CHAPTER NINETEEN

Strike-Slip Tectonics

19.1 INTRODUCTION

Every year, a few earthquakes startle the residents of California. Some tremors do little more than rattle a few dishes, but occasionally the jolting of a great earthquake tumbles buildings, ruptures roads, and triggers landslides. Most earthquakes in California signify that a sudden increment of slip has occurred somewhere along the San Andreas fault zone. This strike-slip fault zone, which contains countless individual faults, accommodates northward motion of the Pacific Plate relative to the North American Plate (Figure 19.1a and b). Most North American residents have heard of the San Andreas Fault—but it's not the only major strike-slip fault zone on this planet! Strike-slip faults cut both continental and oceanic crust in many places. Examples in continental crust include the Queen Charlotte Fault of western Canada, the Alpine Fault of New Zealand (Figure 19.2a), the faults bordering the Dead Sea (Figure 19.2b), the Anatolian Faults of Turkey, the Chaman Fault in Pakistan, and the Red River and Altyn Tach Faults of China. Strike-slip faults in oceanic crust most commonly occur along mid-ocean ridges, where they trend perpendicular to the ridge axis and link segments of the ridge. But there are important examples that link nonaligned segments of trenches as well.

As we noted in Chapter 8, a **strike-slip fault,** in the strict sense, is a fault on which all displacement occurs in a direction parallel to the strike of the fault (i.e., slip lineations on a strike-slip fault are horizontal). Thus, strict strike-slip displacement does not produce uplift or subsidence. In the real world, strike-slip movement is commonly accompanied by a component of shortening or extension. Specifically, **transpression** occurs where there is a combination of strike-slip movement and shortening, and can produce uplift along the fault. **Transtension** occurs where there is a combination of strike-slip movement and extension, and can produce subsidence along the fault.

In this chapter, we discuss the nature of deformation within strike-slip fault zones (both oceanic and continental), and we review the tectonic settings in which

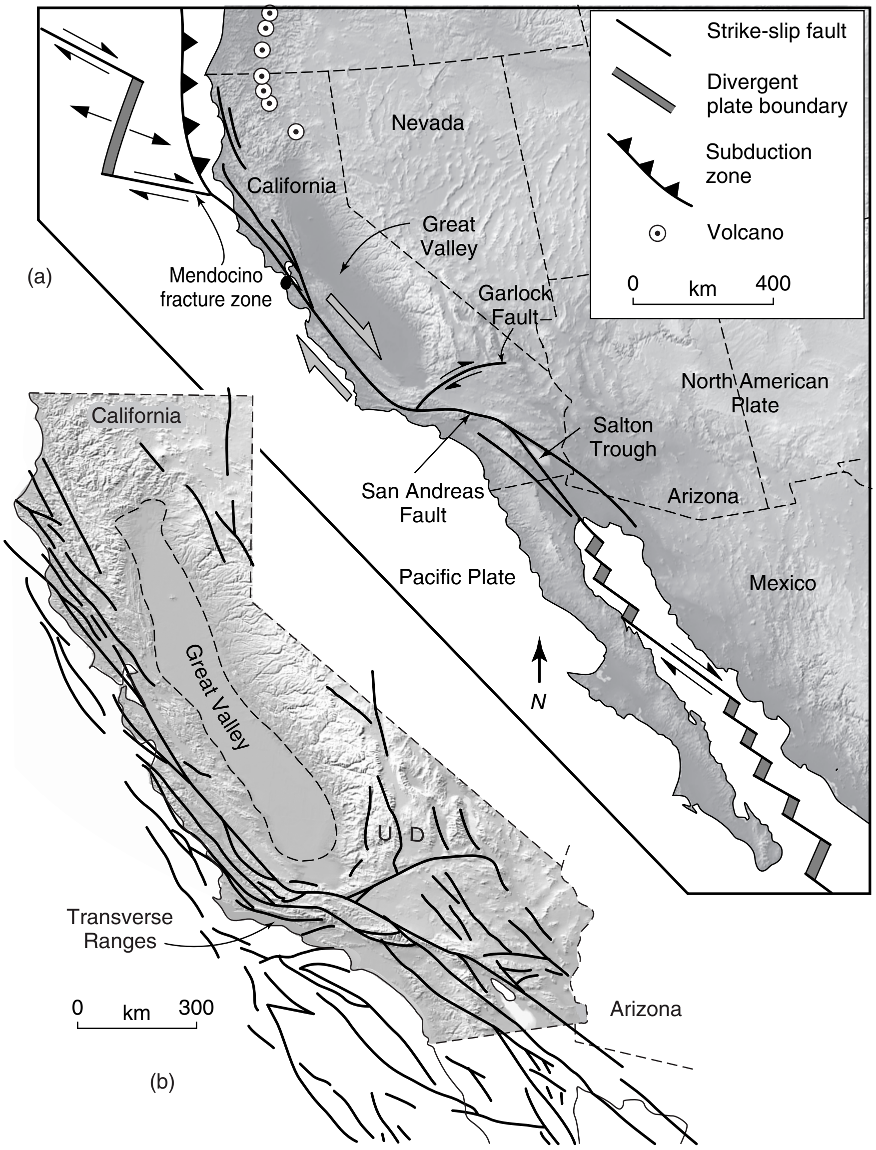


FIGURE 19.1 (a) Regional map of the plate boundary between the North American and Pacific Plates. The San Andreas Fault is the strike slip fault zone that defines this boundary in California. (b) An enlargement of California, showing the major strike slip faults. J & F = Juan de Fuca Plate.

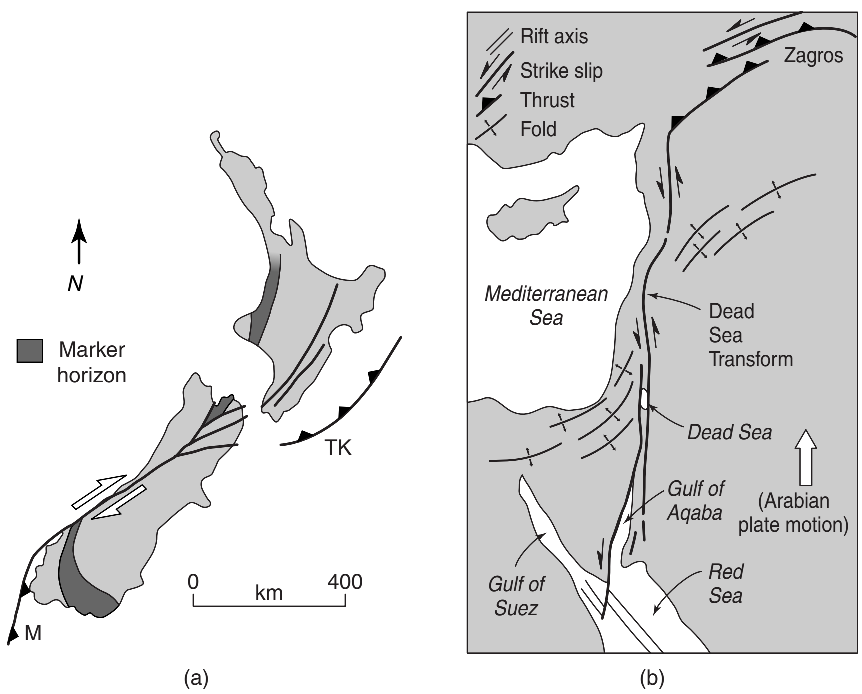


FIGURE 19.2 Examples of major strike-slip faults. (a) The Alpine Fault in New Zealand links the Macquarie Trench (M) with the Tonga-Kermadec Trench (TK). (b) The Dead Sea transform (DST) runs from the Gulf of Aqaba north to the western end of the Zagros Mountains. It accommodates northward movement of the Arabian Plate.

TABLE 19.1 TERMINOLOGY FOR STRIKE-SLIP FAULTS

Flower structure	An array of faults in a strike-slip fault zone that merges at depth into a near-vertical fault plane, but near the ground surface diverges so as to have shallower dips. In a positive flower structure, there is a component of thrusting on the faults, and in a negative flower structure, there is a component of normal faulting.
Horsetail	A group of fault splays at the endpoint (fault tip) of a fault; the splays diverge to define a fanlike array in map view.
Lateral escape	The strike-slip movement of a crustal block in a direction perpendicular to the regional convergence direction in a collisional orogen; the block is essentially squeezed laterally out of the way of the colliding blocks.
Lateral ramp	A surface connecting two non-coplanar parts of a thrust fault or normal fault (i.e., a ramp that is roughly parallel to the transport direction). Strike-slip movement occurs on a lateral ramp.
Non-coplanar	An adjective used to describe a geometric arrangement of two planes in which the planes have the same attitude, but are not aligned end-to-end to form a single plane.
Pull-apart basin	A sedimentary basin that forms at a releasing bend along a strike-slip fault.
Releasing bend	In the context of describing a strike-slip fault, it is a map-view bend, in the fault plane, whose orientation is such that the block on one side of the fault pulls away from the block on the other, causing transtensional deformation.
Restraining bend	In the context of describing a strike-slip fault, it is a map-view bend, in the fault plane, whose orientation is such that the block on one side of the fault pushes against the block on the other, causing transpressional deformation.
Stepover	A geometry that occurs where the end of one fault trace overlaps the end of another, non-coplanar fault trace.
Strike-slip duplex	An arrangement of sigmoidal-shaped fault splays (in map view) at a stepover, a restraining bend, or a releasing bend, whose geometry in map view resembles the cross-sectional geometry of the duplexes that occur in thrust-fault or normal-fault systems.
Strike-slip fault	Any fault on which displacement vectors are parallel to the strike of the fault, in present-day surface coordinates. The use of this term is purely geometric, and has no genetic, tectonic, or size connotation.
Tear fault	A traditional term for strike-slip faults that occur in a thrust sheet and accommodate differential displacement of one part of a thrust sheet relative to an adjacent part; in more recent jargon, a tear fault is a vertical lateral ramp.
Transcurrent fault	A strike-slip fault that has the following characteristics: it dies out along its length; the displacement across it is less than the length of the fault; the length of the fault increases with time and continued movement; displacement on the fault is greatest at the center of the fault trace and decreases toward the endpoints (tips) of the fault.
Transfer fault	A fault trending roughly perpendicular to the axis of a rift that links together two non-coplanar parts of the rift; transfer faults accommodate map-view offset of the locus of extension. Kinematically, transfer faults are a type of transform fault. If a rift succeeds in evolving into a mid-ocean ridge, the transfer fault becomes an oceanic transform fault.
Transform fault	A strike-slip fault that has the following characteristics: once formed, displacement across it can be constant along the length of the fault; displacement across it can be much greater than the length of the active fault; its length can be constant, increase, or decrease with time; it terminates at another fault. Transform faults can be lithosphere plate boundaries, in which case they terminate at intersections with other plate boundaries.
Transpression	A combination of strike-slip and compressional deformation; transpression occurs where a fault is not parallel to the map projection of regional displacement vectors, so that there is a component of compression across the fault (leading to shortening).
Transtension	A combination of strike-slip and tension; transtension occurs where a fault is not parallel to the map projection of regional displacement vectors, so that there is a component of tension across the fault (leading to extension).
Wrench fault	A traditional synonym for a strike-slip fault. It was commonly used in the petroleum geology literature, typically in reference to a regional-scale continental strike-slip fault. We abandon the term, as it tends to be used inconsistently.

these faults develop. You will find that a wide variety of complex subsidiary structures develop in strike-slip fault zones. We begin the chapter by explaining the kinematic distinction between transfer faults and transcurrent faults, the two basic classes of strike-slip faults. Basic terms used in discussing strike-slip faults are provided in Table 19.1.

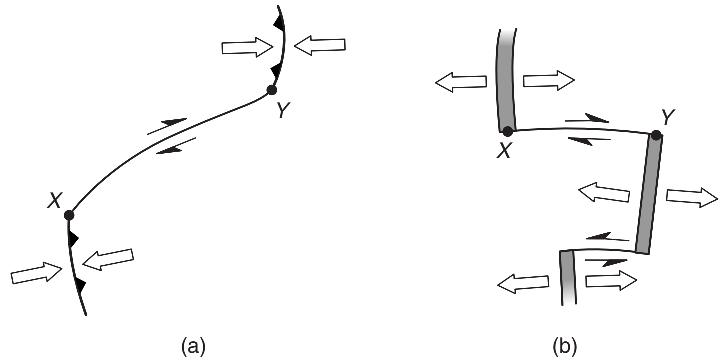


FIGURE 19.3 (a) Map sketch showing a transform fault linking two convergent plate boundaries. The fault trace occurs only between *X* and *Y*. (b) Map sketch showing a transform fault linking two ridge segments.

19.2 TRANSFORM VERSUS TRANSCURRENT FAULTS

When examining the role that strike-slip faults play in crustal deformation, we find it convenient to distinguish between two kinematic classes of faults—transform faults and transcurrent faults.[1] A fault in one class differs from a fault in the other class in terms of the geometry of its endpoints (the locations along strike where the fault terminates), the way that slip magnitude varies along the fault's length, and the way that the fault's geometry evolves through time. To make this statement more concrete, let's look at the specific characteristics in more detail.

19.2.1 Transform Faults

J. Tuzo Wilson introduced the term "transform faults" to the geologic literature in the early 1960s, in reference to the third category of plate boundary (distinct from convergent and divergent boundaries). While some strike-slip faults are, indeed, plate boundaries, the term **transform fault** can be applied more broadly to describe any strike-slip fault that has the following characteristics:

- The active portion of a transform fault terminates at discrete endpoints. At the endpoints, the transform intersects other structures (Figure 19.3). For example, the transform can terminate at a shortening structure (e.g., convergent boundary, thrust fault, or stylolite) or at an extensional structure (e.g., divergent boundary, normal fault, or vein).
- The length of a transform fault can be constant, or it can increase or decrease over time. For example, in Figure 19.4a, the spreading rate on ridge segment *A* is the same as the spreading rate on ridge segment *B*. As a consequence, the length of the transform fault connecting these ridge segments remains constant over time. In contrast, Figure 19.4b shows a transform fault where one end terminates at a ridge segment, but the other end terminates at a trench. If the rate of subduction at the trench exceeds the rate of spreading on the ridge, then the length of the transform fault connecting them decreases over time. In Figure 19.4c, the length of the transform fault

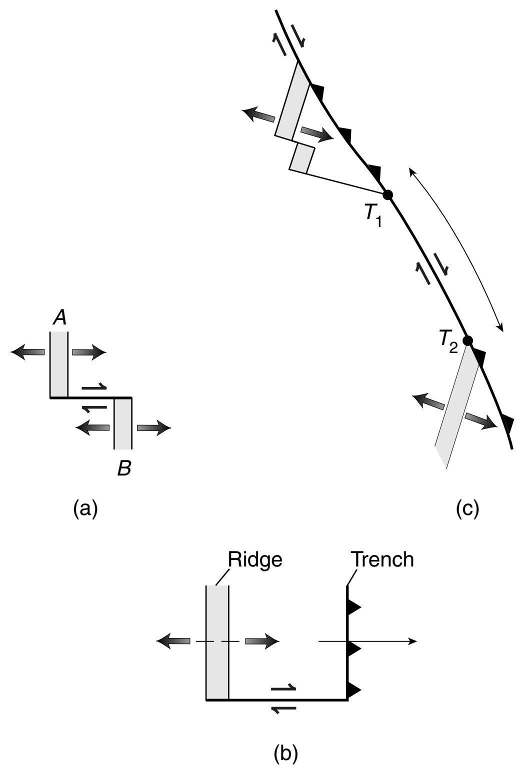


FIGURE 19.4 The length of a transform fault can change with time. (a) Transform length stays constant if spreading rates on ridge segments at both endpoints are the same. (b) Transform length decreases if the spreading rate at one endpoint is less than the subduction rate at the other. (c) Transform length increases as the two triple junctions (T_1 and T_2) defining the endpoints move apart.

[1]This discussion is based on a paper by R. Freund (1974).

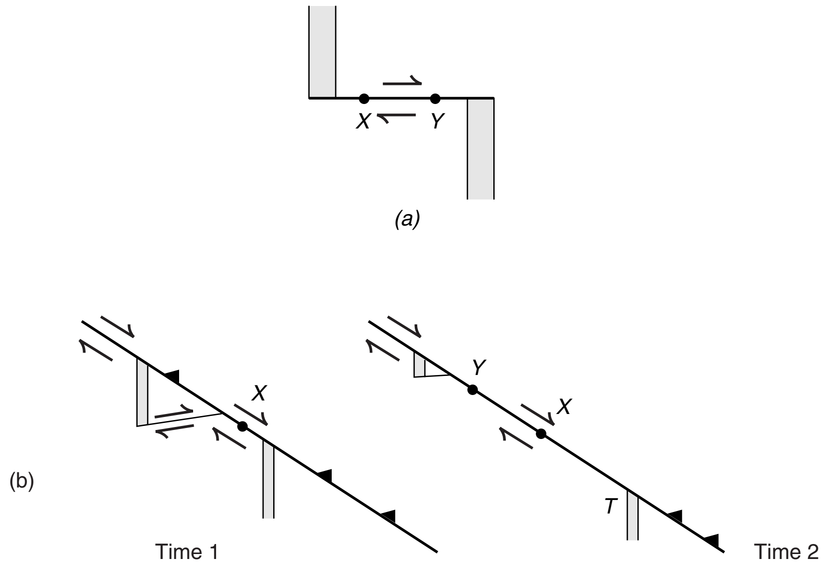


FIGURE 19.5 (a) The amount of displacement remains the same along the length of a transform if the length of the transform stays constant or decreases. Displacement at *X* is the same as the displacement at *Y*. (b) If the transform length changes with time, then the amount of slip varies along the length. At time 1, the fault is fairly short. At time 2, the length of the fault is longer. Displacement at *X*, in the middle of the fault, is greater than displacement at point *Y*, near an endpoint.

between two triple junctions labeled (T_1 and T_2) increases with time if the triple junctions move apart.

- In cases where the fault length is constant or decreasing, the amount of displacement along the length of a transform fault is constant. For example, the displacement at point *X* on the fault in Figure 19.5a is the same as the displacement at point *Y*. If the length of a transform fault increases over time, however, the amounts of displacement on the younger portions of the fault are less than the amounts on the older portions (Figure 19.5b).
- Displacement across a transform fault can be greater than the length of the fault itself. For example, consider a 10-km long transform fault that links two ridge segments. If more than 10 km of spreading occurs on the ridge segments, then there will be more than 10 km of slip on the transform fault (Figure 19.6).

Transform faults occur in a variety of settings and at a variety of scales (from mesoscopic to regional). At a mesoscopic scale, transform faults can link non-coplanar stylolites or non-coplanar veins. In Chapter 16, we saw how transfer faults link segments of rifts. Transfer faults are oriented at a high angle to the rift segments and terminate at normal faults. Kinematically, transfer faults behave like transform faults.

Where rifting is successful, so that a new divergent plate boundary forms, transfer faults in the rift evolve into classic transform faults that link spreading segments of a divergent plate boundary (Figure 19.7a and b). Note that such transform faults are plate boundaries, as first recognized by J. Tuzo Wilson.[2] Let's see how such transform faults display the kinematic characteristics we described above by looking at an example, a 10-km long transform fault linking two segments of the Mid-Atlantic Ridge (Figure 19.7c). Each end of the transform fault terminates abruptly at a ridge segment (i.e., at a zone of extension) trending at a high angle to the transform. If the spreading rate on each ridge segment is the same, then the length of the transform stays constant over time, regardless of the amount of slip that has occurred along it. Further, the same amount of slip occurs everywhere along the length of the fault. (This makes sense if you remember that the transform fault initiated at the same time as the ridge segments—it was "born" with the length that it now has. Note also that the sense of slip on the faults must be compatible with the spreading directions on the ridge segments.) The amount of slip along an oceanic transform fault can be much greater than the length of the fault. For example, if there has been 1000 km of spreading on the ridge segments, there must be 1000 km of displacement on the 10-km long transform.

Again we note that not all plate-boundary transform faults link segments of mid-ocean ridges. Some link non-coplanar segments of subduction zones, or link other transform faults (Figure 19.2a). For example, a transform fault links the east-dipping convergent plate boundary along the west coast of South America to the west-dipping convergent plate boundary of the Scotia Sea (Figure 19.7c). Note that in cases where a transform fault links a ridge and a trench, the fault length may decrease with time (Figure 19.4b); this happens if the subduction rate is faster than the spreading rate.

[2]Wilson wrote his paper about transform faults in 1965. His prediction of the sense of slip for transform faults was confirmed a couple of years later by Lynn Syke's study of fault-plane solutions for earthquakes on oceanic transforms and stands as one of the proofs of plate tectonics.

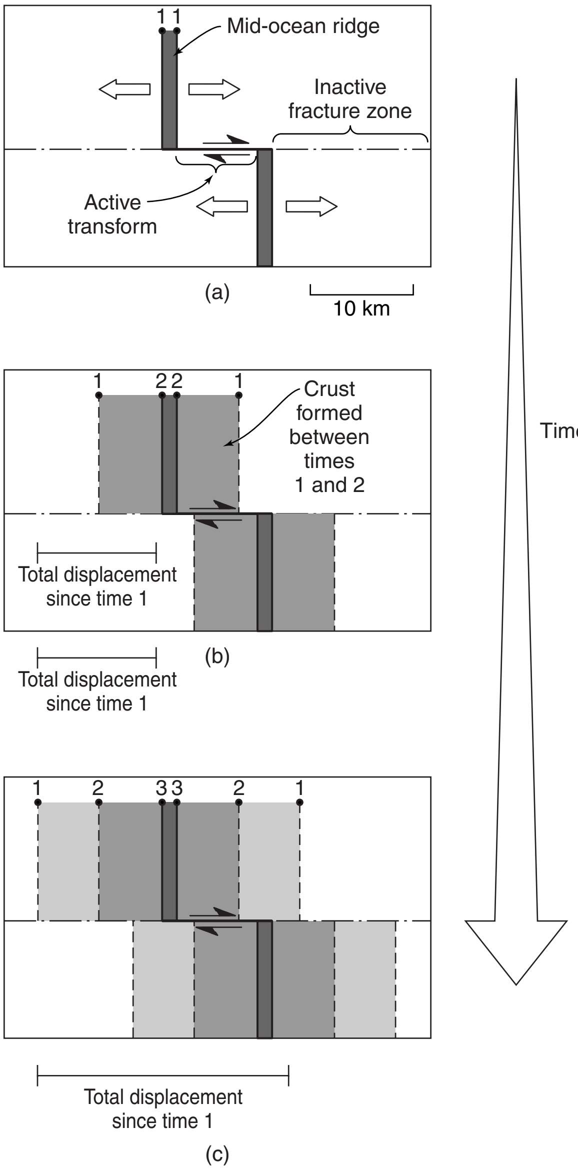


FIGURE 19.6 The amount of slip on a transform can exceed the length of the transform. (a) At time 1, the transform is 10 km long. Its length does not change over time, even though spreading at the ridges occurs continuously. (b) At time 2, the amount of displacement on the fault is already 1.5× the length of the fault. (c) At time 3, the displacement on the fault is 3.0× the length of the fault.

19.2.2 Transcurrent Faults

Transcurrent faults differ kinematically from transform faults in a number of ways:

- Transcurrent faults die out along their length. This means that at the endpoint of a transcurrent fault, the fault does not terminate abruptly at another fault, but either splays into an array of smaller faults (sometimes called a **horsetail**), or simply disappears into a zone of plastic strain. Typically, fault

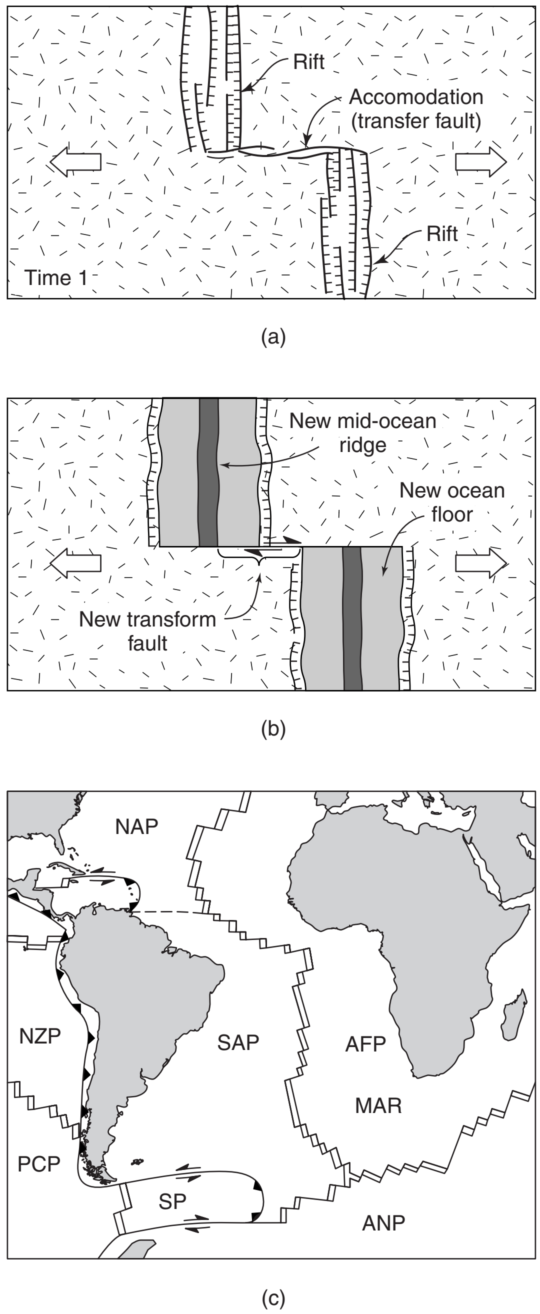
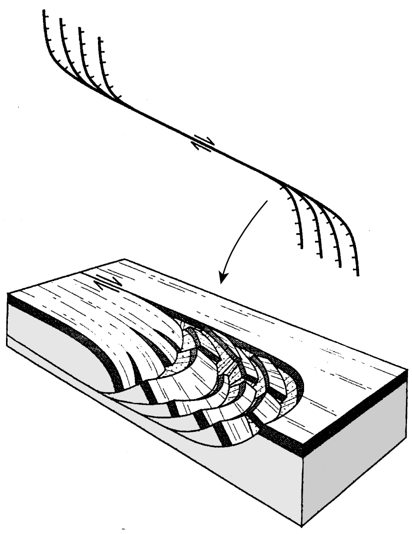


FIGURE 19.7 (a) Sketch map showing a transfer fault linking two rift segments. (b) Later, the rift segments have evolved into new mid-ocean ridges, and the transfer fault has evolved into an oceanic transform fault. Note that this fault still has the same length as when it originated. (c) Simplified map showing the transform faults along the Mid-Atlantic Ridge in the South Atlantic Ocean. SAP = South American Plate, NAP = North American Plate, AFP = African Plate, ANP = Antarctic Plate, NZP = Nasca Plate, CS = Caribbean Sea, SS = Scotia Sea, MAR = Mid-Atlantic Ridge.

splays comprising a horsetail are curved. Depending on the direction of curvature with respect to the sense of displacement, thrust or normal components of displacement occur in faults of the horsetail, and these movements will be accompanied either by folding and uplift where there is a thrust component, or by tilting and subsidence where there is a normal component. (Figure 19.8).

- Transcurrent faults initiate at a point and grow along their length as displacement increases (Figure 19.9a). As a consequence, short faults have a small amount of displacement, while long faults have a large amount of displacement. Thus, fault displacement (as measured in map view) is proportional to fault length.
- Displacement across a transcurrent fault is greatest near the center of its trace and decreases to zero at the endpoints of the fault (Figure 19.9b).



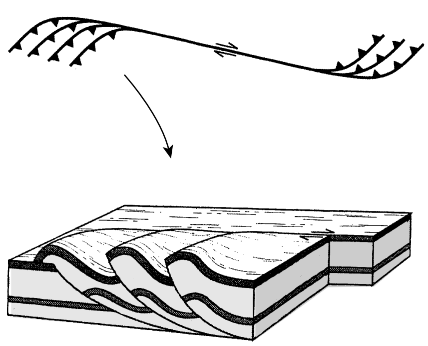


FIGURE 19.8 Terminations along transcurrent faults. (a) Here, the fault terminates in a horsetail composed of a fan of normal faults. (b) Here, the fault terminates in a horsetail composed of a fan of thrust faults.

- The displacement on a transcurrent fault must always be less than the length of the fault (Figure 19.9b).

Transcurrent faults typically develop in continental crust as a means of accommodating development of regional strain. For example, slip on a conjugate system of mesoscopic transcurrent faults results in shortening of a block of crust in the direction parallel to the line bisecting the acute angle between the faults. As is the case with transform faults, transcurrent faults can form at any scale, from mesoscopic to regional.

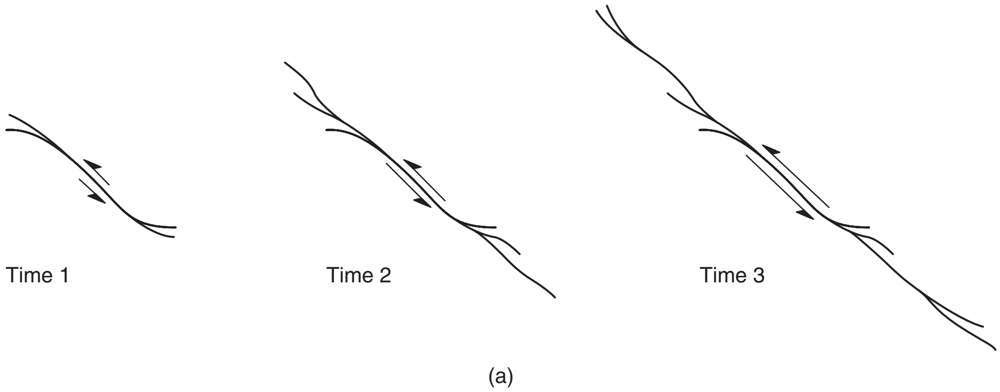
19.3 STRUCTURAL FEATURES OF MAJOR CONTINENTAL STRIKE-SLIP FAULTS

Major continental strike-slip faults, meaning ones that have trace lengths ranging from tens to thousands of kilometers, are not simple planar surfaces. Typically, such faults have locally curved traces, divide into anastomosing (braided) splays, include several parallel branches, and/or occur in association with subsidiary faults and folds (Figure 19.10). In this section, we describe the structural complexities of large continental strike-slip faults and suggest why these complexities occur.

19.3.1 Description of Distributed Deformation in Strike-Slip Zones

Regional-scale strike-slip deformation in continental crust does not produce a single, simple fault plane. Rather, such shear produces a broad zone containing numerous individual strike-slip faults of varying lengths, as well as other structures such as normal and reverse faults, and folds.

Let's first look at the array of individual strike-slip faults that occurs in a large continental strike-slip fault zone. As an example, consider the San Andreas fault zone of California. The San Andreas Fault "proper" is one of about 10 major strike-slip faults, and literally thousands of minor strike-slip faults, that slice up western California (Figure 19.1). In some localities, faults have sinuous traces, so neighboring faults merge and bifurcate to define, overall, an anastomosing array (Figure 19.10). Elsewhere, the zone includes several subparallel faults. Locally, one fault dies out where another, parallel, but non-coplanar one initiates. The region between the endpoints of two parallel but non-coplanar



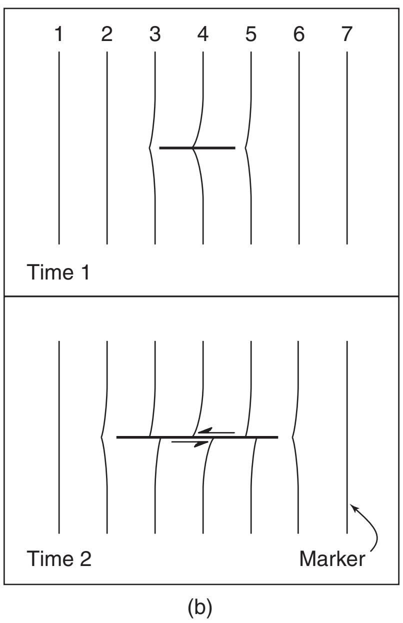


FIGURE 19.9 (a) Growth of a transcurrent fault. As time passes, the fault lengthens, and the displacement on the fault increases. In the process, some horsetail splays are abandoned. (b) Map sketch illustrating how the displacement varies along a transcurrent fault (horsetails are not shown). The heavier line is the fault, and the thin lines are marker lines. At time 1, the fault is short and has offset only marker 4. At time 2, the fault has grown, and now offsets markers 3, 4, and 5. Note that the markers just beyond the tips of the fault are starting to bend, prior to rupturing. Note also that the displacement on the fault is less than the length of the fault.

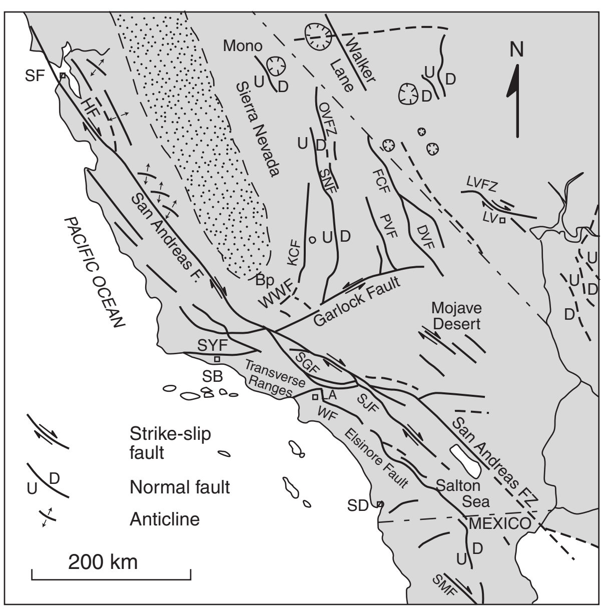


FIGURE 19.10 Faults of southern California. Note that faults anastomose and that, locally, folds form adjacent to the faults. Also note the large bend in the San Andreas Fault at its intersection with the Garlock Fault. B = Bakersfield, LA = Los Angeles, LV=Las Vegas, SB = Santa Barbara, SD = San Diego, SF = San Francisco, DVF = Death Valley Fault, FCF = Furnace Creek Fault, HF = Hayward Fault, KCF = Kern Canyon Fault, LVFZ = Las Vegas fault zone, OVFZ = Owens Valley fault zone, PVF = Panamint Valley Fault, SGF = San Gabriel Fault, SJF = San Jacinto Fault, SNF = Sierra Nevada Fault, SYF = Santa Ynez Fault, WF = Whittier Fault, WWF = White Wolf Fault.

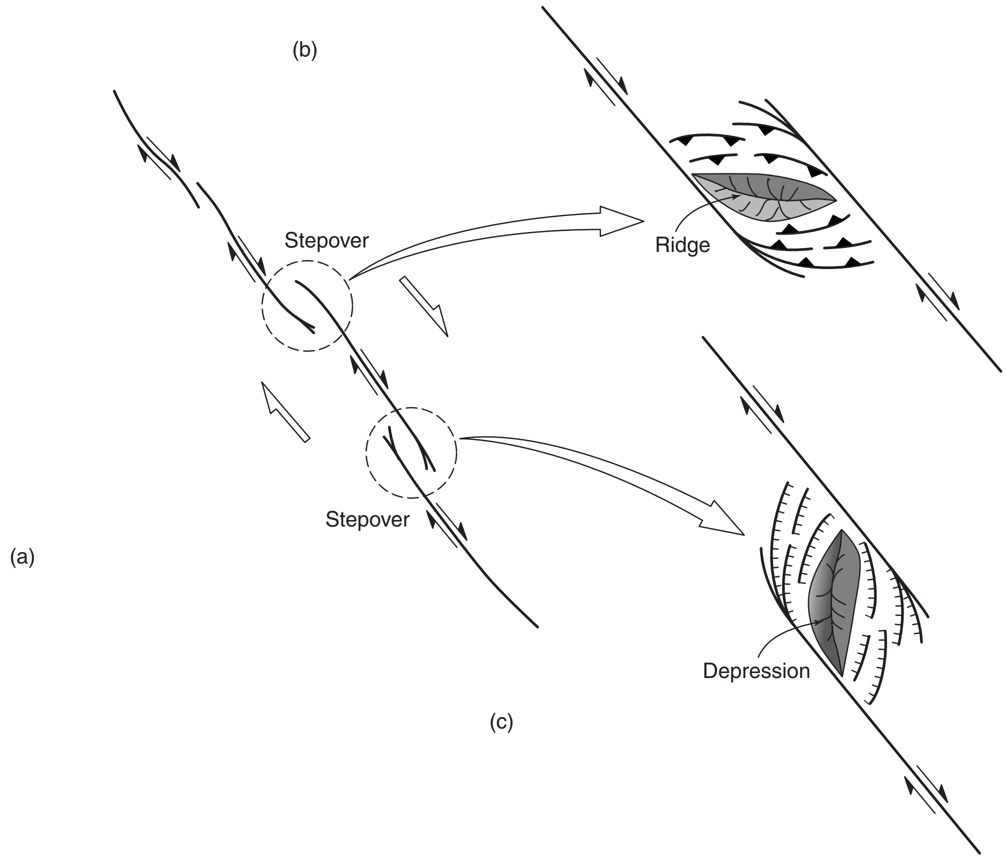


FIGURE 19.11 A stepover along a strike-slip fault. (a) Sketch map showing how regional dextral shear can be distributed along fault segments that are not coplanar. Slip is relayed from one segment to another at a stepover. (b) At a restraining stepover, compression and thrusting occur. (c) At a releasing stepover, extension and subsidence occur.

faults is called a **stepover** (Figure 19.11). Localized faulting occurs in the stepover region to accommodate the transfer of slip.

Continental strike-slip fault zones also may contain *en echelon* arrays of thrust faults, folds, and normal faults, as well as subsidiary strike-slip faults (Figure 19.12). Typically, the thrust faults and folds (Figure 19.13) trend at an angle of 45° or less to the main fault, and the acute angle between subsidiary thrust faults and the main fault (or between the fold hinges and the main fault) opens in the direction of shear. The normal faults also trend at an angle of 45° or more with respect to the main fault, but the acute angle between subsidiary normal faults and the main fault opens opposite to the direction of shear. Most subsidiary strike-slip faults trend at a shallow angle to the main fault.

19.3.2 The Causes of Structural Complexity in Strike-Slip Zones

Why do continental strike-slip fault zones contain so many subsidiary faults? To understand this complexity, we'll review an experiment that we introduced in Chapter 8. Take a homogeneous clay slab and place it over two adjacent wooden blocks (Figure 19.14). The clay cake represents the weaker uppermost crust, and the wooden blocks represent stronger crust at depth. Now, push one of the blocks horizontally so that it shears past its neighbor. As the blocks move relative to one another, the clay cake begins to deform, partly by plastic mechanisms and partly by brittle failure. Initially, this brittle deformation yields arrays of small strike-slip faults called **Riedel shears.** Two sets of Riedel shears, labeled R and R′, develop. We can picture these shears as forming a conjugate system relative to the far-field maximum compressive stress driving development of the overall fault zone. Eventually, P shears develop (see Chapter 8), which finally link with Riedel shears to form a throughgoing strike-slip fault. With this model, we can speculate that some of the subsidiary faults in a strike-slip zone initiated as Riedel shears or as P shears.

The model just described suggests that, even if the upper crust were homogeneous, one might expect strike-slip zones to contain subsidiary strike-slip faults. In reality, the upper crust is heterogeneous.

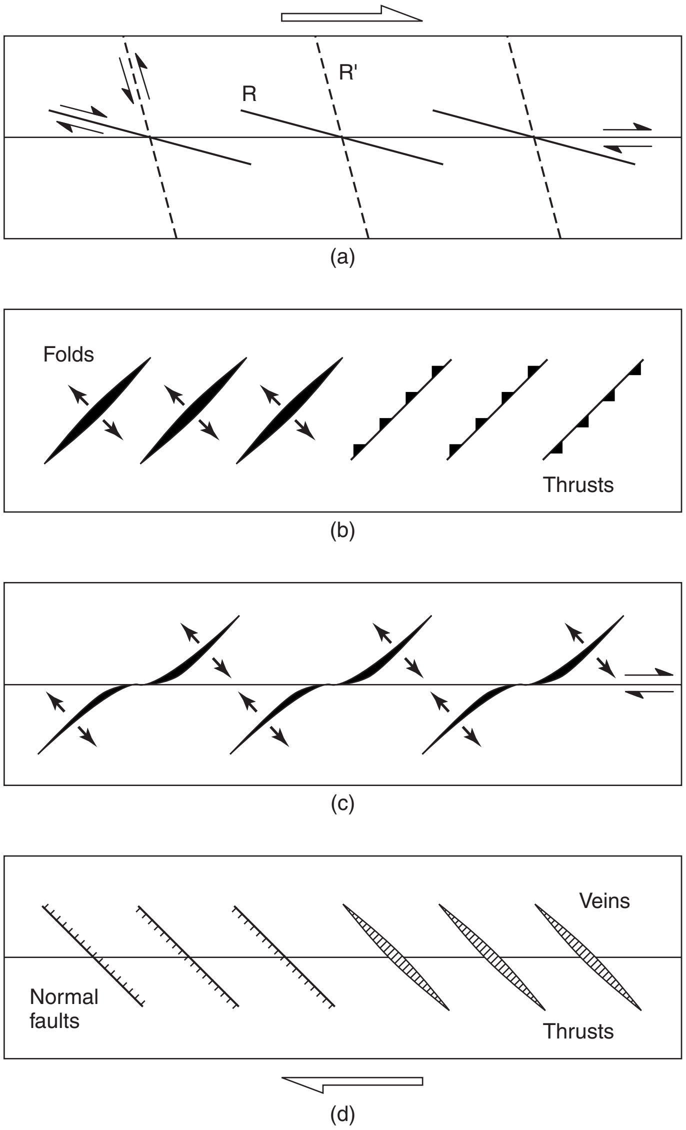


FIGURE 19.12 Arrays of subsidiary structures associated with dextral shear. (a) Subsidiary strike-slip faults (Riedel shears: R and R′); (b) *en echelon* folds, and *en echelon* thrusts; (c) *en echelon* folds which formed and then were later offset by shear on a strike-slip fault; (d) *en echelon* normal faults and veins.

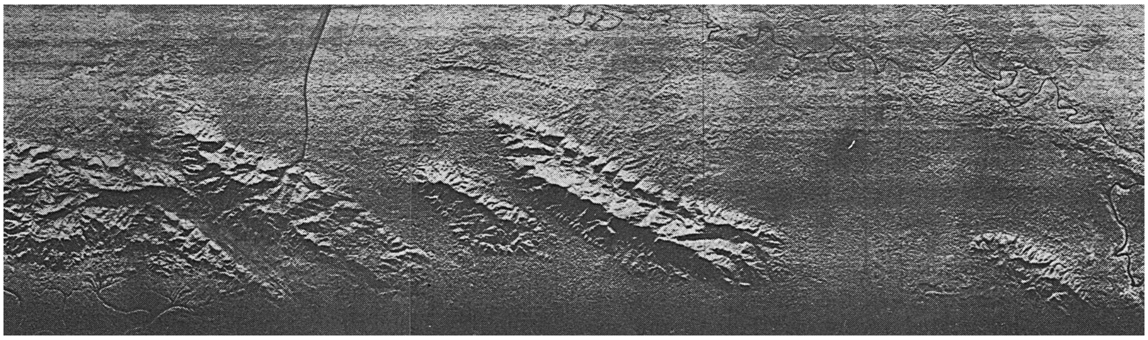


FIGURE 19.13 A side-scan radar image from the Darien Basin in eastern Panama showing an array of *en echelon* anticlines whose formation has arched up the land surface, creating a set of ridges. The field of view is about 50 km. Note that the geometry of these structures indicates left-lateral shear.

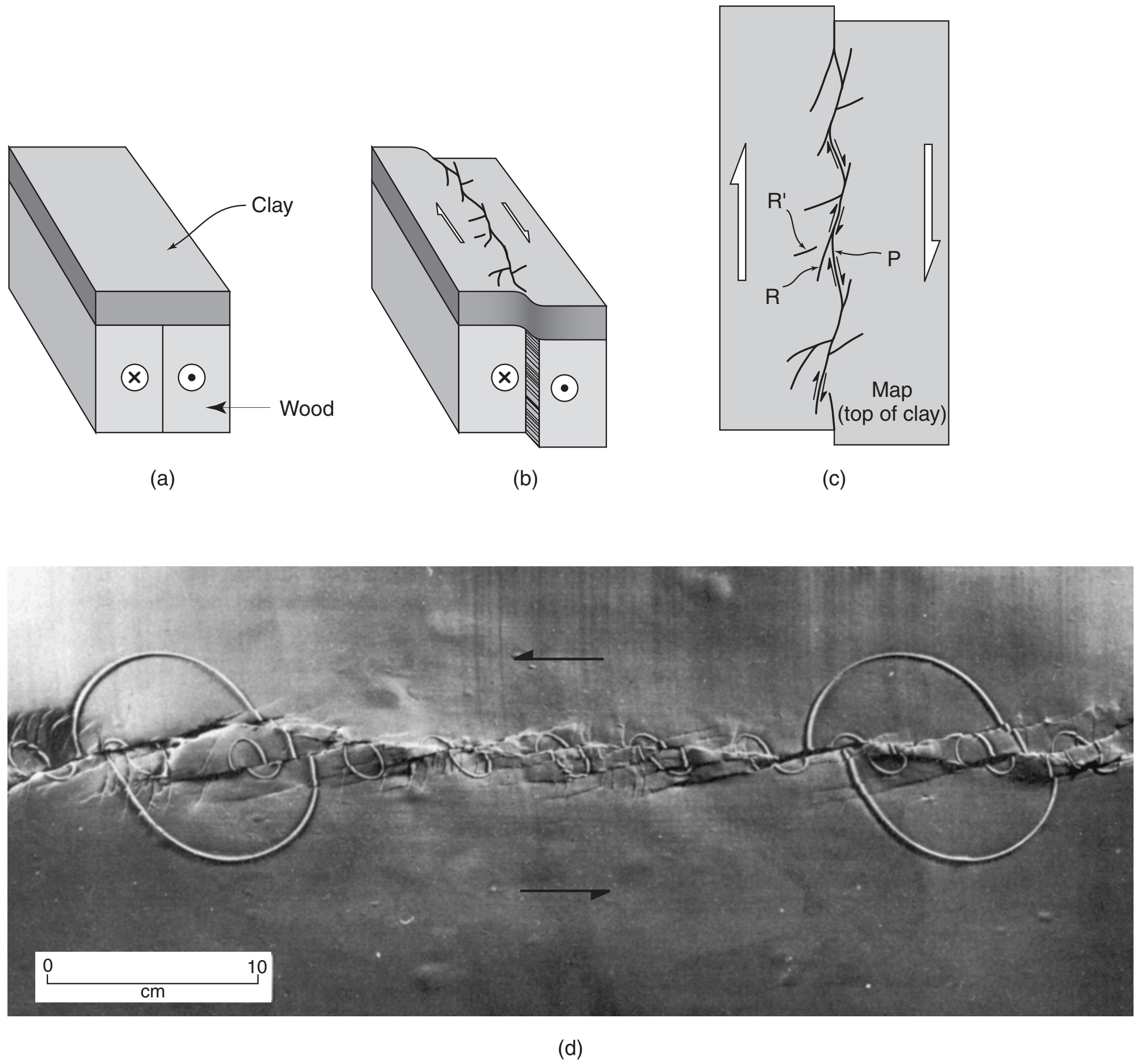
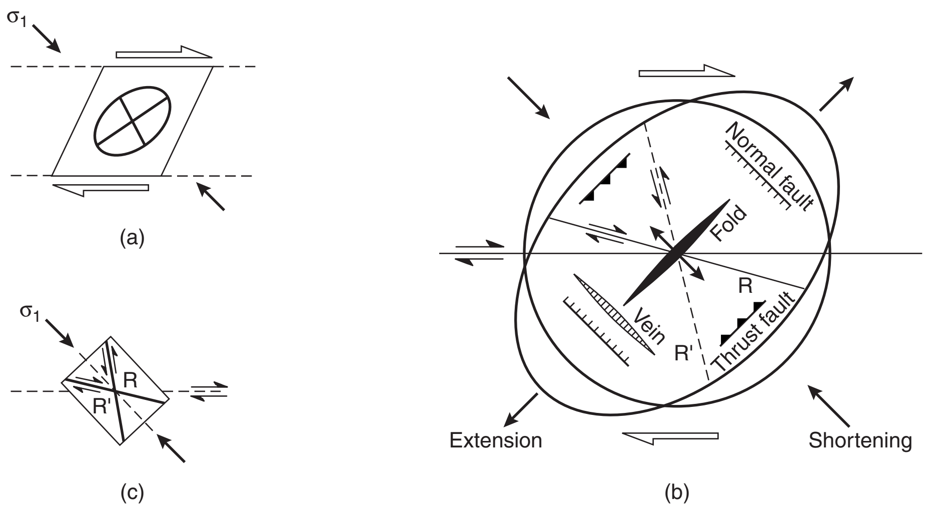


FIGURE 19.14 A laboratory model of strike-slip fault development. (a) Before deformation, a clay cake rests on two wooden blocks that were pressed together. The clay represents the weak uppermost crust, and the wood blocks represent the stronger lower crust. The vertical boundary between the two blocks represents the strike-slip fault. (b) As deformation begins, Riedel shears develop in the clay cake. (c) A map view of the top surface of the clay cake, showing a later stage of deformation, in which Riedel shears have been linked by P fractures. A throughgoing fault has just developed. (d) An example of a clay-cake experiment, this one for left-lateral shear.

Crust may contain a variety of different rock types with different strengths, and the contacts between rock units may occur in a variety of orientations. In addition, the crust contains preexisting planar weaknesses such as joints, old faults, and foliations. All these heterogeneities cause stress concentrations and local changes in stress trajectories. As a result, faults may locally bend, and they may split to form two strands on either side of a stronger block. In sum, the process by which the fault is formed, as well as the crust's heterogeneity, ensures that strike-slip fault zones include a variety of subsidiary fault splays.

To picture why *en echelon* arrays of thrusts, folds, and normal faults have the orientations that they do, picture a block of crust that is undergoing simple shear in map view. An imaginary square superimposed on the zone transforms into a rhomb, and an imaginary circle transforms into a strain ellipse (Figure 19.15a and b). In the direction parallel to the short axis of the ellipse, the crust shortens, so thrusts and folds develop perpendicular to this axis. In the direction parallel to the long axis of the ellipse, the crust stretches, so normal faults and veins develop perpendicular to this axis. As deformation pro-



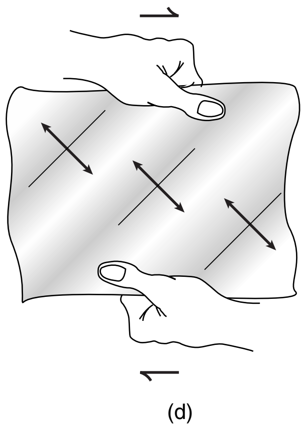


FIGURE 19.15 A strain model explaining the origin of subsidiary structures along a strike-slip fault. (a) A map view of dextral simple shear. A square becomes a parallelogram, and a circle in the square becomes an ellipse. (b) A detail of the strain ellipse showing that folds and thrusts form perpendicular to the shortening direction, while normal faults and veins form perpendicular to the extension direction. R and R′ shears form at an acute angle to the shortening direction. (c) Note that R and R′ are similar to conjugate shear fractures formed in rock cylinder subjected to an axial stress. (d) You can simulate formation of *en echelon* folds with a sheet of paper.

gresses, the main strike-slip fault eventually slices the block in two, and the two halves of the subsidiary faults and folds are displaced with respect to one another. Note that R and R′ shears can be pictured as conjugate shear fractures whose acute bisector between the faults is parallel to σ_1, the regional maximum principle stress (Figure 19.15c). Also note that you can simulate deformation in a strike-slip fault zone by shearing a piece of paper between your hands (Figure 19.15d); the ridges that rise in the center of the paper represent folds.

19.3.3 Map-View Block Rotation in Strike-Slip Zones

Progressive slip on strike-slip faults may result in map-view rotation of crustal blocks (i.e., rotation of the blocks about a vertical axis). Such rotation can happen is one of three ways. (1) As illustrated by contrasting Figure 19.16a with Figure 19.16b, progressive simple shear causes material lines in the fault zone to rotate; (2) In places where several parallel strike-slip faults dice up a crustal block, the block as a whole can rotate around a vertical axis in the same way that books on a bookshelf tilt around a horizontal axis when they stay in contact with the shelf but shear past one another (Figure 19.16b). This process is the map-view equivalent of the process leading to rotation by slip on planar normal faults (see Chapter 9); (3) Small crustal blocks caught in a strike-slip zone may rotate about a vertical axis like ball bearings between two sheets of wood (Figure 19.16c).

19.3.4 Transpression and Transtension

The trace of the San Andreas Fault in southern California is not just a featureless line on the surface of the Earth. At some localities, the trace lies within a train of marshy depressions (called sag ponds); elsewhere, the trace is marked by 50-m high ridges

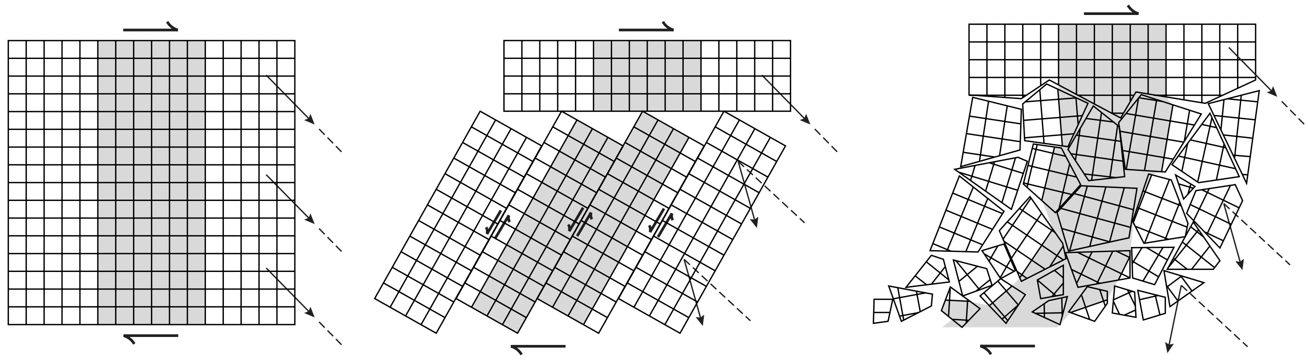


FIGURE 19.16 Mechanisms of block rotation in a right-lateral strike-slip zone. (a) A map view of a grid being subjected to dextral simple shear. (b) The grid lines rotate. Further, fault-bounded blocks may rotate intact as slip on bounding faults increases (bookshelf model). (c) Alternatively, the fault zone may break into smaller blocks that rotate by different amounts. In this case, large rotations may occur locally. The amount of rotation is tracked by the paleomagnetic declination in each block (solid arrows) relative to a reference direction (dashed line).

(called pressure ridges) that contain tight folds (Figure 19.17). The presence of such topographic and structural features tells us that motion along the fault is not perfectly strike-slip. Ridges form in response to **transpression,** a combination of strike-slip displacement and compression that yields a component of shortening across the fault. This shortening causes thrusting and uplift within or adjacent to the fault zone. Notably, where the fault zone contains a broad band of weak fault rocks (gouge and breccia), transpression squeezes the fault rocks up into a fault-parallel ridge, much like a layer of sand squeezes up between two wood blocks that have been pushed together (Figure 19.18a). Topographic depressions reflect **transtension,** a combination of strike slip and extension. The extensional component causes normal faulting and subsidence (Figure 19.18b). Transpression or transtension can occur along the entire length of a fault zone, if the zone trends oblique to the vectors describing the relative movement of blocks juxtaposed by the fault. Such a situation develops where global patterns of plate motion change subsequent to the formation of the fault, for a fault is a material plane in the Earth and cannot change attitude relative to adjacent rock once it has formed.

The dimensions of transpressive or transtensile structures forming along a strike-slip fault depend on the amount of cross-fault shortening or extension, respectively. Where relatively little transpressive or transtensile deformation has occurred, cross-fault displacement results in relatively small pressure ridges or sags, with relief that is less than a couple of hundred meters. If, however, transpression or transtension has taken place over millions of years, significant mountain ranges develop adjacent to the fault. For example, transpression along the Alpine Fault of New Zealand has resulted in uplift of the Southern Alps, a range of mountains that reaches an elevation of up to 3.7 km above sea level.

Seismic-reflection studies of large continental strike-slip faults indicate that subsidiary faults in transpressional or transtensional zones within strike-slip systems are concave downwards, and merge at depth into the main vertical fault plane (Figure 19.19a). Thus, in cross section, large continental strike-slip faults resemble flowers in profile, with the petals splaying outwards from the top of a stalk. This configuration of faults is, not surprisingly, referred to as a **flower structure.**[3] In transpressive zones, a **positive flower structure** develops, in which the slip on subsidiary faults has a thrust-sense component (Figure 19.19b); whereas in transtensile zones, a **negative flower structure** develops, in which the slip on subsidiary faults has a normal-sense component (Figure 19.19c).

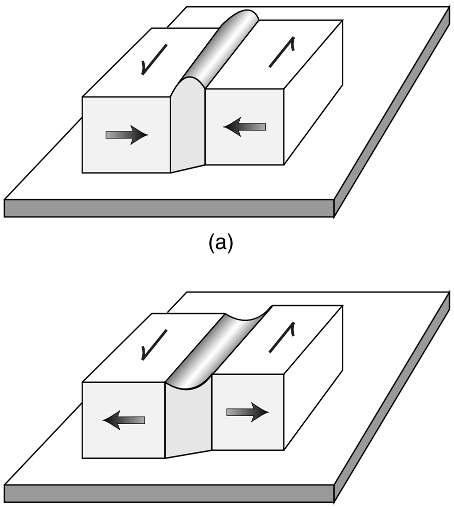
[3]Some authors prefer the term "palm structure" for this arrangement of faults, because the shape of the faults more closely resembles fronds on a palm tree than petals on most flowers. But the term "flower structure" tends to be used more commonly. For discussion of these structures, see Sylvester (1988).

(a)

(b)

(c)

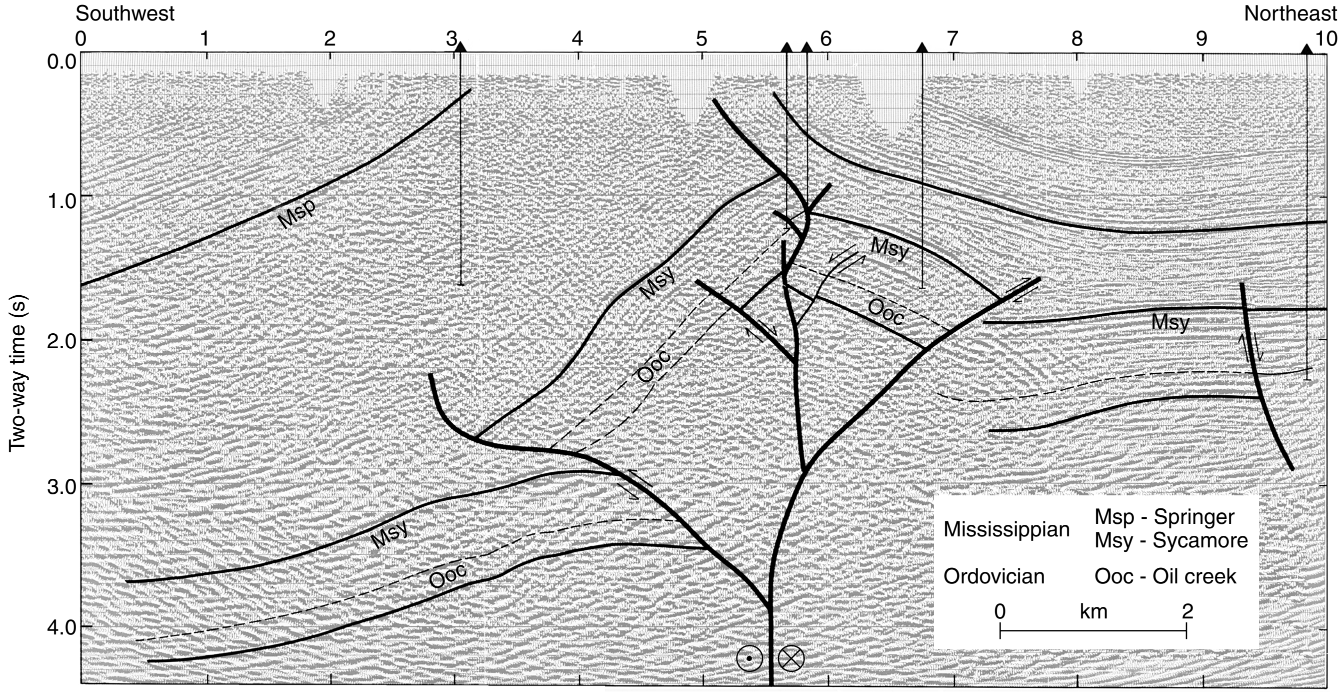
FIGURE 19.17 (a) Air photo showing the trace of the San Andreas Fault, north of San Francisco (Tomales Bay). Note that the faulting has locally caused a water-filled depression to form. (b) Photograph of pressure ridges along the San Andreas Fault, San Luis Obisbo County, California. (c) A cross section of a pressure ridge in a road cut across the San Andreas Fault near Palmdale, California.



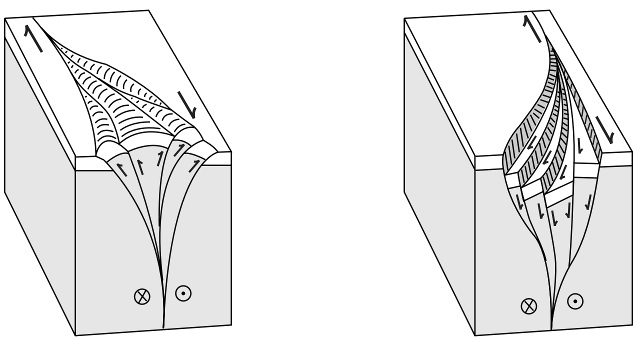
(a)

(b)

FIGURE 19.18 Simple wood block model illustrating the concept of transpression and transtension. (a) When blocks shear and squeeze together (transpression), sand is pushed up; (b) When blocks shear and pull apart (transtension), sand sags.



(a)



(b) (c)

FIGURE 19.19 (a) Seismic-reflection profile across a strike-slip fault in the Ardmore Basin, Oklahoma, showing a positive flower structure. Msp and Msy are Mississippian formations, and Ooc is Ordovician. (b) Block diagram of a positive flower structure. (c) Block diagram of a negative flower structure.

19.3.5 Restraining and Releasing Bends

In many locations, transpression and transtension take place at distinct bends in the trace of a strike-slip fault.[4] To see why, picture a fault that, overall, strikes east-west, parallel to the trend of regional displacement vectors (Figure 19.20a and b). Along segments of the fault that strike exactly east-west, motion on the fault can be accommodated by strike-slip motion alone, for the fault plane parallels the regional displacement vectors. But at fault bends, where the strike of the fault deviates from east-west, the fault plane does not parallel regional displacement vectors, and there must be either transpression or transtension, depending on the orientation of the bend. Let's look at the geometry of these bends in more detail.

A bend at which transpression takes place is called a **restraining bend,** because the fault segment in the

[4]Recall that a fault bend, in the context of discussing strike-slip faults, is a portion of a fault where the strike of the fault changes. Elsewhere in this book, we have also used the term "fault bend" when discussing dip-slip faults; a fault bend along a dip-slip fault is a location where the fault's dip changes.

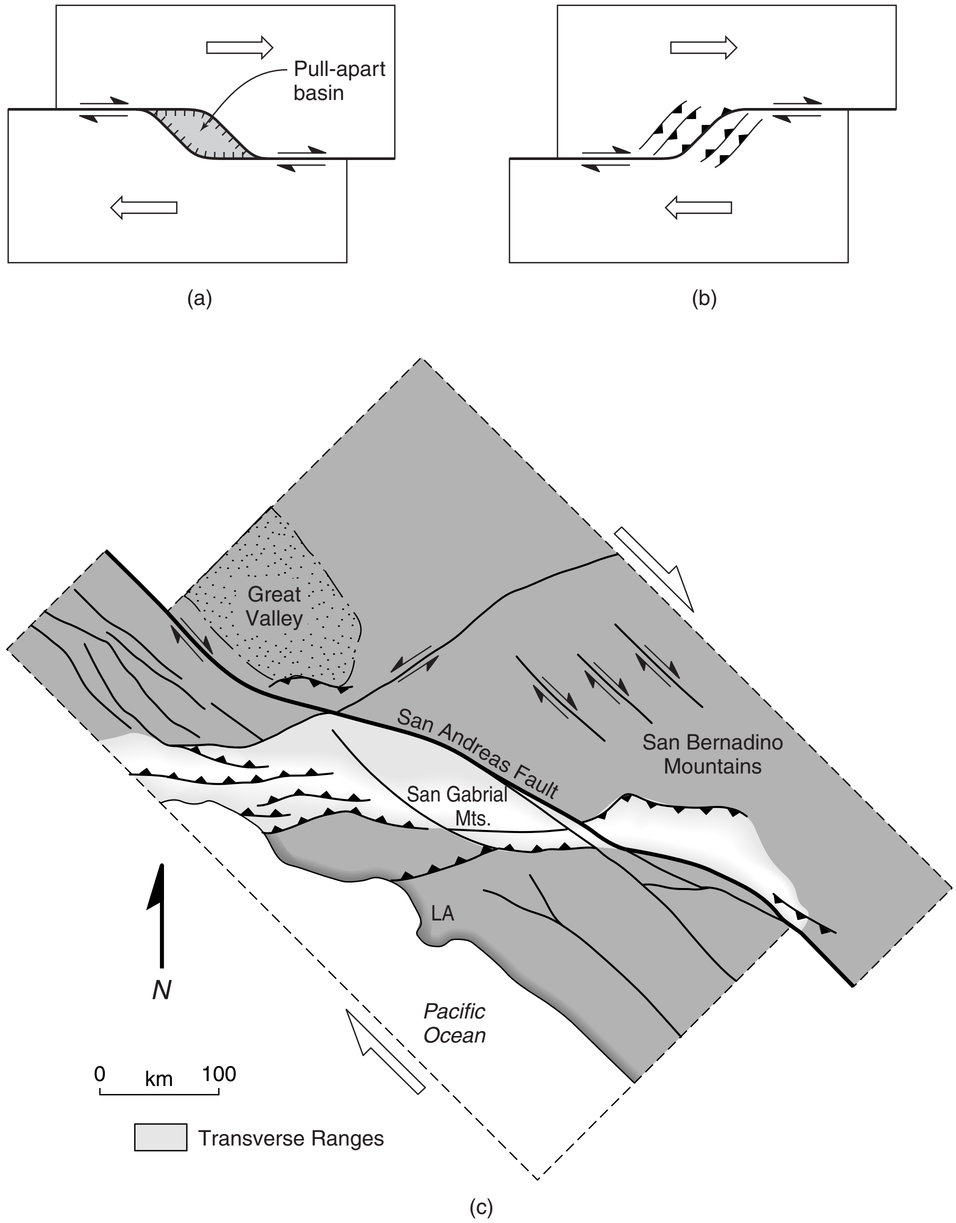


FIGURE 19.20 Map-view models of fault bends along strike-slip faults. The "edges" of the crustal blocks are provided for reference. (a) Releasing bend at which normal faults and a pull-apart basin have formed. (b) Restraining bend at which thrust faults have formed. (c) Application of this model to the San Andreas Fault north of Los Angeles (LA). The dashed lines outline imaginary reference blocks. The San Andreas Fault bends along the margin of the Mojave Desert.

bend inhibits motion. In other words, at restraining bends, pieces of crust on opposite side of the fault push together, causing crustal shortening. Along regional-scale restraining bends, a fold-thrust belt can form, causing the uplift of a **transverse mountain range** (i.e., a mountain range trending at an angle to the regional trace of the fault). For example, the Transverse Ranges just north of Los Angeles in southern California developed because of shortening across a large restraining bend along the San Andreas Fault (Figure 19.20c).

A bend at which transtension occurs is called a **releasing bend** because, at such bends, opposing walls of the fault pull away from each other (Figure 19.20b). As a consequence of this motion, normal faults develop, and the block of crust adjacent to the bend subsides. Displacement at a releasing bend yields a negative flower structure or, in cases where the bend is

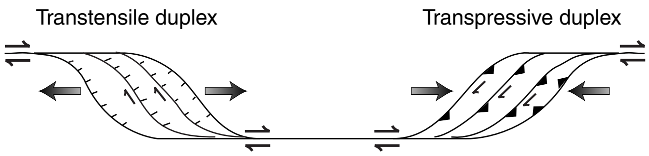


FIGURE 19.21 Map-view sketch of strike-slip duplexes formed along a dextral strike-slip fault.

large, and large amounts of extension have taken place, a pull-apart basin. A **pull-apart basin** is a rhomboid-shaped depression, formed along a releasing bend and filled with sediment eroded from its margin. The dimension and the amount of subsidence in a pull-apart basin depends on the size of the bend and on the amount of extension. Notably, formation of small pull-apart basins involves brittle faulting only in the upper crust, but formation of large pull-apart basins involves thinning of the lithospheric mantle, so that after extension ceases, the floor of the basin thermally subsides (see Chapter 16). Examples of present-day pull-apart basins include the Dead Sea (Figure 19.2b) at the border between Israel and Jordan, and Death Valley in eastern California. In both of these basins, the land surface lies below sea-level, creating an environment in which summer temperatures become deadly hot.

Both restraining bends and releasing bends can exist simultaneously at different locations along the same fault. As a consequence, a region of crust moving along the fault may at one time be subjected to transtension and then, at a later time, be subjected to transpression. When this happens, a negative flower structure, or pull-apart basin formed at a releasing bend, becomes inverted and changes into a positive flower structure, and normal faults bordering a pull-apart basin become reverse faults. Inversion causes sediment that had been deposited in negative flower structures or pull-apart basins to thrust up and over the margins of a strike-slip fault zone.

19.3.6 Strike-Slip Duplexes

A **strike-slip duplex** consists of an array of several faults that parallels a bend in a strike-slip fault (Figure 19.21). The map-view geometry of a strike-slip duplex resembles the cross-sectional geometry of thrust-fault duplex or a normal-fault duplex. Strike-slip duplexes formed at restraining bends can also be called **transpressive duplexes,** while those formed at releasing bends can also be called **transtensile duplexes.**

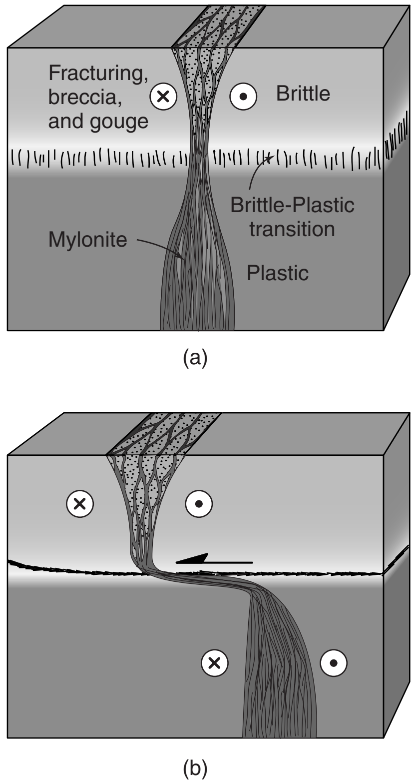


FIGURE 19.22 (a) A regional-scale strike-slip fault may consist of a broad zone of breccia and gouge at shallow crustal levels. At deeper layers, the zone may narrow into a zone of cataclasite and, at great depth, it broadens into a zone of mylonite, grading down into schist. (b) The shallower portion of some strike-slip faults may be offset, relative to the deeper portion, by a detachment at depth.

19.3.7 Deep-Crustal Strike-Slip Fault Geometry

What do regional-scale strike-slip fault zones look like at progressively greater depths in the crust? Based on studies of exposed crustal sections of ancient faults, geologists surmise that the fault zone is very weak near the surface, and thus is very broad. With increasing depth, the crust becomes stronger and the fault zone becomes narrower. Then, at the brittle–plastic transition, the crust becomes weaker, and the fault zone widens (Figure 19.22a).

Recent studies suggest that the concept of a strike-slip fault zone cutting down through the entire crust as a continuous vertical zone of deformation may be an oversimplification in some locations. There is growing evidence that regional-scale strike-slip faults are locally offset at subhorizontal detachments deep in the crust, so that the strike-slip displacement in the upper crust does not lie directly over displacements in the lower crust and underlying mantle lithosphere (Figure 19.22b).

19.4 TECTONIC SETTING OF CONTINENTAL STRIKE-SLIP FAULTS

Until now, we have focused our discussion on characteristic structural features that occur in strike-slip fault zones. Now, we broaden our perspective and turn our attention to the tectonics of strike-slip faulting. Specifically, we describe the various plate settings at which strike-slip fault zones develop. Several of these occur in southern Asia (Figure 19.23).

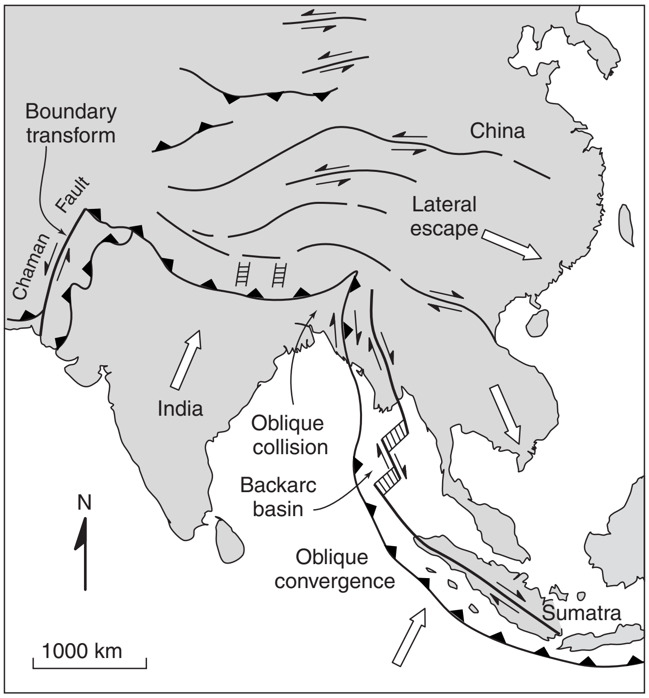


FIGURE 19.23 Sketch map of southern Asia, showing the collision of India. Strike-slip faults have developed in several settings here. A boundary transform (the Chaman Fault) delimits the northwestern edge of the Indian subcontinent. Strike-slip faults also form due to oblique collision, oblique convergence, and lateral escape. Small rifts have developed just north of the Himalayas.

19.4.1 Oblique Convergence and Collision

Strike-slip faults occur along convergent plate boundaries where the vector describing the relative motion between the subducting and overriding plates is not perpendicular to the trend of the convergent margin (Figure 19.24a). At such **oblique-convergent plate margins,** the relative motion between the two plates can be partitioned into a component of dip-slip motion (thrusting) perpendicular to the margin, and a component of horizontal shear (strike-slip faulting) parallel to the margin. Present-day examples illustrate that the strike-slip faults of oblique-convergent plate boundaries develop in a variety of locations across the margin, including the accretionary wedge, the volcanic arc, and the backarc region.

Partitioning of relative movement into dip-slip and strike-slip components accompanies the oblique collision of two buoyant lithospheric masses (Figure 19.24b). The strike-slip component of motion displaces fragments of crust laterally along the orogen. When the colliding mass is a small exotic terrane, the terrane may be sliced up by strike-slip faults after it docks (Figure 19.24c). For example, Wrangelia, an exotic crustal block that was incorporated into the western margin of North America during Mesozoic oblique convergence, was sliced by strike-slip faults into fragments that were then transported along the margin. As a result, bits and pieces of Wrangelia occur in a discontinuous chain that can be traced from Idaho to Alaska (see Figure 17.29).

Strike-slip faulting also develops at collisional margins where one continent indents the other. The vise created when two continents converge may cause blocks of crust caught between the colliding masses to be squeezed laterally out of the zone of collision, a process called **lateral escape.** The boundaries of these "escaping" blocks are strike-slip faults. For example, during the Cenozoic collision of India with Asia, India pushed northward into Asia. Stress from this collision resulted in a fan-shaped pattern of strike-slip faults accommodating the escape of blocks of Asia eastward (Figure 19.23). A similar phenomenon is currently happening in the eastern Mediterranean, where the northward movement of the Arabian Peninsula along the Dead Sea transform has resulted in the westward escape of Turkey, squeezed like a watermelon seed between the North Anatolian Fault and the East Anatolian Fault (Figure 19.25).

19.4.2 Strike-Slip Faulting in Fold-Thrust Belts

The traces of thrust faults trending nearly perpendicular to the regional transport direction dominate the map pattern of fold-thrust belts. Locally, however, these belts contain strike-slip faults whose traces trend nearly parallel to the regional transport direction (Chapter 18). Some of these faults, known as **lateral ramps,** cut a thrust sheet into pieces that move relative to one another. In places where lateral

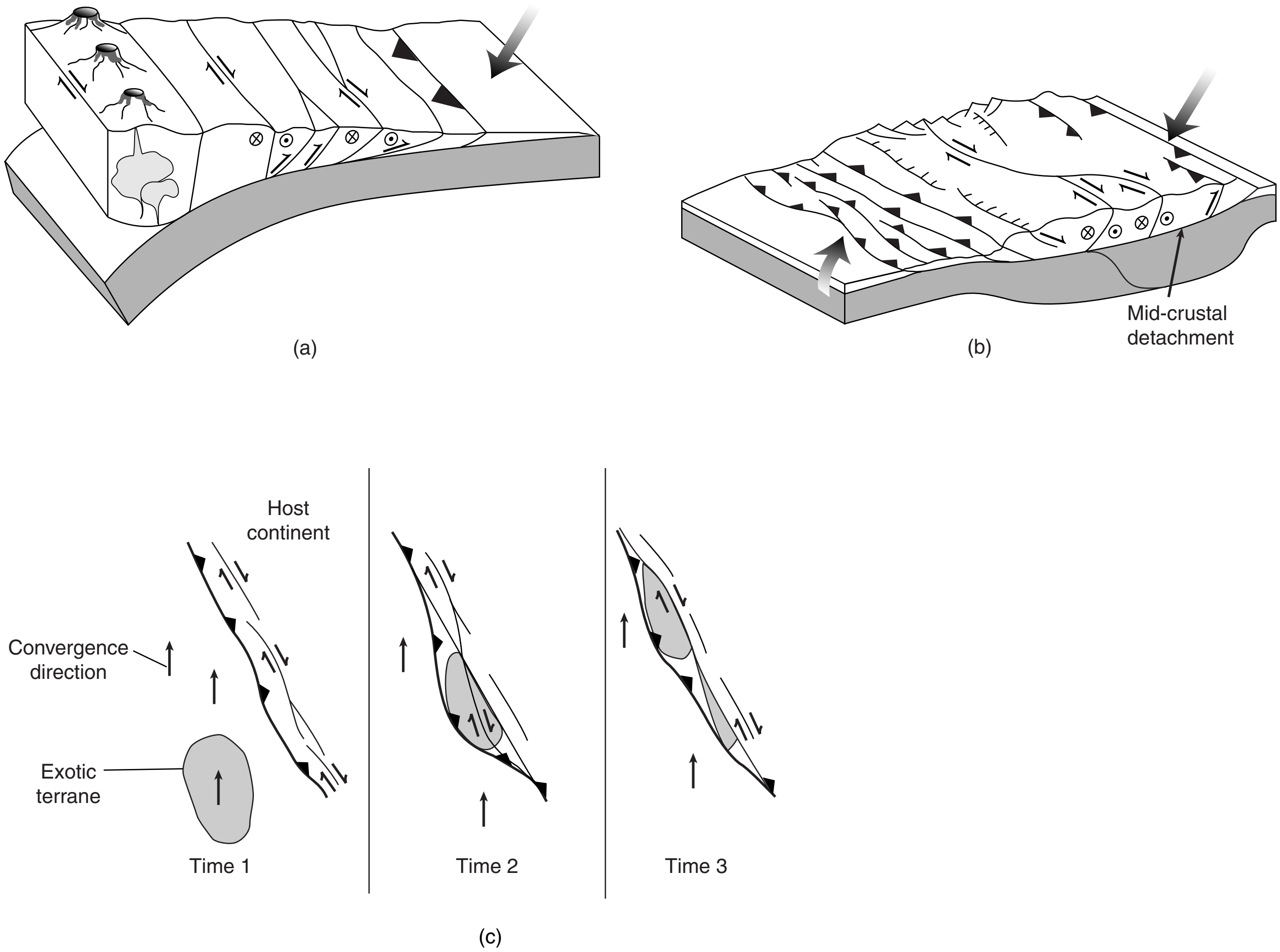


FIGURE 19.24 (a) Strike-slip faulting at an oblique convergent margin. Note that the faults occur at various locations across the width of the margin. The large arrow indicates the relative movement of the downgoing plate. (b) Strike-slip faulting at an oblique collisional margin. Note that in this portrayal, the faults terminate at depth at a mid-crustal detachment horizon. (c) Map-view sketches showing progressive stages during oblique docking of an exotic terrane. Note how the terrane is sliced by faults subsequent to docking, and slivers slip along the length of the orogen.

ramps have a near-vertical dip, they are also called **tear faults.** Lateral ramps or tear faults can accommodate along-strike changes in the position of a frontal ramp with respect to the foreland. Examples of lateral ramps bound the two ends of the Pine Mountain thrust sheet in the southern Appalachians (Figure 19.26).

Locally, thrust sheets contain conjugate mesoscopic strike-slip fault systems, which accommodate shortening of thrust sheets across strike and simultaneous stretching of thrust sheets along strike. Examples of conjugate strike-slip systems stand out in the Makran fold-thrust belt of southern Pakistan, and in the Jura fold-thrust belt of Switzerland and France (Figure 19.27).

19.4.3 Strike-Slip Faulting in Rifts

As we noted earlier, rifts typically consist of a chain of distinct segments. Adjacent segments in the chain may differ in amount of extension, and in the dominant dip direction of normal faults. Also, the axis of one segment may not align with the axis of the adjacent segment in the chain. Each segment in a rift is linked to its neighbor by an **accommodation zone.** In places where an accommodation zone consists of a strike-slip fault, it can also be called a **transfer fault** (see Chapter 16; also Figure 19.7). Where transtension or transpression occurs, flower structures may develop along transfer faults.

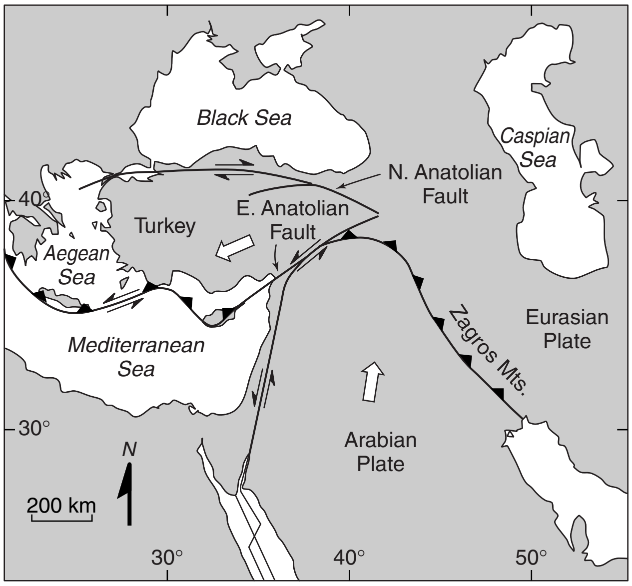
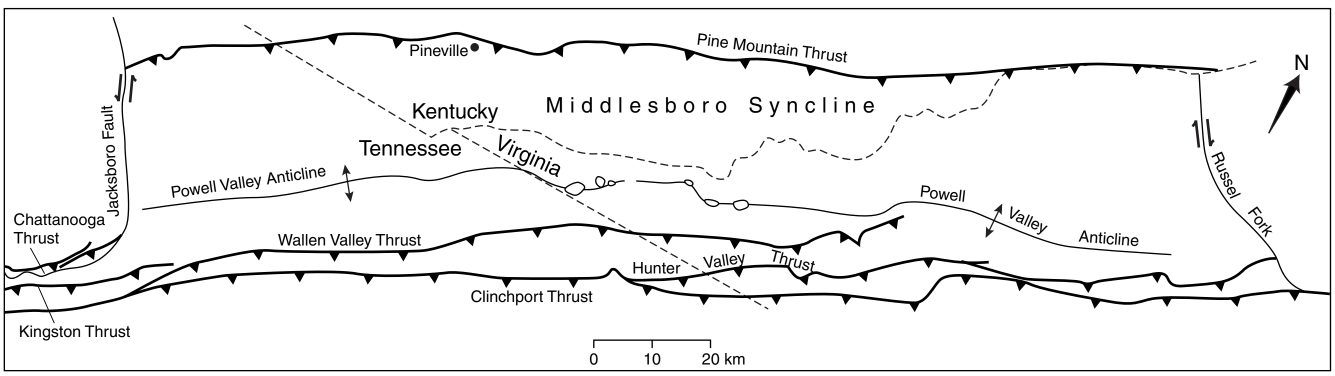


FIGURE 19.25 Sketch map of the region from the eastern Mediterranean to the Caspian Sea. Lateral escape in response to the northward movement of the Arabian Plate is squeezing Turkey out to the west. Escape is accommodated by slip on the North Anatolian Fault and the East Anatolian Fault.

The Garlock Fault in southern California is one of the largest examples of a strike-slip within a rift environment (Figure 19.22). This fault, a ~250-km long left-lateral strike-slip fault, forms the northern border of the Mojave Desert and intersects the San Andreas Fault at its western end. It defines the boundary between two portions of the Basin and Range Rift: the "Central Basin and Range" lying to the north, and the Mojave block lying to the south. The Garlock Fault exists because these two portions have not developed the same amount of extensional strain. Specifically, a greater amount of extension has occurred in the Central Basin and Range than in the Mojave block (Figure 19.28). To accommodate this difference, the crust to the north of the fault has slid westward, relative to the crust to the south.[5] Notably, this motion pushed the portion of the San

[5]This model was first proposed by B. C. Burchfiel and G. A. Davis (1973).



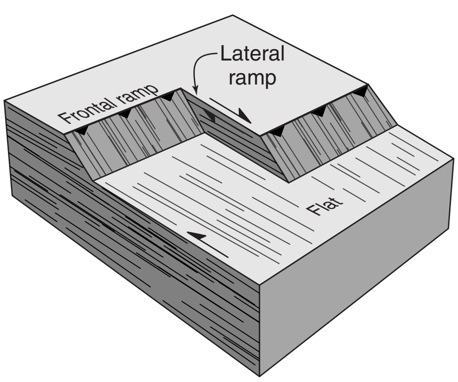


FIGURE 19.26 (a) Map of the Pine Mountain thrust system in the southern Appalachians (eastern USA), showing lateral ramps (Jacksboro Fault and Russell Fork Fault). (b) Block diagram indicating the concept of a lateral ramp or tear fault.

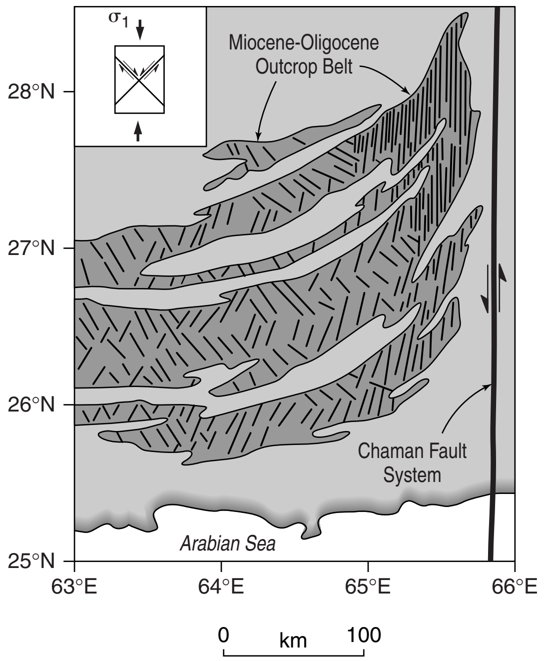


FIGURE 19.27 Simplified map of the Makran fold-thrust belt in Pakistan, illustrating the conjugate system of strike-slip faults contained within thrust slices.

Andreas Fault north of its intersection with the Garlock Fault to the west, creating a major restraining bend in the San Andreas fault zone. The presence of this bend is responsible for development of the Transverse Ranges just north of Los Angeles (Figure 19.20).

Motion between the Pacific and North American Plates across the Transverse Ranges bend in the San Andreas Fault also generates compressive stress within the Mojave block. This stress has caused some of the normal faults formed by rifting in the block in the past to be reactivated today as strike-slip faults. Motion on these faults results in rotation, around a vertical axis, of crust within the Mojave block (Figure 19.20; also Figure 19.16b).

19.4.4 Continental Transform Faults

Where transform plate boundaries cross continental lithosphere, strike-slip deformation disrupts the continental crust. Several examples of continental transforms crop out on the surface of the Earth.

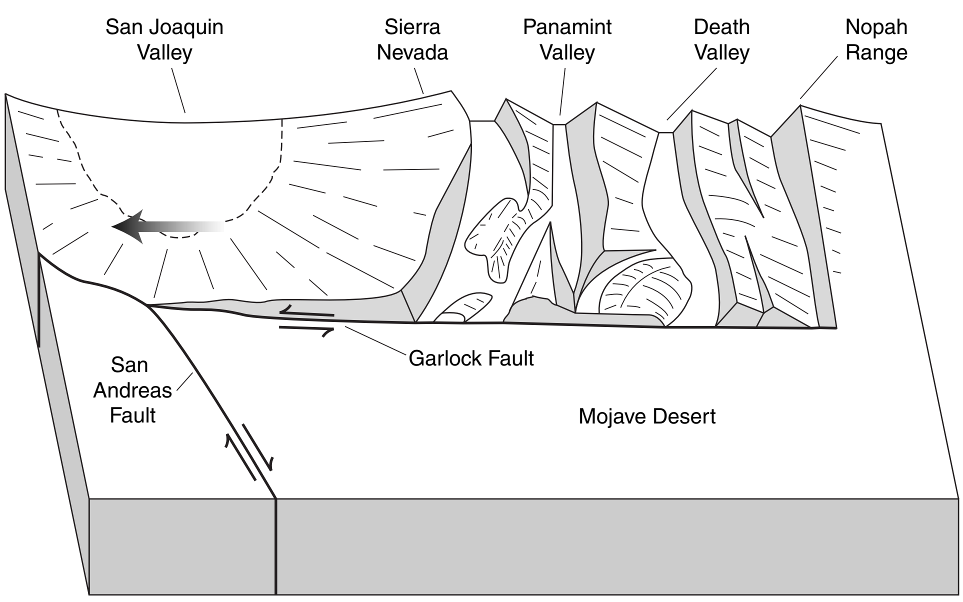


FIGURE 19.28 Block model of the Garlock Fault, southern California (USA), accommodating the offset from the extended Basin and Range Province to the north and the Mojave Desert to the south.

North Americans are probably familiar with the San Andreas fault zone of California (Figure 19.1), which spans the distance between the Gulf of California rift on the south and the Mendocino oceanic transform on the north. As we noted earlier, this right-lateral fault zone is structurally complex, with numerous splays, many of which are seismically active. Subsidiary structures, such as pull-aparts, flower structures, and transverse ranges, mark locations along its trace where the trace of the fault zone is not exactly a small circle around the Euler pole describing the motion between the North American and Pacific Plates. Also, *en echelon* folds and faults form where strain has not been accommodated by strike-slip displacement. Other important examples of continental transforms include the Alpine Fault of New Zealand, which accommodates motion between the Australian and the Pacific Plates (Figure 19.2), and the Chaman Fault of Pakistan, which accommodates motion between India and Asia on the west side of India (Figure 19.23).

19.5 OCEANIC TRANSFORMS AND FRACTURE ZONES

All mid-ocean ridges (i.e., plate boundaries along which seafloor spreading occurs) consist of non-coplanar segments that range in length from tens to hundreds of kilometers. Each segment is linked to its neighbor by an oceanic transform fault. These **oceanic transform faults** are plate boundaries; strike-slip deformation within them accommodates the motion of one oceanic plate laterally past another as seafloor spreading progresses (Figure 19.7). The length of oceanic transform faults varies from 10 to 1000 km. As measured across strike, they are relatively narrow (less than a few kilometers wide); but even considering this narrow width, they are so abundant that between 1% and 10% of the oceanic lithosphere has been affected by deformation related to strike-slip faulting.

It is important to remember that transform faults occur only between ridge segments; they do not extend beyond them. There are, however, pronounced topographic lineaments, known as **fracture zones,** that extend beyond the tips of the ridge segments. Seismicity in the ocean floor shows that earthquakes occur along the ridge segments and along the transform faults between them, but that they are rare along fracture zones. Thus, fracture zones are not active fault zones.[6] Note also that when you cross a transform boundary, you pass from one plate to another. However, when you cross a fracture zone, you stay on the same plate.

Satellite gravity measurements, side-scan sonar images, dredge hauls, cores, and submarine photographs lead to the conclusion that oceanic transform zones and fracture zones are not featureless lines on the surface of the sea floor. Rather, they contain escarpments, ridges, and narrow troughs (Figure 19.29). The bathymetric complexity of oceanic transform zones and fracture zones develops in response to a variety of phenomena. First, since the depth of ocean floor depends on the age of the underlying lithosphere (see Chapter 16) and the ocean floor on one side of a zone is not the same age as the ocean floor on the other side (except at the point on a transform fault halfway between two ridge tips), there must be a change in ocean-floor depth across a zone. Second, in places where transforms are not precisely small circles around an Euler pole (see Chapter 14), transpression and transtension generate a flower structure; the flower structure becomes inactive once a transform zone becomes a fracture zone, but the faults comprising it do not disappear. Third, slip along a transform zone pervasively fractures the crust. Seawater circulates through the fracture network and reacts with crustal basalt, altering olivine to form a hydrated mineral called serpentine. This process increases the volume of the crust, because serpentine is less dense than olivine, and thus causes the sea floor to rise. Fourth, heat radiating from the tip of a ridge segment can cause isostatic uplift of the lithosphere beyond the tip. This uplifted lithosphere gradually drifts away from the ridge termination, as sea-floor spreading progresses, forming an elongate ridge (an intersection high) bordering the fracture zone (Figure 19.29). At a distance from the ridge, the intersection high cools and subsides.

Notably, the development of escarpments in oceanic transform zones and fracture zones, coupled with the weakening of rock by pervasive fracturing and serpentinization, sets the stage for submarine slope failure. The debris that tumbles down the escarpments collects

[6]Some authors, therefore, use the term "inactive transform," instead of fracture zone, for these lineaments.

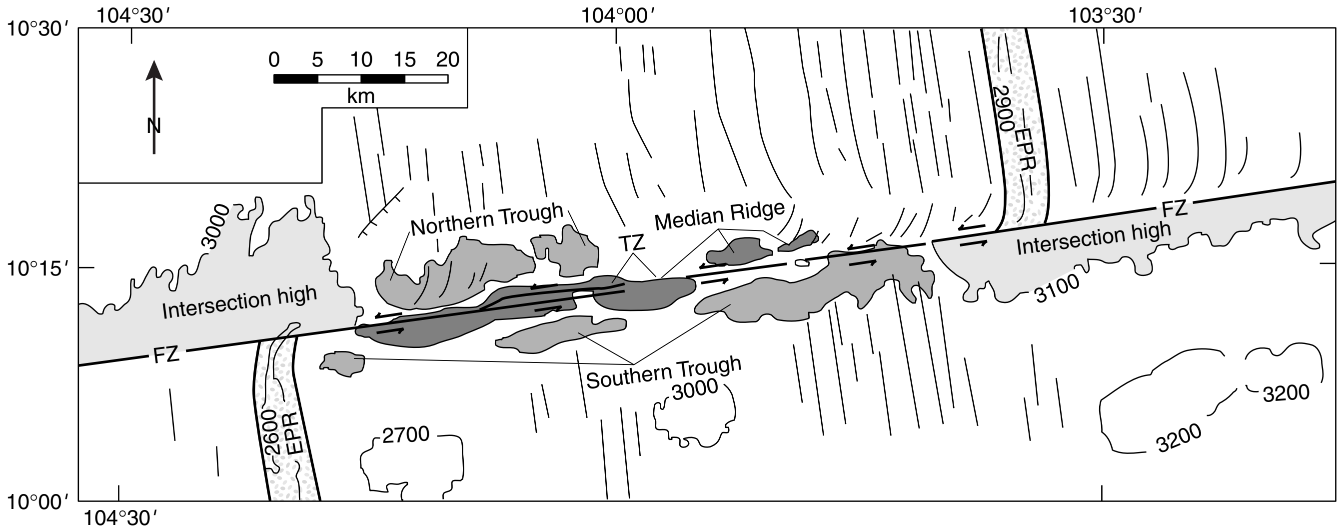


FIGURE 19.29 Topography of the Clipperton fracture zone (FZ) and transform zone (TZ) of the East Pacific Rise. Note intersection highs at ridge tips, and trough and ridges along the transform zone. Contours in meters below sea level. EPR = East Pacific Rise.

in thick piles consisting of sedimentary breccia. Thus, in contrast to normal ocean crust, the oceanic crust of transform zones and fracture zones typically has a coating of sedimentary breccia.

Examination of a map showing global plate boundaries (Figure 14.14a) reveals that not all oceanic transform faults offset ridge segments. Some are plate boundaries that link subduction zones, or link subduction zones to ridges. Examples include the north and south border of the Scotia Plate between Antarctica and South America, and the northern border of the Caribbean Plate. The Alpine Fault of New Zealand, which we discussed earlier, connects along strike to oceanic transforms.

19.6 CLOSING REMARKS

With this chapter on strike-slip tectonics, we close our discussion of tectonic settings. We have focused on describing processes and terminology, with the goal of developing a framework of concepts that we can use to describe tectonic settings. You will have noticed that we mention several features in more than one of these chapters. This necessary and appropriate redundancy emphasizes that classes of structures are not limited to one tectonic setting. In the remainder of this book, we use our structural and tectonic framework to examine specific orogens.

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