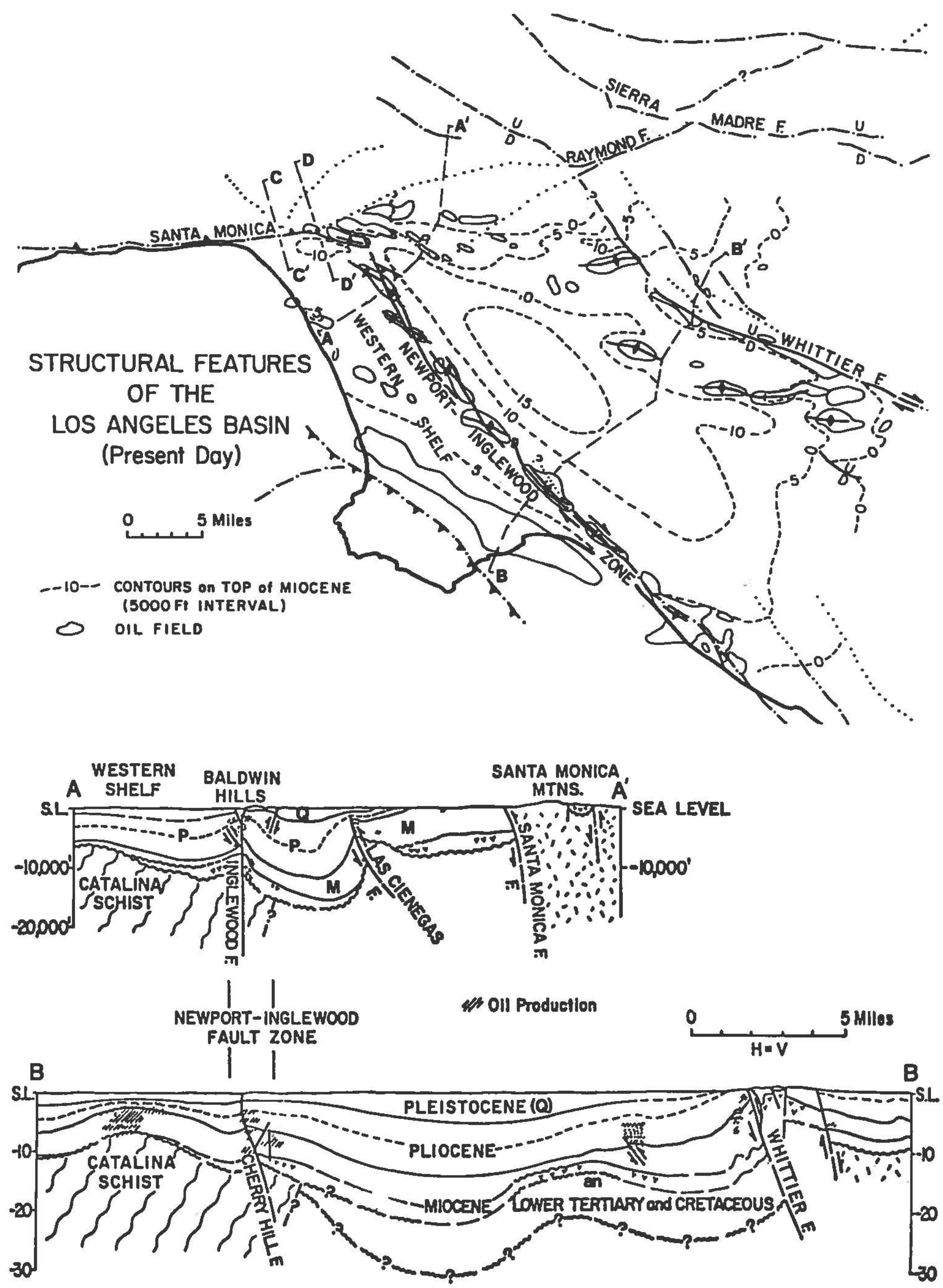


Fig. 12.16 Generalized structural map (A) and regional cross sections (B) of Los Angeles basin. Abbreviations: M, upper and upper-middle Miocene; P, Pliocene; Q, Quaternary; an, Anaheim nose. Vertical scale of (B) is in kilofeet. (Reproduced with permission from Wright, 1987.)

shear, such as the rhomb-shaped Bir Zreir graben in eastern Sinai, formed by rotation and consequent subsidence between two bent strike-slip faults of the Dead Sea Rift zone (Eyal and Eyal, 1986). Several southern California Neogene basins appear to have formed during clockwise rotation of as much as 90° during Neogene time near the margins of the east-trending Transverse Ranges and in the Mojave Desert (Fig. 12.15) (Luyendyk and Hornafius, 1987; Luyendyk, 1989).

Los Angeles Basin, Southern California

The Los Angeles basin (Fig. 12.16) is regarded by some writers as a transrotational basin, formed by differential clockwise rotation of blocks in a wide zone of right simple shear (Luyendyk et al., 1980, 1985; McKenzie and Jackson, 1986; Luyendyk and Hornafius, 1987; Luyendyk, 1991). Other writers (Crouch and Suppe, 1993) have proposed that it formed by large-magnitude crustal extension in early Miocene time (25–22 Ma), and after its Miocene and Pliocene sedimentary succession accumulated during rapid rifting from 16 to 5 Ma, it was deformed by late Pliocene and younger right-slip faults.

The roughly triangular basin is approximately 40 km wide and 60 km long (Fig. 12.16A). It consists of several triangular sub-blocks and basins bounded by fault zones that also control the main hydrocarbon-bearing structural traps in the basin. Many of the faults are active, judging from historic earthquakes, as well as from abundant geologic evidence of recent deformation. Along its north flank, a series of north-dipping Quaternary thrust faults carries crystalline basement of the Transverse Ranges southward over the basin (Dibblee, 1982; U.S. Geological Survey, 1987). The east and southeast flanks of the basin consist of a faulted shelf of deformed Upper Cretaceous and lower Tertiary strata that rest on Mesozoic sialic basement overlain by Oligocene, Miocene and Pliocene sedimentary rocks, some of which are faulted and intruded by Miocene igneous rocks. The Elsinore-Whittier right-slip fault system extends northwestward from the Peninsular Ranges into the eastern part of the Los Angeles basin, where it must intersect somehow in the subsurface with the northern basin-margin thrust faults. The southwest side of the basin is bounded by the northwest-striking Newport-Inglewood fault zone. Crouch and Suppe (1993) suggested that the Newport-Inglewood fault zone was the locus of the main extensional breakaway for the Early Miocene rifting that formed the basin. Interpretations of geologic, seismic and seismic-reflection data indicate that the entire basin is underlain by an active, blind, south-dipping regional décollement, which with its imbricate thrusts, ramps and folds, has accommodated 15 to 30 km of north-south shortening in the last 2 to 4 my (Davis and Namson, 1989). Many modern topographic highs and lows in the basin correspond to structural highs and lows, and, together with deep thrust earthquakes, attest to continuing deformation.

The Tertiary sedimentary succession in the Los Angeles basin is dominated by Neogene deep-marine turbidites and hemipelagic units that contain the greatest concentration of petroleum per unit area of any basin in the world (Biddle, 1991). Miocene strata include organic-rich, siliceous shale of the Monterey Formation and related units, which are the primary source rocks. Miocene volcanic rocks are present at depth nearly everywhere in the central part of the Los Angeles basin, but crop out only locally along its edges. This implies that magma intruded in response to cracking of the crust and opening of the basin, and it rose nearly hydrostatically to a compensating level within the sedimentary fill (Wright, 1991). Some of the volcanic rock units are also rotated clockwise, based on interpretations of paleomagnetic vectors (Hornafius et al., 1986), supporting the conclusion that transrotational deformation, in part, followed Early Miocene formation of the basin by rifting and detachment faulting (Crouch and Suppe, 1993).

Thick strata consisting of thin-bedded Pliocene marine turbidites, derived largely from sources north and east of the basin, are the principal reservoir rocks for petroleum. The Pliocene and Quaternary clastic marine and nonmarine strata in the central part of the basin are nearly 4 km thick, based on drilling and seismic-reflection studies (Blake, 1991; Mayer, 1991; Wright, 1991) (Fig. 12.16B).

Mid-Pliocene (younger than 5 Ma) transpression may have caused the Newport-Inglewood fault zone to be reactivated as a right-slip fault (Crouch and Suppe, 1993). On the sea floor above the fault, right-stepping, en-echelon, anticlinal hills formed (Yeats, 1973) that guided turbidity currents down the axes of synclinal troughs. Continued folding of the anticlines tilted the edges of the sand-filled synclines and formed some oil traps (Harding, 1973). Extensive folding and faulting also generated abundant structural traps in and along the margins of the basin (Wright, 1991). Now the basin is filled so that alluvial fans stretch from the bounding mountains to the north across the basin almost to the sea, and surface streams carry sedimentary debris to offshore basins.

Many modern topographic highs and lows correspond to structural highs and lows, attesting to the recency of deformation.

Mojave Desert Basins, Southern California

The Mojave region experienced a major episode of uplift, extensional rifting, and detachment faulting during Early Miocene time, followed by regional, northwest-southeast simple shear that caused irregularly shaped blocks to rotate differentially about subvertical axes (Dokka and Travis, 1990). Several basins formed between misfit areas among the faulted pieces, edges and corners of the rotated blocks. More than 1000 m of Miocene and Pliocene alluvial, fluvial, lacustrine, volcaniclastic and evaporite deposits accumulated in the largest of the basins, the Barstow basin, which is 100 km long and 20 km wide (Link, 1980; Woodburne et al., 1990). These deposits are moderately folded and faulted, but little rotated (MacFadden et al., 1990). Early Middle Miocene evaporite deposits in Kramer basin (Dibblee, 1980) contain 100 m of sodium borate (borax), one of the largest deposits in the world. It was deposited mostly from hot springs into an ephemeral playa lake (Siefke, 1980) that formed along a minor right-slip fault (Dibblee, 1967).

Transpressional Basins

Narrow basins that are marginal to transpressional principal displacement zones or positive flower structures may result from flexural subsidence induced by loading from branching reverse faults. This flexurally induced subsidence is similar to the mechanisms of foreland-basin development, and these marginal strike-slip basins are, in fact, miniforeland basins that may be overidden by advancing allochthons. Small piggy-back basins are also common in these settings (e.g., Namson and Davis, 1988). Rapid sedimentation in transpressional basins induces additional subsidence by sediment loading.

Sediments deposited in transpressional basins in deep-marine settings along transform faults may consist mostly of pelagic or hemipelagic mudstone with some proximal landslide, debris-flow, volcaniclastic or talus deposits. Where deep-marine transpressional basins are situated close to larger landmasses, thick turbidite successions may be deposited. In nonmarine settings, alluvial-fan, lacustrine, fan-delta, delta, fluvio-deltaic and turbidite facies are most common. Most sediment is derived from the adjacent transpressional uplift blocks, but sediment may also be derived from more distal sources.

In more complex transpressional regimes, marked by multiple faults, fault strands and interfault basins, transpressional basins may originate from synclinal downfolding or fault-dominated subsidence marginal to reverse faults. These types of transpressional basins have more complex patterns of sedimentation and structural evolution. Individual transpressional basins may have multiple sources of sediment; longitudinal, rather than transverse sediment transport directions, may predominate. We discuss two California basins as examples of transpressional basins, the Ventura and San Joaquin basins. It is clear, however, that both basins have had long and complex histories and should be grouped with polyhistory basins. Here, we focus only on their last stages of evolution, which have been transpressional.

Ventura Basin, Southern California

The Ventura basin of southern California is located in the southwestern part of the Transverse Ranges province (Fig. 12.17) (Crowell, 1976). The west-trending and west-plunging narrow basin developed in Early Miocene time (about 22 Ma) in response to 90° clockwise rotation of the Transverse Ranges province (Luyendyk et al., 1980, 1985; Luyendyk, 1989). Although early evolution of the basin was transtensional or transrotational, its later history has been dominated by transpression (Crowell, 1976). As the province underwent rotation, extensive left-lateral faults sliced the block into a series of sub-blocks, some of which underwent uplift, and others of which underwent subsidence in response to rotation-induced shear.

Like the nearby Los Angeles basin, deposition commenced in Miocene time (14–6 Ma) with diatomaceous and phosphatic mud (now the Monterey Formation) in protected and moderately deep water. These deposits lapped across a subsea shoulder or ridge with seaknolls on it that was uplifted into an

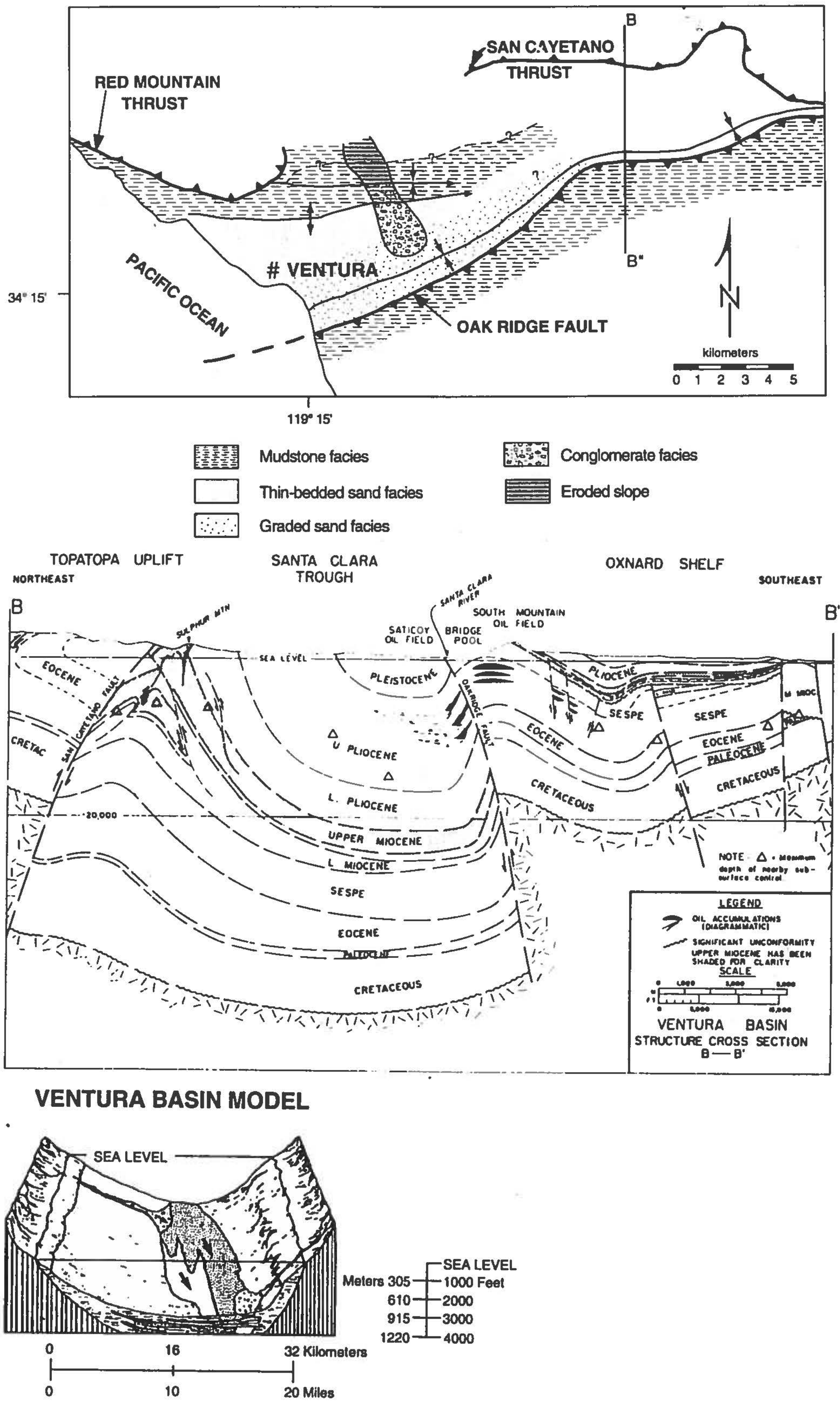


Fig. 12.17 Generalized map, cross section and depositional model of Ventura basin, California. A) Facies distribution of Pliocene and Pleistocene deep-sea sedimentary rocks (modified from Hsü et al., 1980); B) Structure section B-B′ (from Nagle and Parker, 1971); C) Process model of turbidity-current deposition in deep-sea basin of Ventura type (modified from Hsü et al., 1980).

antiformal arch along the length of the basin. Major subsidence occurred in Early Pliocene time, and one of the thickest Pliocene bathyal marine sequences in the world was deposited, with sedimentation rates exceeding 2 mm/y. Locally, basin inversion has occurred during the past one-half million years.

The thick Pliocene deposits consist of sandy siliciclastic turbidites transported axially down the basin from basement source areas more than 50 km to the east (Hsü, 1977; Hsü et al., 1980). These deposits form spectacular petroleum reservoirs in prolific oil fields along the axes of inversion anticlines. Coarser-grained submarine debris flows were shed into the basin from the subparallel north and south flanks of the basin. Basin inversion also caused uplift of older rocks along basinward directed reverse faults along the north and south flanks of the basin (Fig. 12.17B).

The entire western Transverse Ranges province has been interpreted as a fold-and-thrust belt by some workers (Namson and Davis, 1988). The abundant left-lateral faults, rotation history, and spatial association of the basin with the San Andreas fault suggest to us that it forms a type of fold-and-thrust belt within a wide zone of transpressional deformation. The Pliocene to Holocene Ventura basin formed within this zone as a complex transpressional basin characterized by rapid subsidence and high rates of sedimentation typical of strike-slip basins.

San Joaquin Basin, Central California

The late Cenozoic San Joaquin basin in central California (Bartow, 1991) developed in the southern part of the Great Valley, a forearc basin that originated in Mesozoic time as a result of lithospheric plate convergence (see Chapter 6). The San Joaquin basin is a deep (12 km), irregular-shaped basin that began to subside rapidly in Late Oligocene and Early Miocene time in response to transpression extension (Goodman and Malin, 1992). Subsequent middle Miocene to Holocene subsidence and basin-margin uplift have been partly in response to associated with the San Andreas fault (Wilcox et al., 1973). Subsidence increased at about 16 Ma, with deposition of phosphatic shale and diatomaceous strata, the principal source rocks for petroleum.

The southern and southwestern flanks of the San Joaquin basin are bounded by transpressional regions of the San Andreas fault (Fig. 12.18; Wilcox et al., 1973; Harding, 1976). The San Emigdio Range and Tehachapi Mountains along the south flank of the basin are being thrust northward over the basin along the Pleito thrust and along several buried thrusts. The Temblor Range along the southwestern flank is being thrust northeastward over the basin along mostly buried thrust faults (Webb, 1981; Namson and Davis, 1988; Medwedeff, 1989; Ryder and Thomson, 1989). Marginal to the thrusts is a series of recent lacustrine basins, including, from southeast to northwest, Kern, Buena Vista, and Tulare lakes.

Structural mapping of basinwide lacustrine and volcanic ash units within the Tulare Formation, a widespread and thick Pliocene to Holocene siliciclastic unit in the San Joaquin Valley and surrounding San Emigdio and Temblor Ranges, reveals that these units have been structurally and topographically depressed adjacent to the allochthonous masses. The lacustrine transpressional basins form mini-foreland basins depressed by structural loading and resultant flexural bending. Some workers attribute the convergent foldthrust belt entirely to thrust tectonics (Namson and Davis, 1988; Medwedeff, 1990). Together with Harding (1973), however, we maintain that the transpressional setting is a consequence of partitioned convergent strike slip.

The San Joaquin basin originally opened westward toward the Pacific Ocean in early Miocene time (20 to 16 Ma); it was subsequently truncated, probably rotated and offset about 300 km by the San Andreas fault. The offset western part of the Early Miocene basin lies in the Santa Cruz Mountains (Bartow, 1991). During Miocene time (12–7 Ma), the block southwest of the San Andreas fault was uplifted and provided detritus to fan deltas and turbidite systems into the deep basin (Ryder and Thomson, 1989; Fig. 12.18). Prior to that, most of the sediment came into the basin through submarine canyons from the Sierra Nevada region to the east and flowed northward down the axis of this mid-Miocene trough. At the same time, however, en-echelon folds were growing and forming submarine hills along the west edge of the basin, partly as a result of shear along the San Andreas fault. The growing folds focused turbidity currents along synclinal valleys and through saddles to basin plains. The basin became isolated from the Pacific at the beginning of the Pliocene and has filled with nonmarine deposits of Pliocene and Quaternary age.

About 4 Ma, motion along the San Andreas transform changed from predominately transtensional to transpressional. The consequent shortening caused blind thrusts to develop along the west edge of the basin, further deforming the early en-echelon anticlines. The thrust faults and anticlines are active today.

Polygenetic Basins

In divergent settings, strike-slip basins may develop along transfer or accommodation zones that link major loci of normal faults. They may also form along the flanks of individual detached crustal segments, especially in horseshoe faults, where low-angle strike-slip faults mark the margins of detached segments (Fig. 12.3E). In convergent settings, strike-slip basins develop along transfer zones that link intact individual segments of larger thrust or underthrust crustal segments (Fig. 12.3E).

Death Valley Basin, Eastern California

Death Valley is a north-trending, asymmetric half graben that has formed as a result of regional extension (Fig. 12.19). Although extension commenced about 15 Ma, the present physiography of the valley has developed in the last 4 my. The east side of the half graben is bounded by a major range-front fault zone with oblique normal slip, the Black Mountains fault zone. Structural relief exceeds 6,000 m across the Black Mountains fault zone, as measured from the base of the sedimentary section in the deepest part of the half graben to the top of basement rocks in the adjacent bounding mountain range to the east.

Alluvial, fluvial, and lacustrine sediments exceed 3000 m in thickness on the east side of the basin next to the Black Mountains fault, and thin westward to less than a few-hundred meters. Some sediment is derived from the bounding mountains and transported into the half graben as debris-flow-dominated alluvial fans that are small and steep on the east side, and by larger, gentler-sloping and regionally extensive stream-flow-dominated fans on the west side (see also Fig. 3.32). The axial and ephemeral Amargosa River transports northward-fining sediment into the valley from a larger area surrounding Death Valley. Evaporitic playa deposits and wind-blown dune fields are also present in the axial parts of the valley (close to the east side).

Northwest-striking right-slip faults at the north and south ends of central Death Valley are tear faults or transfer faults that separate the half graben from differentially extending tracts to the north and south (Burchfiel and Stewart, 1966). At the north end, the Furnace Creek fault extends southeastward across the north end of the valley to the Black Mountains fault zone. At the south end, the Death Valley fault zone extends northwest, cutting the southern end of the Black Mountains fault zone and the southern end of the graben, there bending northward into minor normal faults along the west edge of the valley. The amount of displacement is about 25 km along both of these fault zones.

In addition to being a pull-apart between two fault systems, as proposed originally by Burchfiel and Stewart (1966), Death Valley has been interpreted as one of several half grabens that developed locally in a regionally extended crustal slab (Wernicke, 1985; Wernicke et al., 1988; Hodges et al., 1989). The graben formed as the hanging wall extended northwestward, parallel to the strike-slip transfer faults that accommodate or transfer the extension from one half graben to another (Burchfiel et al., 1987).

Vienna Basin, Austria

The Vienna basin, one of several Miocene basins in the Carpathian Mountains of Austria and Czechoslovakia, is a rhombohedral basin along an active left-stepping, left-lateral tear fault between two thrust blocks of the Carpathian thrust belt (Fig. 12.20).

The basin is about 200 km long and 60 km wide. It is filled with Miocene shallow-marine to locally brackish clastic strata (as thick as 6 km) that were deposited during two episodes of basin subsidence; Pliocene and Quaternary fluvial deposits, as thick as 200 m, cap the Miocene sequence (Royden, 1985).

Three overstepping and northeast-striking left-slip faults controlled the early and late episodes of development of the two rhombohedral subbasins that make up the Vienna basin. The northernmost fault has accumulated about 80 km of Cenozoic left slip.

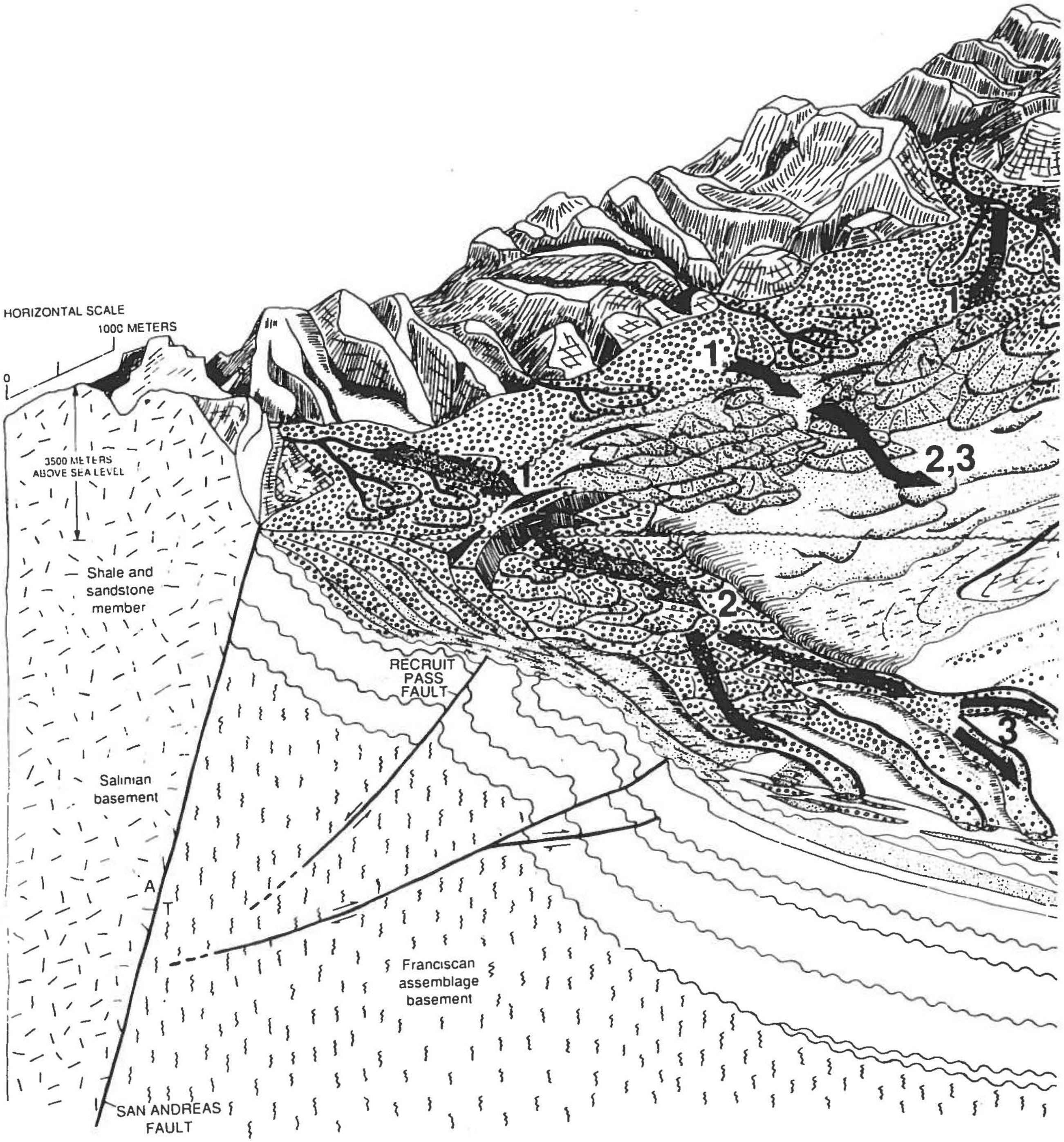
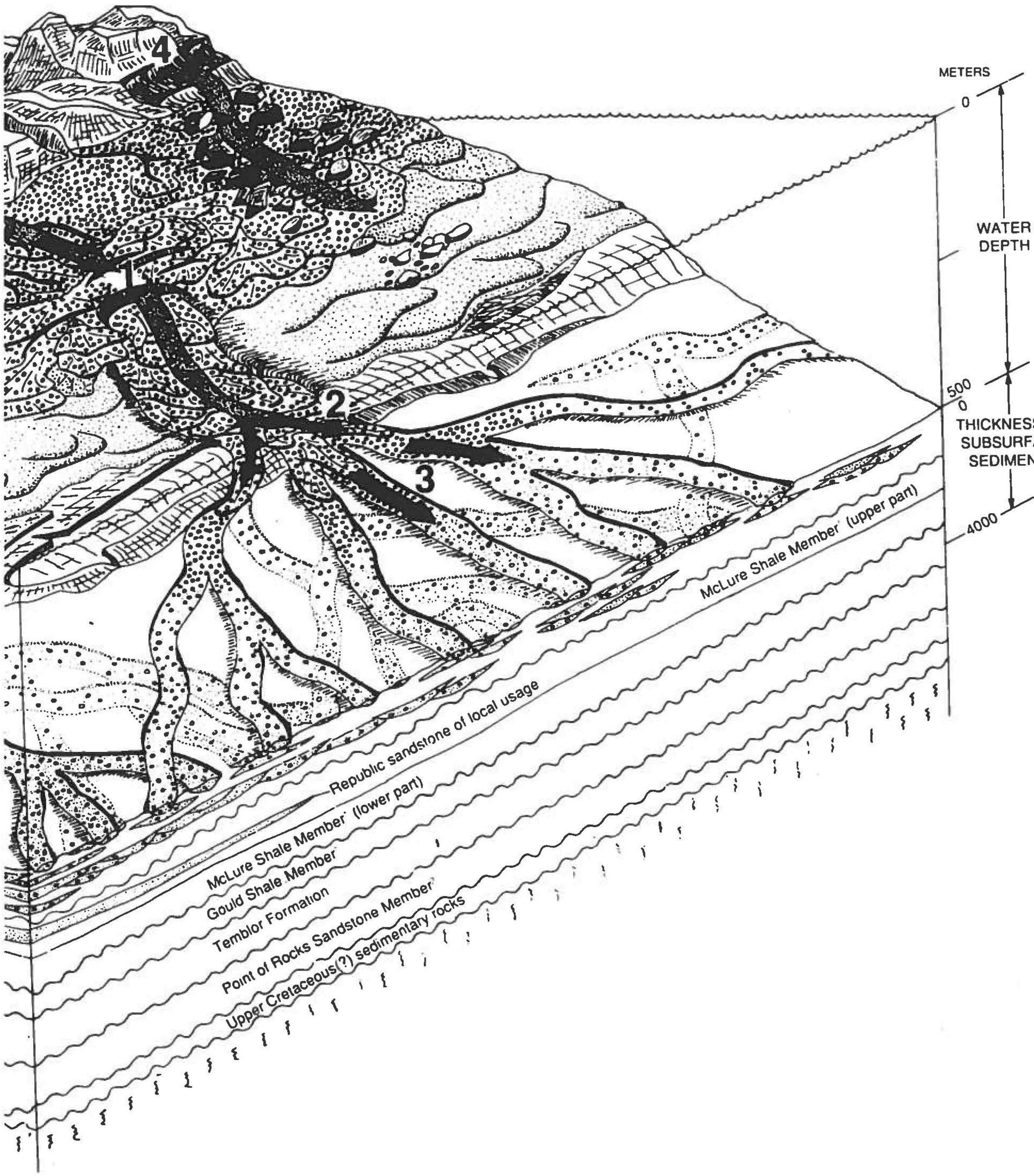


Fig. 12.18 Schematic reconstruction of depositional environments, depositional processes and paleotectonic setting of Santa Margarita Formation, southern San Joaquin basin, California. Water depth is greatly exaggerated to allow depiction of fan deltas and submarine fans.



Numerals: 1) subaerial debris-flow deposit; 2) subaqueous debris-flow deposit; 3) high-concentration turbidity-current deposit; and 4) debris-avalanche deposit. (Reproduced with permission from Ryder and Thomson, 1989.)

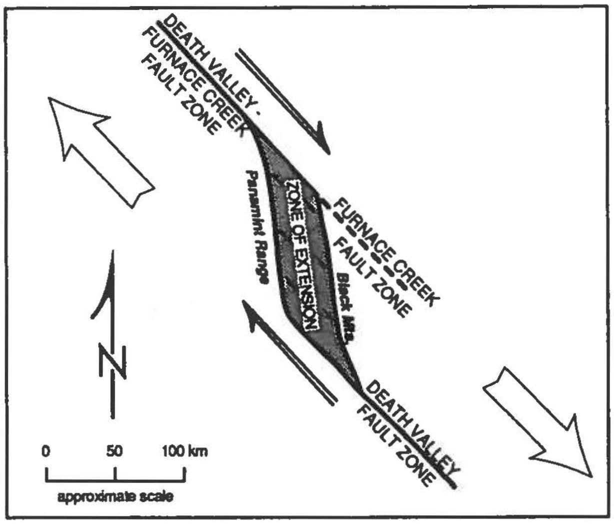


Fig. 12-19 Diagrammatic sketch map of Death Valley "pull-apart." (Modified from Burchfiel and Stewart, 1966.)

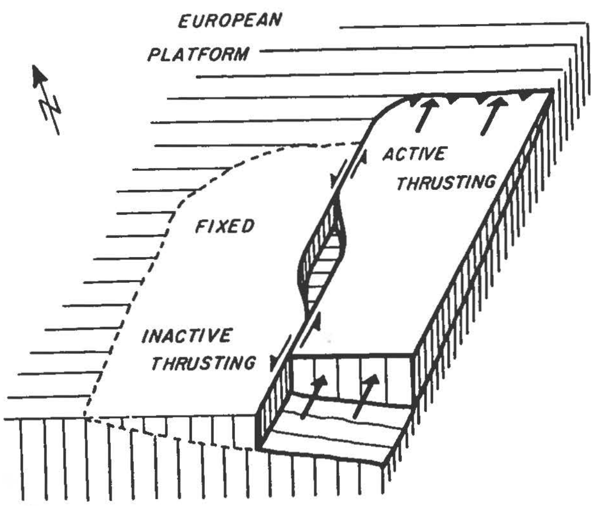


Fig. 12.20 Schematic diagram of opening of Vienna basin as a fault-bend basin at a left step in a left-slip tear fault within an allochthon thrust northeastward onto the European platform (Reproduced with permission from Royden, 1985).

The faults are roughly parallel to structural trends in the underlying nappes.

The depocenters of the two episodes of subsidence do not coincide. Subsidiary synsedimentary faults along the northwest margin of the basin dip 40° to 50° basinward and have several kilometers of normal separation at the surface. The basin fill thickens northwestward across the basin, and the older strata dip more steeply than the younger.

Left-slip faults developed contemporaneously with thin-skinned thrusts (Royden, 1985). The Vienna basin is confined entirely to the allochthon and formed along a releasing bend; therefore, in this type of polygenetic basin, mantle was not involved and a thermal anomaly is not present beneath the basin.

Polyhistory Basins

Polyhistory basins have complex structural patterns that result from alternating episodes of strike-slip, compressional, and extensional faulting. These basins commonly form in areas of changing plate-tectonic framework. They may develop as polygenetic basins along thrust-sheet boundaries in collisional orogens that are later modified by post-collisional extensional faults; in this type of setting, strike-slip basins along tear faults confined to the upper plate of thrust sheets may undergo reversal of movement along tear faults during subsequent extensional faulting. Some collapse basins, such as the strike-slip-bounded, Devonian Hornelen basin of western Norway (Steel and Gloppen, 1980), may have resulted from this process.

Bowser Basin, Western Canada

The northwestern part of the North American Cordillera contains Jurassic, Cretaceous and Cenozoic "successor" basins that record complex tectonic and depositional adjustments to the accretion and juxtaposition of tectonostratigraphic terranes. The Tyaughton-Methow, Bowser and Gravina-Nutzotin basins are three such complex successor basins that developed along the continental margin in response to accretionary events (Fig. 12.21). Oblique convergence and transform motion of oceanic and arc-type lithospheric terranes has generated a wide region of generally right-lateral displacement (Eisbacher, 1981, 1985).

The Bowser basin originally developed in Late Jurassic time above and marginal to the belt of faults along which the Stikine terrane accreted to North America. The Stikine terrane, primarily made up of an early Mesozoic calc-alkaline magmatic-arc complex, subsided by Late Jurassic time to form the basement of the Bowser basin. Initially, Bowser basin sediments were transported westward into the basin from uplifted, previously accreted terranes to the

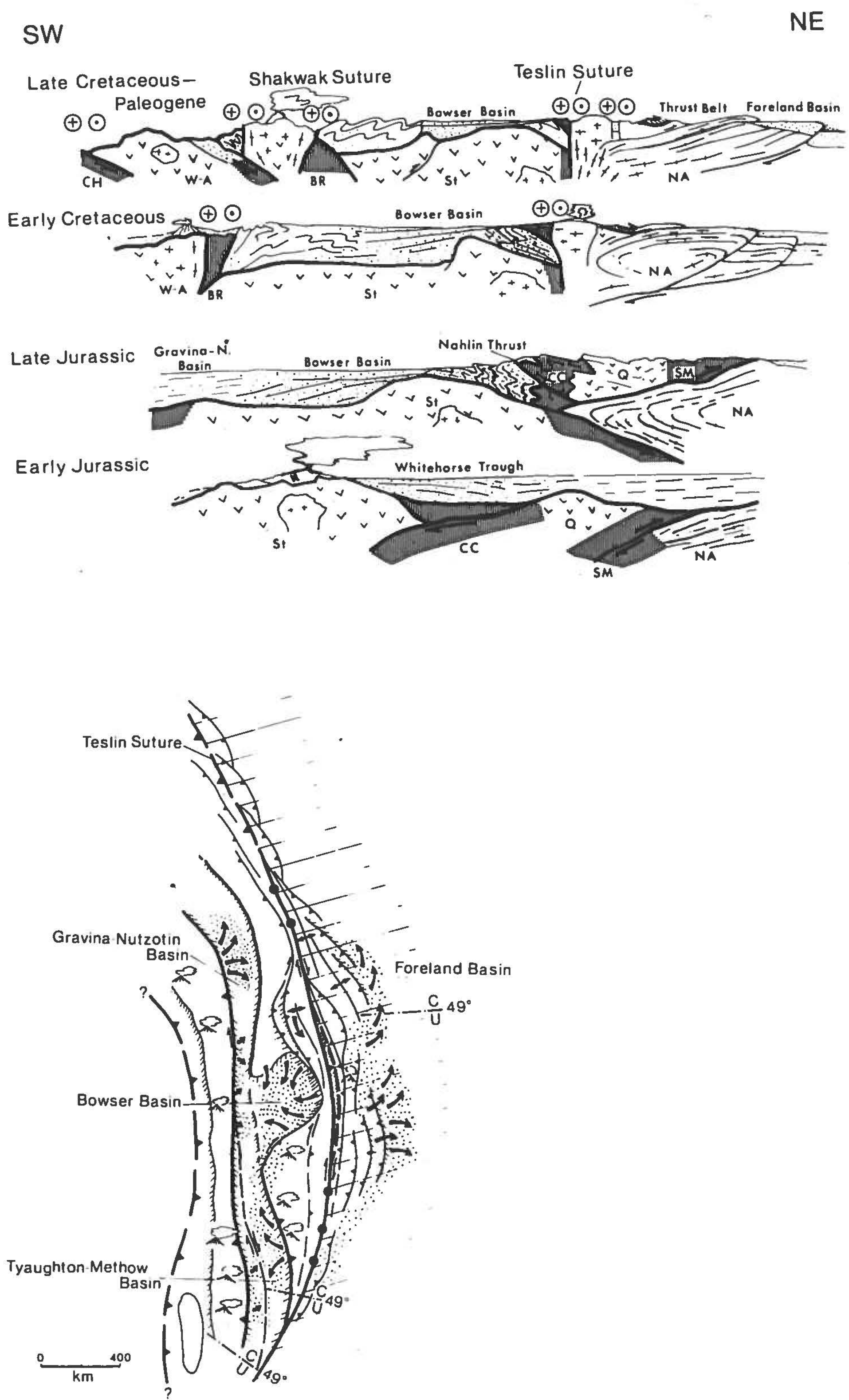


Fig. 12.21 Cross sections and map illustrating tectonic evolution of Bowser basin. A) Schematic evolution of peri-collisional clastic basins and strike-slip faults in a southwest-northeast transect through north-central Canadian Cordillera. Vertical ruling indicates oceanic-type terranes and ophiolitic mélanges (SM, Slide Mountain terrane; CC, Cache Creek terrane; BR, Bridge River terrane; CH, Chugach terrane). V-pattern indicates island arcs and oceanic plateaus (Q, Quesnel terrane; St, Stikine terrane; W-A, Wrangell-Alexander terrane). NA, North American craton. Cretaceous-Paleogene right slip developed along zones of complex crustal convergence characterized by multiple changes in structural vergence and subsequent pluton activity. B) Paleogeographic model of Middle Jurassic to Early Cretaceous setting of clastic basins near Teslin suture. Diagonal ruling indicates extent of North American continental crust locally overlapped in west by ophiolitic thrust masses of closing Teslin suture, and characterized by uplift of metamorphic core complexes. Accreted terranes west of thrust belt are shown approximately 1500 to 2000 km south of their present positions with respect to North America (49th parallel is used as displaced marker line). Strike-slip faults and arc polarity west of marginal Gravina-Nutzotin-Bowser-Tyaughton-Methow troughs are speculative. Areas of volcanic activity and generalized sediment dispersal depicted by cones and arrows, respectively. (Reproduced with permission from Eisbacher, 1985.)

east; more than 3,000 m of mid-Jurassic to Upper Cretaceous submarine-fan, slope and fluvial-dominated deltaic deposits accumulated in the basin. There is little direct evidence during this early history of the basin for major strike-slip involvement; basin subsidence and basin-margin uplift appear to have resulted primarily from convergence.

Later in the Early Cretaceous, however, right-slip faulting began to dominate. The eastern flank of the basin underwent uplift and transpression along the Teslin suture, an active zone of right slip that widened eastward through time. The eastern Bowser basin was tectonically reactivated, and 500 to 1000 m of generally coarse-grained fluvial and lacustrine strata were deposited unconformably on the older marine fill. Right slip and uplift along the Shakwak suture on the western flank of the basin resulted in eastward transport of nonmarine clastic detritus into the basin. As deformation continued during the Tertiary, the basin narrowed to a trough that filled with 300 to 800 m of uppermost Cretaceous to Eocene alluvial-fan and fluvial units deposited chiefly by east-flowing systems.

Although much detailed sedimentologic and stratigraphic work remains to be done in the basin, the framework established by Eisbacher (1981, 1985), as well as by workers in related basins, including the outboard Queen Charlotte basin, reflects the shifting influences of convergent and strike-slip tectonics on depositional systems and basin morphology.

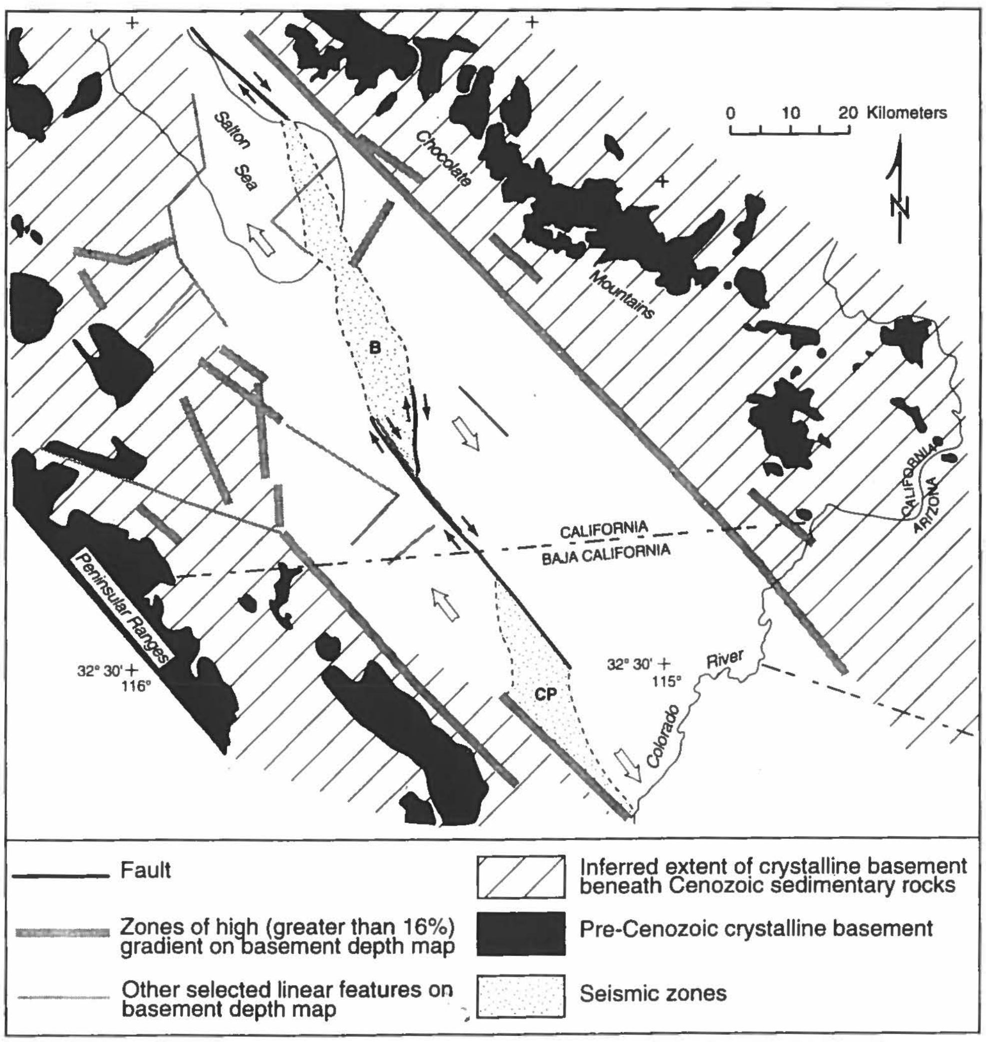


Fig. 12.22 Structural and tectonic map of Salton Trough. Seismic zones (stippled) are spreading centers (B, Brawley; CP, Cerro Prieto) that connect right-stepping right-slip faults and are probable locations of incipient stepover basins in San Andreas strike-slip system that trends obliquely up axis of Salton Trough. Direction of spreading in each stepover is indicated by open arrows. Sediment is carried into basin by Colorado River and by ephemeral drainages from surrounding mountains. (Modified from Fuis and Kohler, 1984.)

Salton Trough, Southern California

The Salton trough is an active strike-slip basin with an active oceanic spreading ridge constituting the southern part of its floor (Fig. 12.22). The trough is an elongate, northnorthwest-trending sedimentary basin that has formed adjacent to the southern terminus of the intracontinental San Andreas fault; the plate boundary passes southeastward across a series of short spreading ridges and long oceanic transform faults to the Rivera triple junction and the East Pacific Rise (see Chapter 2). Many aspects of the basin are well understood as a result of extensive subsurface geothermal exploration, frequent and large earthquakes, excellent exposures of rocks along both basin margins, and numerous studies of modern depositional systems.

Although initial formation of the basin during Middle Miocene time appears to have been the result of extensional tectonics related to formation of the Gulf of California and the Basin and Range (Crowell, 1987), the later and modern history of the basin has reflected right-slip along the San Andreas fault and related transform faults to the south. The initial record of extensional opening is well preserved in outcrops along the western margin of the trough, where pre-rift, braided-stream deposits are overlain in angular unconformity by synrift alluvial-fan, landslide-breccia, lacustrine, fluviodeltaic and evaporitic deposits. The basin in this area appears to have been a half graben, with its principal north-striking border fault to the west. Subsequent increased subsidence allowed the terrestrial succession to be overlain by widespread shallow- to deep-marine, late synrift or postrift strata of the Imperial Formation.

During deposition of the Imperial Formation, the transform boundary jumped eastward into the rift valley, thus initiating activity along the modern San Andreas fault. The later history of the basin has been dominated by strike-slip tectonism associated with several right-slip faults of the San Andreas system, including the San Andreas, Imperial, San Jacinto and Elsinore faults. Coarse-grained and very thick alluvial-fan deposits grade axially into lacustrine mud, similar to the modern setting surrounding the Salton Sea. The later history is dominated by deposition of huge volumes of sediment by the Colorado River, whose delta separates the south end of the Salton trough from the north end of the Gulf of California (Winker and Kidwell, 1986).

The Salton Trough has reached a stage in its tectonic history that involves uplift and arching above an en-echelon chain of hot mantle domes, judging from Quaternary volcanism within the trough, high heat flow, high Bouguer gravity values, and greenschist-facies metamorphism of Pliocene and Pleistocene sediments at depths reached by geothermal drill holes (4000–6000 m). These phenomena are associated with oblique sea-floor spreading and opening of the Gulf of California. Uplift of the mountainous margins may be attributed to thermal expansion of the lithosphere beneath a region wider than the Salton trough itself. Oblique stretching of the region follows the strike of transform faults and agrees with the plate motion between the Pacific and North American lithospheric plates. Faults and deformation patterns outline uplifted blocks and subbasins nested within the Salton Trough. The subbasins are similar to those strike-slip basins in the Gulf of California that have not filled with sediments; orientation of the small basins relative to the strikes of the major transform faults and the plate boundary suggests that their direction of stretching corresponds to the direction of extension in right simple shear (Crowell, 1985).

The Salton Trough reflects the complex evolution of a polyhistory basin with an initial rift origin, an irregularly stretched floor, strike slip, development of en-echelon subbasins, and rerouting of a major fluvial-dominated delta into the widening trough. Tectonic mobility of the Salton trough began about 12 Ma and has intensified during the last 5 my. Modern strike-slip faulting is superposed on Miocene detachment faults which were superposed on Mesozoic and early Tertiary thrusts.

Conclusions

We have described six major types of strike-slip basins based on their fault patterns and mechanisms of formation; however, strike-slip basins constitute the most complicated of basin types. They form in diverse tectonic settings, and are deformed and reformed as fault blocks rise, fall, converge, and diverge in space and time. Their form, structure, depositional setting, and history depend upon the complex interplay of structural, depositional, climatic, and paleogeographic factors.

Most long-lived strike-slip basins are polycyclic and undergo repeated episodes of generally transtensive and transpressive subsidence and uplift within continuing strike-slip settings. Repeated subsidence and uplift commonly yield complex basins that may be difficult to reconstruct, however, and each basin must be evaluated separately.

Laterally displaced source areas, facies, and depocenters provide general criteria for the recognition of areas of strike-slip faulting. Complex and abrupt facies changes within narrow, generally elongate basins characterized by high sedimentation rates, very thick stratigraphic successions, and abundant evidence of paleoseismic activity are typical of the fill of strike-slip basins. Half-graben patterns in transverse cross sections are common. Coarse-grained debris-flow-dominated aprons characterize active strike-slip basin margins, whereas fluvio-deltaic facies characterize the opposite margins, and submarine or sublacustrine turbidite, fluviodeltaic, eolian and playa-evaporite facies typify axial deposystems. Depocenters migrate parallel to the most active strike-slip margin, resulting in accumulation of extraordinarily thick sedimentary sections in small basins. Basinwide upward coarsening sequences record progradational sedimentary systems that result from forced regressions caused by periodic lowering of base level induced by tectonism, sea/lake-level changes and climatic changes.

From our perspective, an understanding of strike-slip basins is best enhanced by a thorough understanding of how other types of basins develop. Strike slip is a significant component of almost all plate-bounding and intraplate settings; thus, to a greater or lesser degree, their understanding helps the understanding of all other tectonic settings. Continued fault displacement and polycyclic episodes of sedimentation and uplift commonly place most long-lived strike-slip basins in the general category of "complex basins." Their original paleotectonic framework may be difficult to recognize and restore.

John C. Crowell (1987), in characterizing Tertiary basins in southern California, wrote: "complexity is the hallmark of their tectonic history," and by inference, their sedimentation history as well. He wrote (p. 240):

> Crustal mobility and sedimentation have gone on together across the whole region, but with belts and areas where, for a time, either deformation or sedimentation has dominated. Cross sections are bound to get more complicated at depth, inasmuch as older beds have been subject to more deformation than younger. Geologists must abandon simplicity as a main guiding principle, and expect complexity instead. Each new bit of information, such as that obtained from a new drill hole or a new seismic line, will help in elucidating the history, but it will also add complexity to the cross section already drawn. No longer does the Principle of Simplicity, so aptly exploited in physics, for example, apply unqualified to our task. We must substitute the Principle of Complexity: in historical or configurational science, new details add complications.

Acknowledgments

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