CHAPTER SIXTEEN

Rifting, Seafloor Spreading, and Extensional Tectonics

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16.1 INTRODUCTION

Eastern Africa, when viewed from space, looks like it has been slashed by a knife, for numerous ridges, bounding deep, gash-like troughs, traverse the landscape from north to south (Figure 16.1a and b). This landscape comprises the East African Rift, a region where the continent of Africa is quite literally splitting apart—land lying on the east side of the rift moves ever so slowly to the east, relative to land lying on the west side of the rift. If this process were to continue for another 10 or 20 million years, a new ocean would form between the two land masses. Initially, this ocean would resemble the Red Sea, which now separates the Arabian Peninsula from Africa. Eventually, the ocean could grow to be as wide as the Atlantic, or wider.

The drama that we just described is an example of rifting. Simply put, **continental rifting** (or simply "**rifting**") is the process by which continental lithosphere undergoes regional horizontal extension. A **rift** or **rift system** is a belt of continental lithosphere that is currently undergoing extension, or underwent extension in the past. During rifting, the lithosphere stretches with a component roughly perpendicular to the trend of the rift; in an oblique rift, the stretching direction is at an acute angle to the rift trend.

Geologists distinguish between active and inactive rifts, based on the timing of the extension. **Active rifts,** like the East African Rift, are places on Earth where extension currently takes place (Figure 16.1). In active rifts, an array of recent normal faults cuts the crust, earthquakes rumble with unnerving frequency, and volcanic eruptions occasionally bury the countryside in ash and lava. The faulting taking place in active rifts yields a distinctive topography characterized by the occurrence of linear ridges separated by nonmarine or shallow-marine sedimentary basins (Figure 16.2). In **inactive rifts,** places where extensional deformation ceased

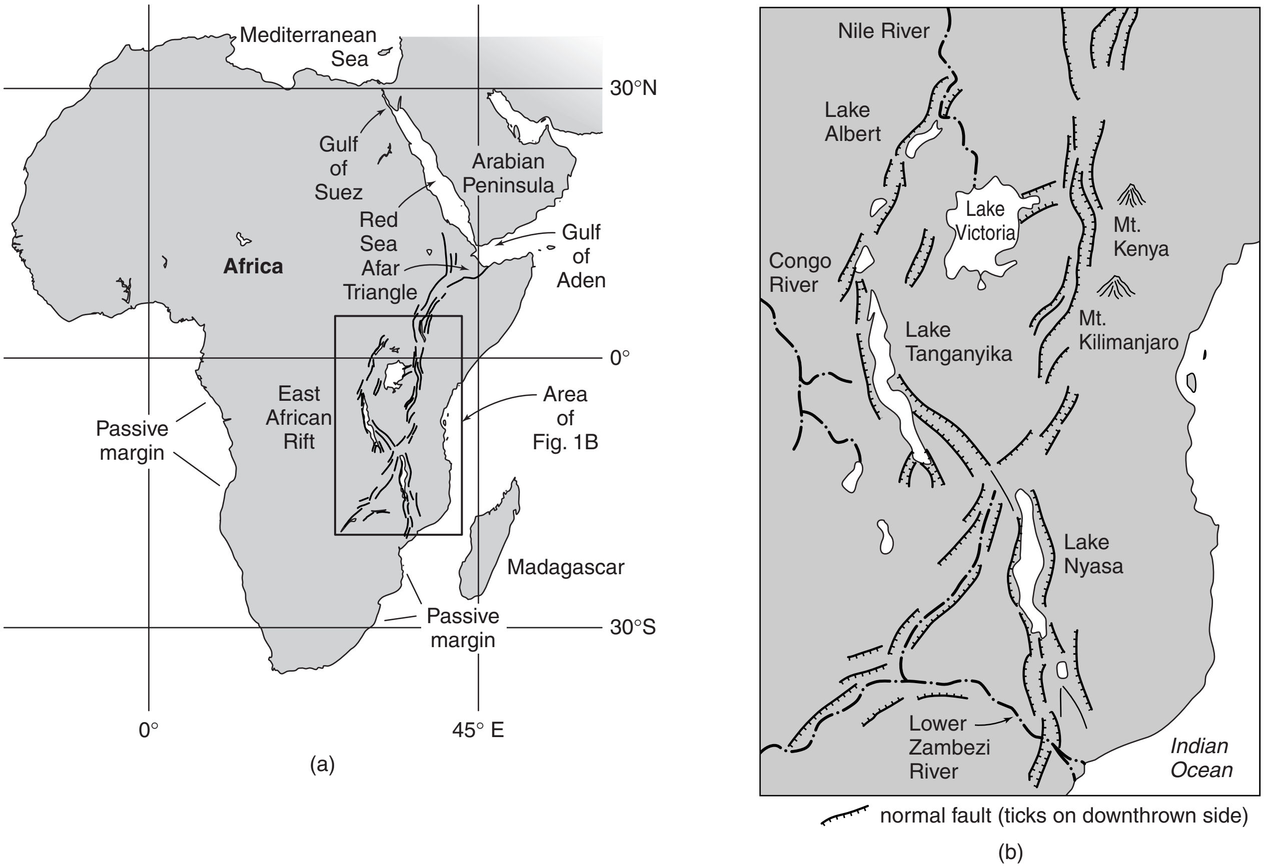


FIGURE 16.1 (a) Map of Africa, showing the East African Rift, the Red Sea, and the Gulf of Aden. As we see in East Africa, a continental rift consists of a belt of normal faults, bounding deep troughs. Some of the troughs may fill with water to become lakes. If the rift is "successful," a narrow ocean basin, like the Red Sea or Gulf of Aden, develops. The Afar Triangle lies at the triple junction between the Red Sea, the Gulf of Aden, and the East African Rift. If seafloor spreading continues for a long time, the narrow ocean could grow to become a large ocean, like the Atlantic. The passive continental margins bordering the ocean basin are underlain by stretched continental crust. (b) Detail of the East African Rift, illustrating the approximate fault pattern.

FIGURE 16.2 Photo of the Gregory Rift in southwest Kenya. Note the escarpments. These are the eroded footwalls of normal faults. Floods bring in sediment to bury the surface of the hanging-wall block. Magmatic activity associated with rifting has led to the growth of the volcano on the left.

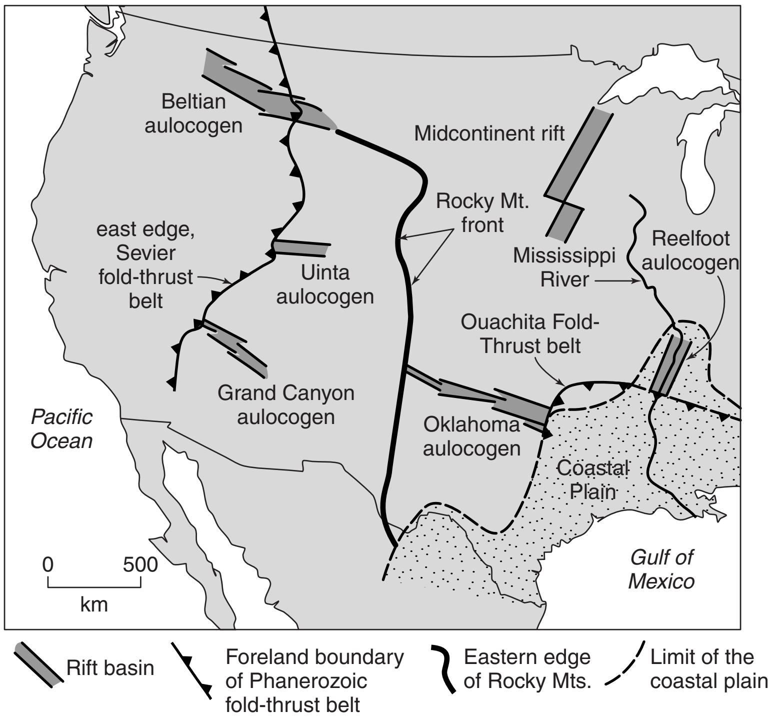


FIGURE 16.3 Sketch map of the western two-thirds of the United States showing inactive Proterozoic rifts that trend at a high angle to North America's Phanerozoic orogens and continental margins. Because these rifts cut into the continent at a high angle, they are called aulacogens. They may be relicts of unsuccessful rifting events that occurred prior to the successful rifting that generated the margins of North America at the beginning of the Phanerozoic. Late Proterozoic successful rifts of the western United States trended north-northeast, forming a north-northeast trending, passive margin. Thus, Phanerozoic convergent-margin tectonism, along this margin, generated orogens at high angles to the aulacogens.

some time ago, we find inactive normal faults and thick sequences of redbeds, conglomerates, evaporites, and volcanics. A preserved inactive rift can also be called an **unsuccessful rift,** in that its existence reflects the occurrence of a rifting event that stopped before it succeeded in splitting a continent in two. Unsuccessful rifts that cut into cratonic areas of continents, at a high angle to the continental margin, are known as **aulacogens** (from the Greek for "furrow"; Figure 16.3).

A **successful rift** is one in which extensional deformation completely splits a continent into two pieces. When this happens, as we have already noted, a new **mid-ocean ridge** (oceanic spreading center) forms between the now separate continent fragments, and seafloor spreading produces new oceanic lithosphere. Typically, 20 to 60 million years pass between inception of a rift and the time (called the **rift–drift transition**) at which active rift faulting ceases and seafloor spreading begins. The amount of lithospheric stretching that takes place prior to the rift–drift transition is variable. Typically, the continental lithosphere stretches by a factor of 2 to 4 times before separation, meaning that the lithosphere of the rift region eventually becomes 2 to 4 times its original width and about one half to one quarter of its original thickness (Figure 16.4). The amount of stretching prior to rifting at a given locality, as well as the overall width of the rift, depends largely on the pre-rift strength of the lithosphere. Rifts formed in old, cold, strong shields tend to be narrow, while rifts formed in young, warm, soft orogens tend to be wide.[1] Once active faulting ceases in a successful rift, the relicts of the rift underlie the continental margins on either side of a new ocean basin. Since no tectonic activity happens along such continental margins after they have formed, we refer to them as **passive margins.** The inactive, rifted crust of passive margins slowly subsides (sinks) and gets buried by sediment eroded from the bordering continent.

[1]For example, the Midcontinent Rift, a 1.1 Ga rift that ruptured the old, cold craton of the central United States, is much narrower than the Basin and Range Province, a Cenozoic rift that formed in the warm, soft North American Cordillera immediately following a protracted period of Mesozoic and Cenozoic convergent tectonism and associated igneous activity.

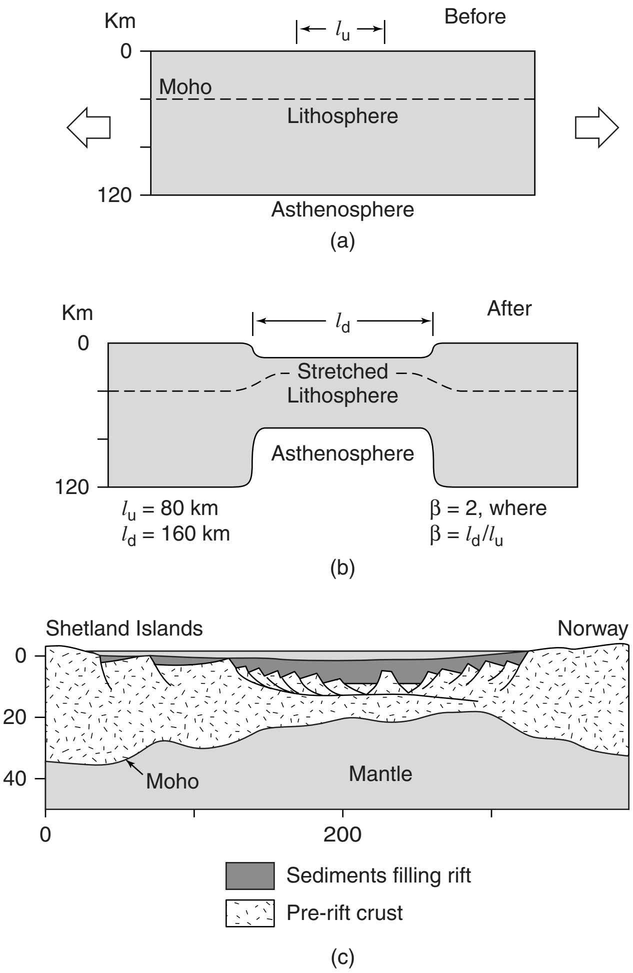


FIGURE 16.4 Illustration of the concept that rifting leads to stretching of the lithosphere. (a) This cross section shows that before rifting, the unstretched lithosphere is 120 km thick. The region that will evolve into the rift is 80 km wide. We call this the undeformed length (l_u). (b) After rifting, the lithosphere in the rift has stretched and thinned, so its deformed length (l_d) is 160 km, and it is 60 km thick. We can represent the strain that results from rifting by a number, β, called the **stretching factor.** In this example, $\beta = l_d/l_u = 2$. The larger the value of β, the greater the amount of stretching and thinning. A value of β = 2 means that the lithosphere has thinned by 50%. (c) In reality, the crust portion of the lithosphere stretches by developing normal faults. This simplified cross-sectional sketch through the Viking Graben (North Sea) shows the faulting in the crust, and the thinning of the crust. (Note: The base of the lithosphere is not shown.) The depression formed by rifting has filled with sediment.

Rifts and passive margins are fascinating regions, and geologists have worked for decades to decipher the complex structural, stratigraphic, and igneous assemblages that they contain. They are also important from a practical standpoint, because they contain significant petroleum resources. In this chapter, we survey the principal features of rifts and passive margins (i.e., of extensional tectonism) and will introduce current speculations about how and why rifting occurs. Table 16.1 summarizes basic rift terminology, for quick reference.

16.2 CROSS-SECTIONAL STRUCTURE OF A RIFT

Let's start our discussion of rifts by developing a cross-sectional image of an active rift. Rifts evolve with time, so the geometry of a rift at a very early stage in development differs from that of a rift just prior to the rift–drift transition. We'll examine rift evolution later, but for now, we'll focus on the geometry of a rift that is at an intermediate stage in its development.

16.2.1 Normal Fault Systems

Rifts are regions in which extensional tectonics (i.e., stretching of continental lithosphere) takes place. Recall from Chapter 6 that, in the brittle field, extensional strain can be accommodated by slip on a "system" (group or array) of normal faults. On pre–1970s cross sections of rifts, geologists implied that normal-fault systems are symmetric, in that the borders of the rift were defined by normal faults that dipped toward the interior of the rift. In this **symmetric rift model,** pairs of normal faults dipping toward each other outline **grabens,** while pairs of normal faults dipping away from each other outline **horsts** (Figure 16.5a). Note that horsts and grabens are bounded on *both* sides by faults and that in old symmetric rift models, horsts formed mountain ridges ("ranges"), and the grabens underlay sediment-filled troughs ("basins"). At depth, in old cross sections, faults simply die out, either in a zone of distributed strain or in a cluster of question marks.

Modern seismic-reflection surveys of rifts do not agree with the above image, but rather show that most rifts are asymmetric. In this **asymmetric rift model,** upper-crustal extension is accommodated by displacement on arrays of subparallel normal faults, most of which dip in the same direction. These faults merge at depth with a regional subhorizontal **basal detachment** (Figure 16.5b). The fault that defines the boundary of the rift, where the basal detachment curves up and reaches the ground surface, is the **breakaway fault.**

TABLE 16.1	TERMINOLOGY OF EXTENSIONAL TECTONICS
Accommodation zones	Normal fault systems are not continuous along the length of a rift. Rather, rifts are divided into segments, whose axes may be offset from one another. Further, the faults of one segment may dip in the opposite direction to the faults of another segment. An accommodation zone is the region of complex structure that links the ends of two rift segments. Accommodation zones typically include strike-slip faults.
Active margin	A continental margin that coincides with either a strike-slip or convergent plate boundary, and thus is seismically active.
Aulacogen	An unsuccessful rift that cuts across a continental margin at a high angle to the margin. Typically, aulacogens transect the grain of an orogen that borders the margin. Aulacogens may represent failed arms of three-armed rifts, or they may simply be older rifts (formed long before the development of the continental margin, during an earlier episode of rifting at a different orientation) that were cut off when the margin formed (Figure 16.3).
Axis (of rift or MOR)	The center line along the length of a rift or a mid-ocean ridge (MOR). The trend of the axis is the overall trend of the rift.
Breakaway fault	The normal fault that forms the edge of the rift. (A breakaway fault forms the boundary between stretched and unstretched crust).
Graben	A narrow, symmetric trough or basin, bounded on both sides by normal faults that dip toward the center of the trough.
Half graben	An asymmetric basin formed on the back of a tilted fault block; one border of the basin is a normal fault.
Horst	An elongate, symmetric crustal block bordered on both sides by normal faults; both faults dip away from the center of the horst.
Listric normal fault	A normal fault whose dip decreases with depth, thereby making the fault surface concave upward.
Midocean ridge	The elongate submarine mountain range that is the bathymetric manifestation of a divergent plate boundary. Though some midocean ridges (e.g., the Mid-Atlantic Ridge) do lie in the center of ocean basins, some (e.g., the East Pacific Rise) do not. Therefore, some geologists use the term "oceanic ridge" in place of "mid-ocean ridge" for these features.
Oblique rifting	Rifting that occurs where the stretching direction is at an acute angle to the rift axis.
Passive margin	A continental margin that is not a plate boundary and, therefore, is not seismically active. It is underlain by the relict of a successful rift. The rift relict subsides and is buried by a thick wedge of sediment.
Planar normal fault	A normal fault whose dip remains constant with depth.
Nonrotational normal fault	A normal fault on which slip does not result in rotation of the hanging-wall block.
Rift (rift system)	A belt of continental lithosphere that is undergoing, or has undergone, extensional deformation (i.e., stretching); also called a continental rift.
Rift–drift transition	The time at which active rift faulting ceases and seafloor spreading begins (i.e., the time at which a mid-ocean ridge initiates, and the relicts of a rift become the foundation of a passive margin).
Rifting	The process by which continental lithosphere undergoes extensional deformation (stretching) by the formation and activity of normal faults.
Rotational normal fault	A normal fault whose hanging wall block rotates around a horizontal axis during slip.
Subsidence	The sinking of the surface of the lithosphere. Subsidence produces sedimentary basins. For example, the relict of a successful rift subsides to form a passive-margin basin.
Successful rift	A rift in which stretching has proceeded until the continent cut by the rift ruptures to form two pieces separated by a new mid-ocean ridge.
Transfer fault	A dominantly strike-slip fault that links two normal faults that are not coplanar; some transfer faults serve as accommodation zones.
Unsuccessful rift	A rift in which extensional deformation ceased prior to rupture of the continent that was cut by the rift.

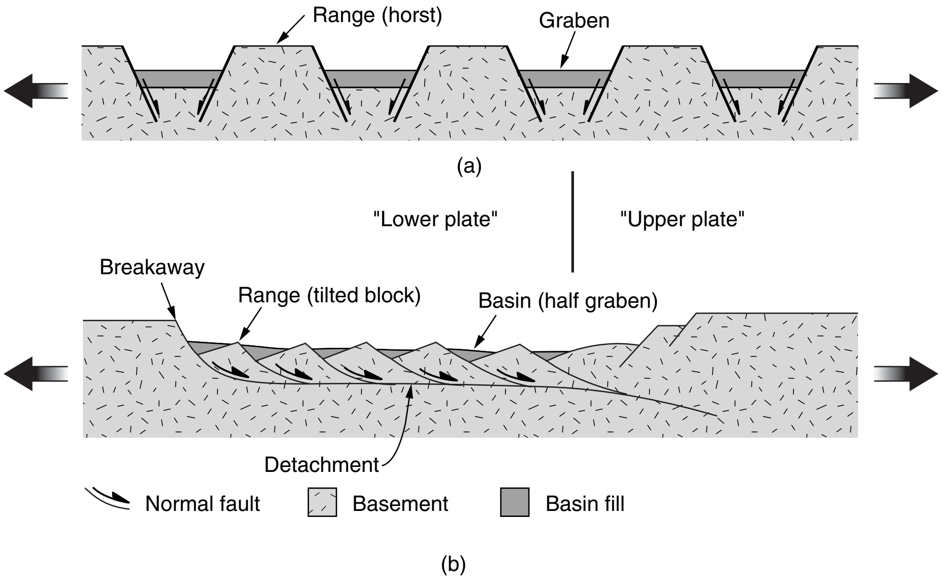


FIGURE 16.5 Contrasting models of rift faulting. (a) Cross section illustrating the old concept of symmetric horsts and grabens. (b) Cross section illustrating the contemporary image of tilted fault blocks and half grabens, all above a detachment.

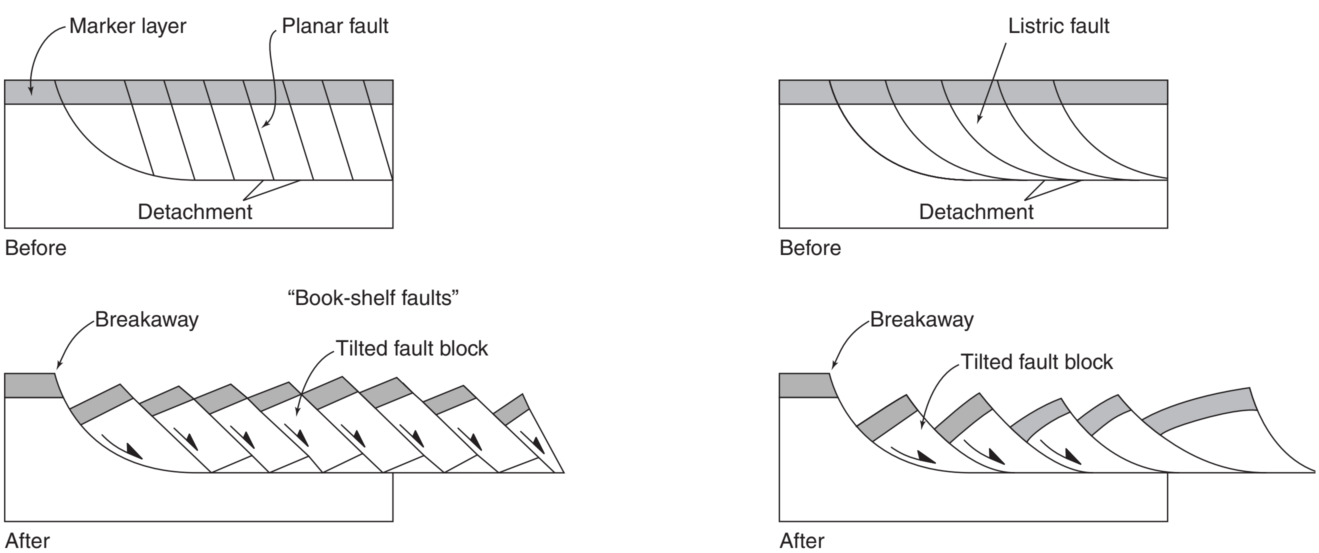


FIGURE 16.6 Geometry of normal fault arrays, in cross section. (a) Parallel rotational faults. Before faulting (top), the faults are parallel and not curved. After faulting (bottom), fault blocks are tilted. In reality, crushing and small-scale faulting at the base of the blocks fill the gaps. (b) Listric faults and associated unfaulted rollover. Before faulting (top), the faults shallow with depth and merge with a detachment. After faulting (bottom), the blocks have moved. The block to the right curves down to maintain contact with the footwall, forming a rollover anticline.

Rifts contain both **planar normal faults,** meaning faults whose dip remains constant with depth (Figure 16.6a), and **listric normal faults,** meaning faults whose dip decreases with depth, thereby making the fault surface concave upward (Figure 16.6b). Movement on listric normal faults results in rotation of the hanging-wall block, so that this block progressively tilts during regional extension, with the amount of tilting proportional to the amount of displacement on the fault. Because of the curvature of listric faults,

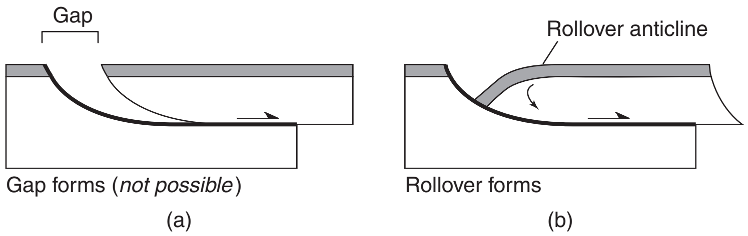


FIGURE 16.7 Formation of a rollover anticline above a listric normal fault. (a) Cross section showing that if the hanging-wall block moves to the right, without bending over, a gap develops. (b) Gravity pulls the lip of the hanging-wall block down, to maintain contact with the footwall, forming a rollover anticline.

movement of the hanging-wall block along a subhorizontal detachment creates a gap (Figure 16.7a). In Earth's gravity, rock isn't strong enough for such gaps to remain open, so either the hanging wall breaks up into fault slices (Figure 16.6b), or it sinks downward and curves to form a rollover anticline (Figure 16.7b).

Though slip on planar normal faults can occur without rotation (i.e., making the faults **nonrotational normal faults**), planar faults typically occur in arrays above a basal detachment; so as slip on the system takes place, the blocks between the faults do progressively rotate. As a result, the blocks tilt, yielding a geometry that resembles a shelf-full of tilted books. As displacement increases, the tilt increases. Faults in such arrays are sometimes referred to as **book-shelf faults.** Any normal fault (listric or planar) on which tilting accompanies displacement can be considered to be a **rotational normal fault.**

Individual normal faults may consist of subhorizontal segments **(flats)** linked by steeply dipping segments **(ramps),** producing a staircase geometry in profile. During displacement on a listric or stair-step normal fault, strata of the hanging wall can bend to form a **rollover fold** (either a rollover anticline or a rollover syncline, depending on the geometry of the underlying fault; Figure 16.8a). Synthetic and antithetic faults may also develop in the hanging-wall block (Figure 16.8b). Recall that a **synthetic fault** is a secondary fault that dips in the same direction as a major fault and that an **antithetic fault** is one that dips in the direction opposite to that of the major fault. In an **imbricate array of normal faults,** adjacent faults die out updip or break the ground surface, whereas in an **extensional duplex,** adjacent ramps merge with the same upper and lower flats (Figure 16.8c). In real rifts, combinations of stair-step faults, imbricate arrays, duplexes, rollovers, antithetic faults, and synthetic faults yield a very complex geometry. This geometry can become even more complex in regions where deposition occurs during faulting (because different stratigraphic levels

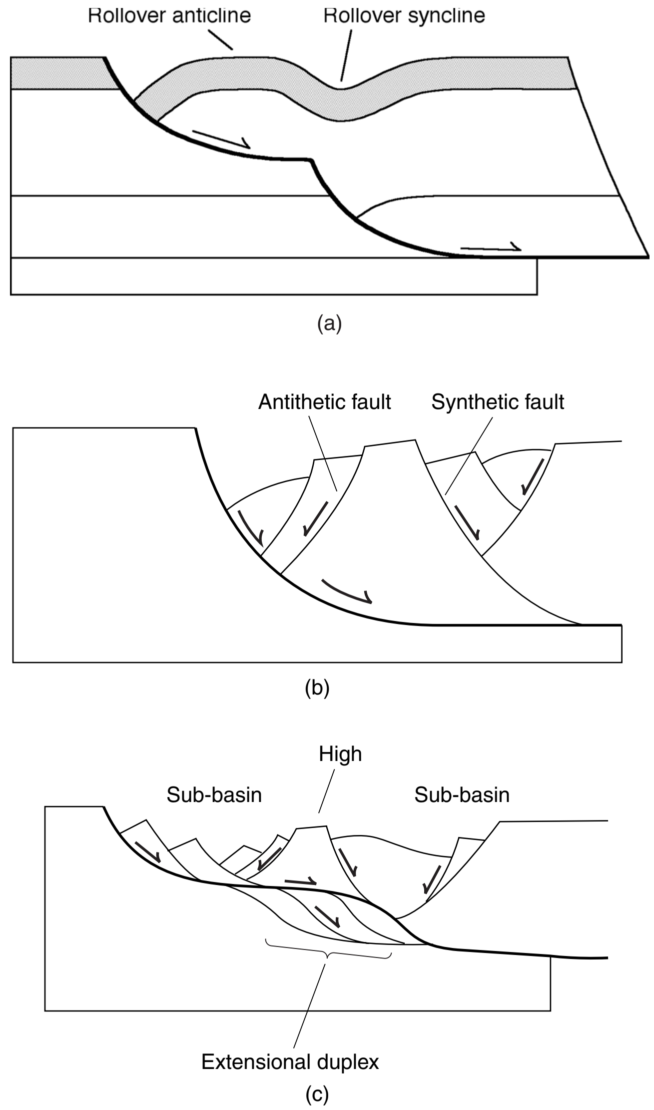


FIGURE 16.8 Complex fault systems and related folds found in rifts. (a) Cross section showing a rollover anticline above a listric normal fault, and a rollover syncline forming at the intersection of a ramp and flat. (b) Here, antithetic faults (dipping toward the main fault) and synthetic faults (dipping in the same direction as the main fault) break up the hanging-wall block. (c) Complex fault system underlain by an extensional duplex. Note the sub-basins and the high block between them.

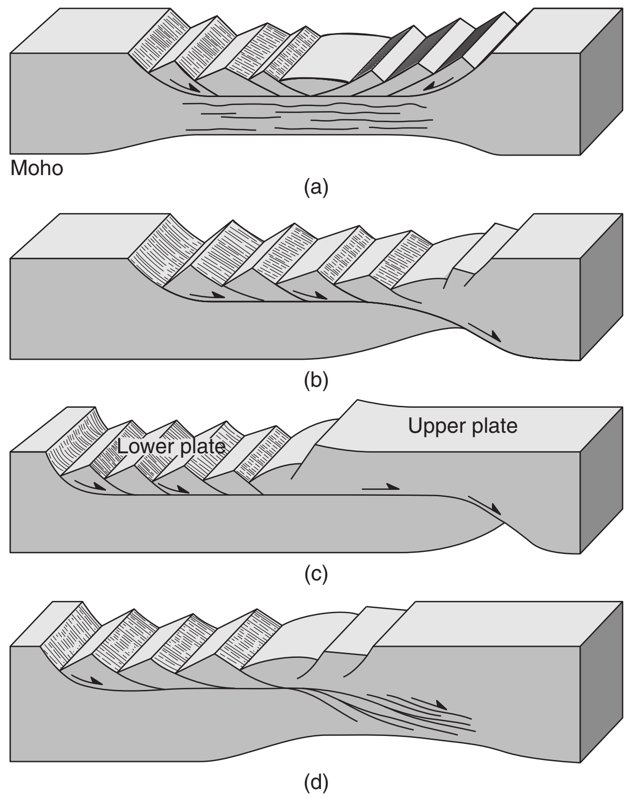


FIGURE 16.9 Models of rifting at the crustal scale. (a) Pure-shear model; (b) simple-shear model; (c) delamination model; (d) hybrid model (simple shear plus broad zone of distributed shear at depth).

are offset by different amounts), and/or where salt movement accompanies rifting (because salt may rise diapirically into overlying strata).

16.2.2 Pure-Shear versus Simple-Shear Models of Rifting

We've concentrated, so far, on the structural geometry of rifts at shallow levels in the continental crust. What do rifts look like at greater depths? In the **pure-shear model** (Figure 16.9a), the detachment defining the base of upper-crustal normal faulting lies at or near the brittle–plastic transition in the crust. Beneath this detachment, the crust accommodates stretching across a broad zone, either by development of penetrative plastic strain, or by movement on an array of anastomosing (braided) shear zones. Geologists apply the name "pure-shear model" to this geometry because a square superimposed on a cross section of the crust prior to extension becomes a rectangle after rifting, due to shortening in the vertical direction and stretching in the horizontal direction.

In the **simple-shear model** (Figure 16.9b), by contrast, the basal detachment cuts down through the crust, and perhaps deeper, as a discrete shear zone. In some versions of this model, the detachment may be subhorizontal for a substantial distance beyond the edge of the rift before bending down, so that the region of upper-crustal extension does not lie directly over the region of deeper extension (this is sometimes called the "delamination model"; Figure 16.9c). Geologists refer to the portion of the detachment that traverses the crust as a **transcrustal extensional shear zone,** or **translithosphere extensional shear zone,** depending on how deep it goes. Because the deeper portions of the shear zone involve portions of the crust where temperatures and pressures are sufficiently high for plastic deformation mechanisms to operate, movement in these zones yields mylonite. In a "combination model," the transcrustal extensional shear zone of the simple-shear model spreads at depth into a diffuse band of anastomosing shear zones and disappears in a zone of distributed strain in the lower crust, and lithosphere beneath the detachment stretches penetratively (Figure 16.9d).

Simple-shear models have the advantage of readily explaining the asymmetry of many rifts; one side of the rift, called the **lower plate,** undergoes extreme stretching, while the other side, called the **upper plate,** undergoes less stretching (Figures 16.9 and 16.10). Quite likely, some combination of pure-shear and simple-shear models represents the actual geometry of rifts.

16.2.3 Examples of Rift Structure in Cross Section

To conclude our discussion of rift structure in cross section, let's look at two documented examples of rifts. In the first example, we look at the Viking Graben in the North Sea Rift. This graben separates the United Kingdom from the mainland of Europe. The second example, the Gulf of Suez, lies between Egypt and the Sinai Peninsula.

The North Sea Rift formed during the Mesozoic and Early Cenozoic, and has been studied intensively because it contains valuable petroleum reserves. A cross-sectional interpretation of the northern portion of the rift, a region called the Viking Graben (Figure 16.11a), shows that the base of the rift has a stair-step profile. Numerous tilted fault blocks underlie half grabens, which have filled with sediment. Some of the deposition is syntectonic, meaning that it occurred as faulting progressed. Note that layers of **syntectonic strata** thicken toward the normal fault that bounds one edge of the half graben and that they have themselves been tilted as fault blocks rotated during later stages of extension. Because the faults grow updip as more sediment accumulates, they are also called **growth faults.**

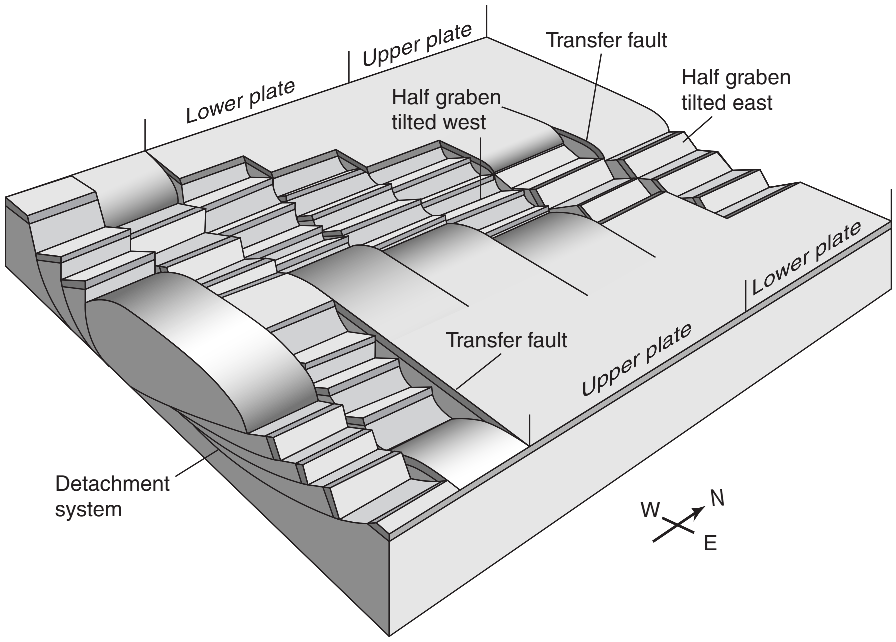


FIGURE 16.10 Block diagram illustrating the concept of upper-plate and lower-plate parts of rifts. Note how the asymmetry of rifts changes at transfer faults (accommodation zones).

In places, hanging-wall blocks contain rollover folds, synthetic faults, and antithetic faults. The magnitude of extensional strain is not uniform across the rift. Thus, this rift includes more than one **rift sub-basin.** A rift sub-basin is a portion of a rift separated from another adjacent portion by a **basement high,** a horst in which the crust has not been stretched as much. The Bergen High in Figure 16.11a is an example of a basement high. The Horda basin is a rift sub-basin.

The Gulf of Suez formed in the Cenozoic during a phase of northward propagation of the Red Sea Rift. According to the cross section interpretation of Figure 16.11b, the rift is very asymmetric, in that most of the faults dip eastward at the latitude of the cross section. Extension lies to the east of a distinct breakaway fault. In the Gulf of Suez, it appears that after the original breakaway had been active for some time, it became inactive as a new breakaway developed to the southwest. This younger breakaway cuts down to a deeper crustal level.

16.3 CORDILLERAN METAMORPHIC CORE COMPLEXES

The simple-shear model of rifting provides a possible explanation for the origin of enigmatic geologic features now known as **Cordilleran metamorphic core complexes** (sometimes referred to as “metamorphic core complexes” or, simply, “core complexes”). Cordilleran metamorphic core complexes are found in a belt that rims the eastern edge of the Cenozoic Basin and Range Province (Figure 16.12), a broad rift that occurs in portions of Idaho, Utah, Nevada, California, Arizona, and Sonora in the North American Cordillera. They were originally named because they are distinct domes of metamorphic rock that occur in the “core” (or hinterland) of the orogen (Figure 16.13). More recently, the phrase “metamorphic core” has been used by geologists in reference to the metamorphic rocks exposed at the center in the domes. An idealized Cordilleran metamorphic core complex includes the following features:

- The interior consists of nonmylonitic country rock (gneiss, sedimentary strata, or volcanics), locally cut by Cenozoic or Mesozoic granite.
- A carapace of mylonite, formed by intense shearing, surrounds the nonmylonitic interior (Figure 16.13). This rock is very fine grained, has very strong foliation and lineation, and contains rootless isoclinal folds whose axial planes are parallel to the foliation.
- Regionally, the mylonite carapace arches into a gentle dome, shaped somewhat like a turtle shell. Notably, the mineral lineation in the mylonite has the same bearing regardless of the dip direction of the foliation. Commonly, shear-sense indicators in the mylonite (e.g., rotated porphyroclasts, C-S

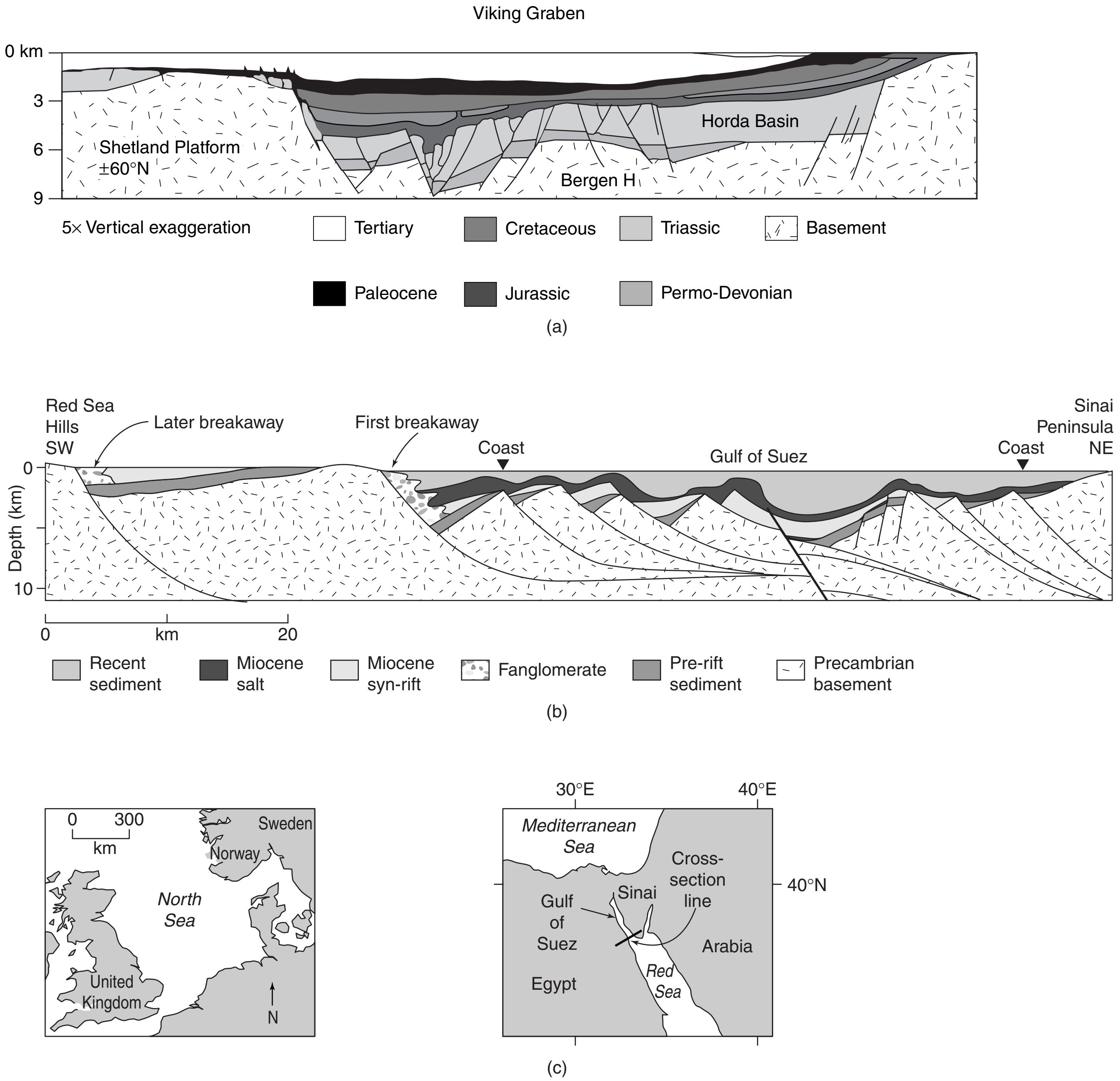


FIGURE 16.11 (a) Cross section of the Viking Graben, in the North Sea Rift. Note that the cross section is at 5 × vertical exaggeration, so the faults look much steeper than they are in nature. Note the basement highs that separate the graben into sub-basins. (b) Cross section of the Gulf of Suez. Note the asymmetry of the rift, in that most faults dip to the northeast; also, note how the original breakaway was abandoned when a later one formed further to the southwest. (c) Location maps showing the North Sea and the Gulf of Suez.

fabric, σ- and δ-tails; see Chapter 12) indicate the same regional displacement sense everywhere in the carapace.

- The basal contact of the mylonite carapace is gradational, such that the degree of mylonitization diminishes progressively downward into the nonmylonitic rocks of the core complex's interior.
- A layer of chloritic fault breccia locally occurs at the top of the mylonite carapace. Rocks above this breccia zone do not contain the mylonitic foliation. The boundary between the chlorite breccia zone and the mylonite can be quite abrupt.
- Movement on rotational normal faults cuts the brittle hanging wall above the chlorite breccia zone,

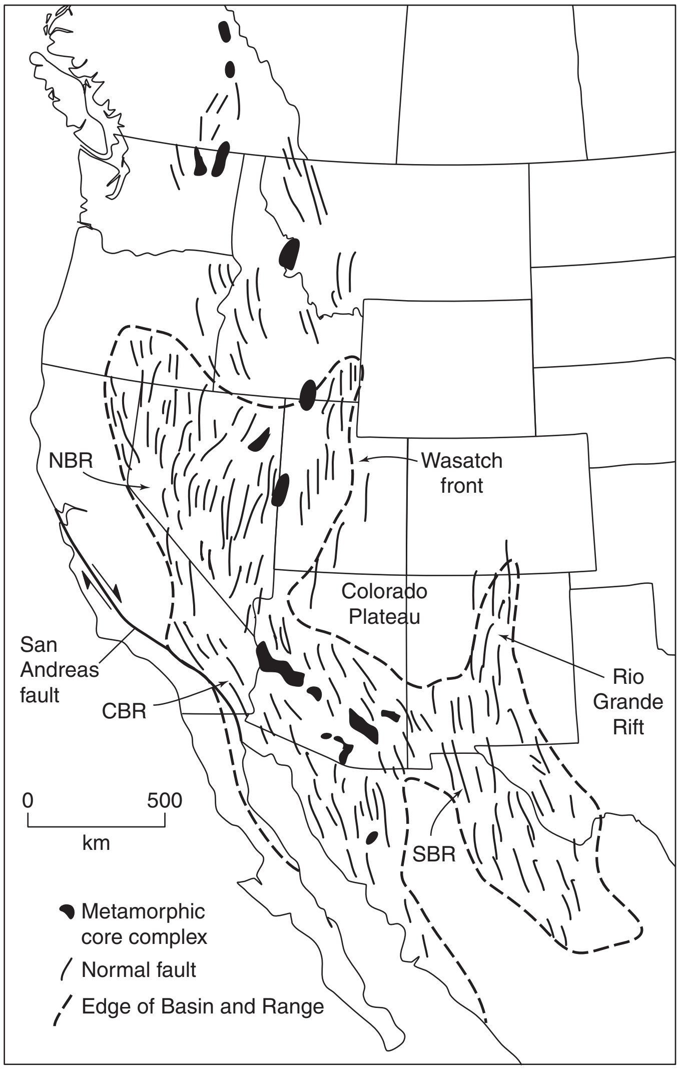


FIGURE 16.12 Simplified geologic map of the Basin and Range Province of the North American Cordillera. The province can be subdivided into three parts: NBR = Northern Basin and Range, CBR = Central Basin and Range, and SBR = Southern Basin and Range. The Rio Grande Rift is an arm of extensional strain that defines the eastern edge of the Colorado Plateau. Metamorphic core complexes are shown in black.

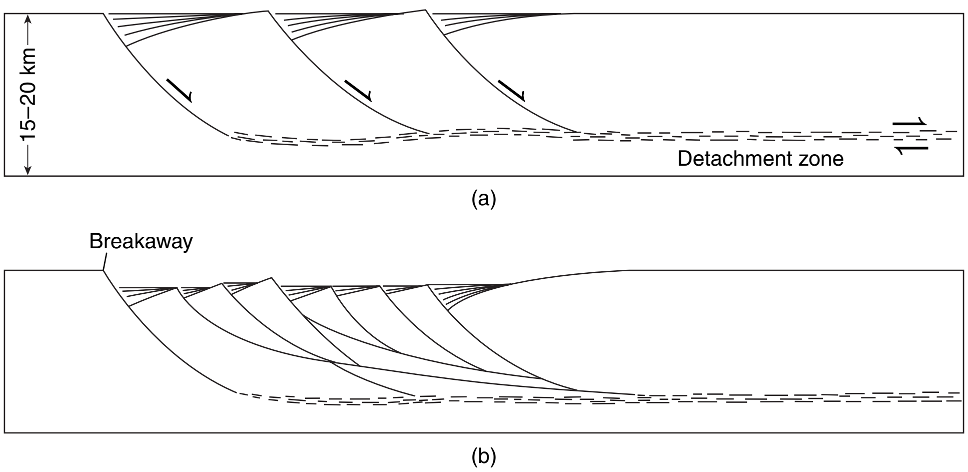


FIGURE 16.13 Idealized cross sections (a–d) showing stages in the development of a metamorphic core complex. (a) An initially subhorizontal, midcrustal ductile detachment zone is formed beneath an array of steeply dipping normal faults in the upper plate; (b) additional normal faults have formed, increasing the geometric complexity; (c) as a result of unloading and isostatic compensation, the lower plate bows upward; (d) extreme thinning of the hanging wall exposes the "metamorphic core" (an exposure of the mylonitic shear zone of the detachment). Some of the hanging-wall blocks have rotated by 90°. (e) Photograph of the contact between tilted fault blocks composed of Tertiary volcanics of the hanging wall and mylonitized basement of the footwall. Whipple Mountains, California (USA).

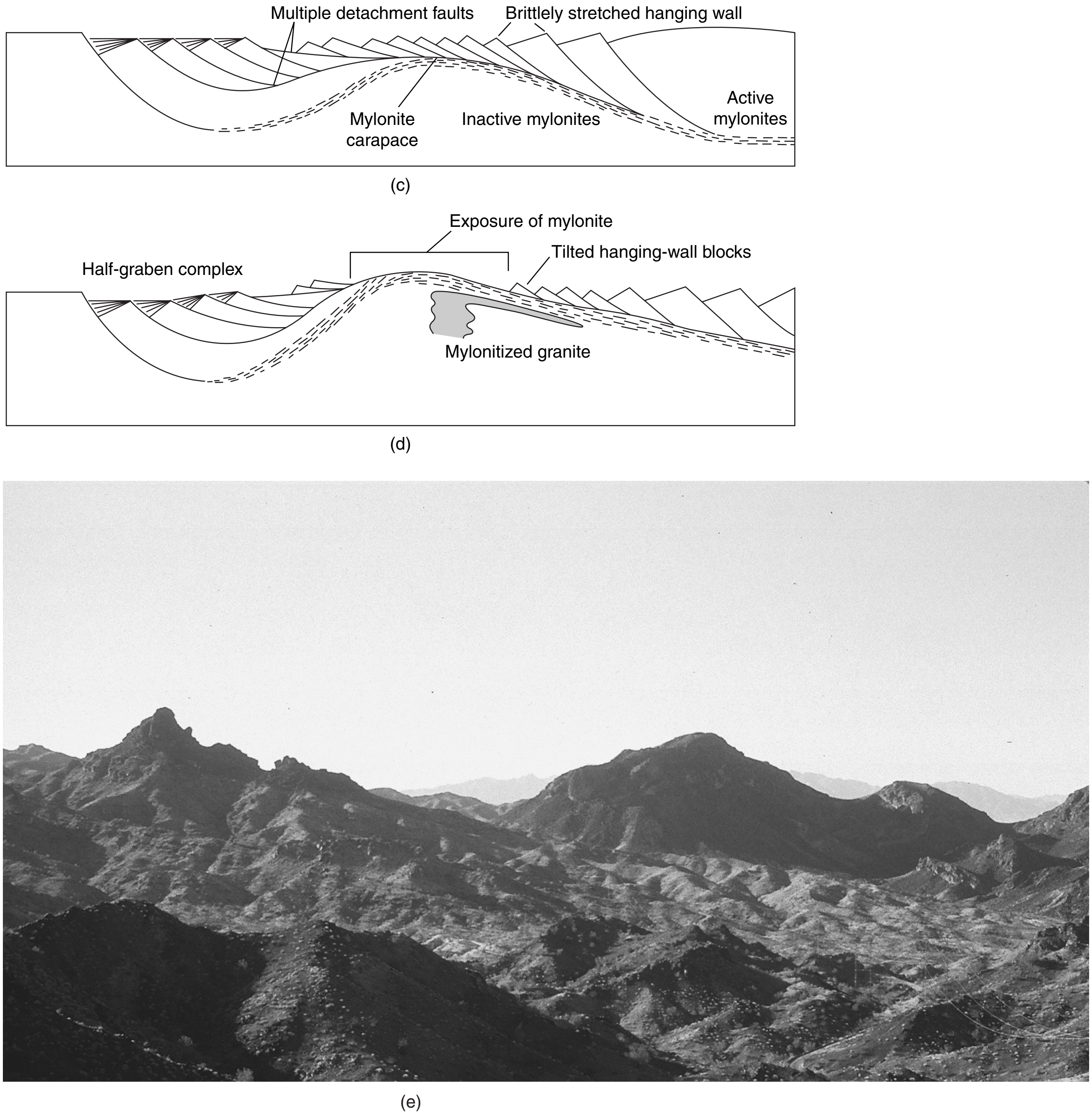


FIGURE 16.13 (Continued)

resulting in large extensional strains in the hanging wall. In fact, locally, faulting has isolated individual blocks of the hanging wall, so that they appear as islands of unmylonitized rock floating on a "sea" of mylonite. In places, stratification in these blocks has been tilted by almost 90°, and thus intersects the mylonite/breccia zone at almost a right angle.

An explanation of how Cordilleran metamorphic core complexes formed remained elusive for many years, until the process of mylonitization, the tools for interpreting shear-sense indicators, and the implications of asymmetric rifting became clear. In the context of these new concepts, geologists consider the complexes to be exposures of the regional detachment at the base of the normal fault system in a rift (Figure 16.13). Specifically, Cordilleran metamorphic core complexes form where normal faulting in a rift has stretched the hanging wall above the rift's basal detachment by so much that it locally thins to zero, so that the footwall beneath the detachment becomes exposed. In this interpretation, the nonmylonitic interior of the core complex consists of footwall rock below the rift's basal detachment, and the mylonitic carapace is an exposure of the detachment itself. Note that the mylonite of the carapace forms at depth in the crust (along a transcrustal extensional shear zone) and then moves up to shallower crustal levels because of the progressive normal-sense motion on the shear zone. The chlorite breccia zone represents the boundary between fault rocks deformed in the brittle field and the mylonite; it forms at shallower depths where deformation leads to brittle failure of rock. Chlorite in this breccia results from retrograde metamorphism accompanying fluid circulation as hot rocks of the footwall are brought up to shallower crustal levels. The mylonitic shear zone arches to form a turtleback-shaped carapace as a consequence of isostatic uplift in response to unloading, when the hanging wall is tectonically thinned and removed. Rocks in the middle crust must flow plastically to fill in the area under the dome.

In sum, Cordilleran metamorphic core complexes represent regions where crustal-scale normal-sense shear on a rift's detachment brings mylonitized footwall rocks up from depth and juxtaposes them beneath brittlely deformed rocks of the hanging wall. The exposed "metamorphic core" represents an exposure of the mylonitic detachment zone, revealed by extreme stretching of the hanging wall.

16.4 FORMATION OF A RIFT SYSTEM

Extension along the entire length of a continent-scale rift system does not begin everywhere at the same time. In a given region, the rift system begins as a series of unconnected **rift segments,** 100–700 km long, each containing a set of normal faults, which die out along their length. The numerous normal faults that comprise an individual rift segment dip dominantly in the same direction. In many locations, individual faults within segments have a C-shaped trace (i.e., are spoon-shaped in three dimensions); at the center of the "C," displacement on the fault is dominantly downdip, whereas at the ends of the C, displacement is strike-slip or oblique-slip (Figure 16.14). The thickness of sediment in the half graben developed above the down-

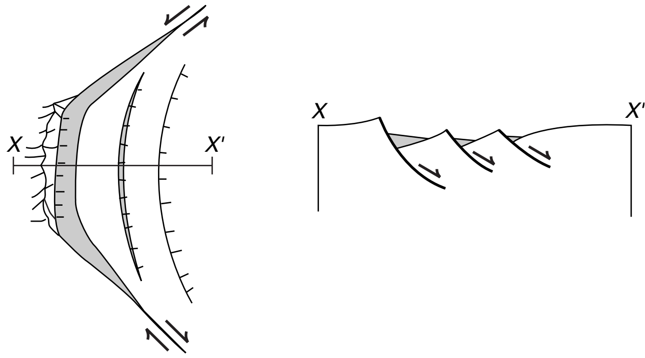
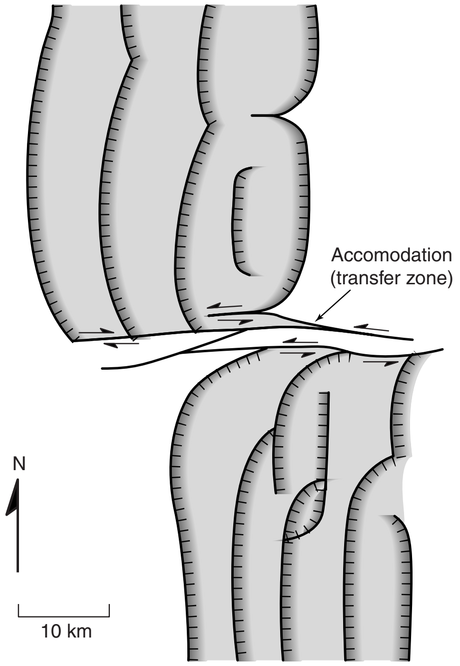


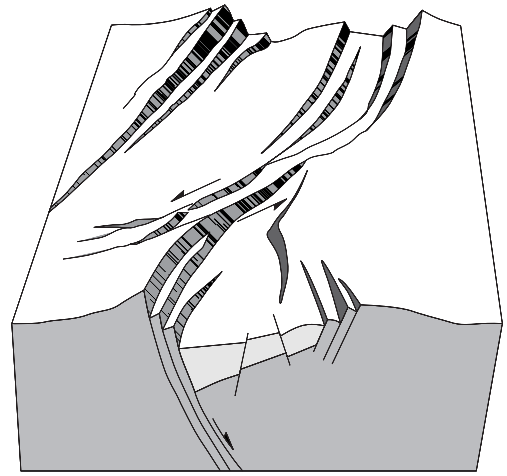
FIGURE 16.14 Map view and cross section of a "C-shaped" half graben. Note that the basin tapers toward its ends as the displacement on the fault changes from dip-slip to strike- or oblique-slip.

dropped hanging-wall blocks diminishes toward the ends of the C. As displacement increases, segments grow along strike until adjacent segments overlap and interact. When this happens, segments link end-to-end to form a continuous zone of extension. Note that segments that ultimately link to form the continuous rift are not necessarily aligned end-to-end.

Regions where two rift segments interact and connect are called **accommodation zones,** and in these zones, you'll find complex deformation involving strike-slip, dip-slip, and oblique-slip faulting (Figure 16.15a and b). The geometry of an accommodation zone depends on whether the faults that it links dip in the same or different directions, on whether the faults were initially aligned end-to-end, on how much the fault traces overlap along strike, and on whether regional stretching is perpendicular or oblique to the traces of faults. As a consequence, accommodation-zone geometry is quite variable. In some cases, accommodation zones consist entirely of strike-slip faults (called **transfer faults**) oriented at a high angle to the regional trend of the rift (Figure 16.c). If the rift is successful, *transfer* faults evolve into the *transform* faults that connect segments of mid-ocean ridges (see Chapter 19 for strike-slip fault terminology).

Let's now look more closely at the sequence of events during which rift segments link up to form a long rift system. At an early stage in rifting, segments are separated from one another along their length by unfaulted crust (Figure 16.16a). As displacement on the faults in the segments increases, the length of the segments also increases. Eventually, the individual segments interact along strike, with one dip direction dominating—at which time opposite-dipping faults cease to be so active; they are preserved as the faults of the "upper plate" (Figure 16.16b). The underlying





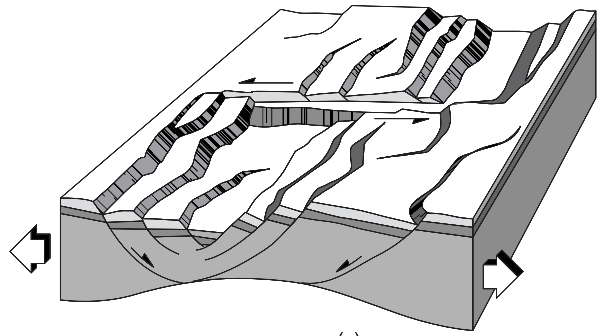


FIGURE 16.15 (a) Map-view geometry of an idealized accommodation zone, formed because the axes of two rifts are not aligned along strike. (b) Three-dimensional model of an accommodation zone. (c) Three-dimensional model of an accommodation zone that is at right angles to rift segments. This could evolve into a transform fault.

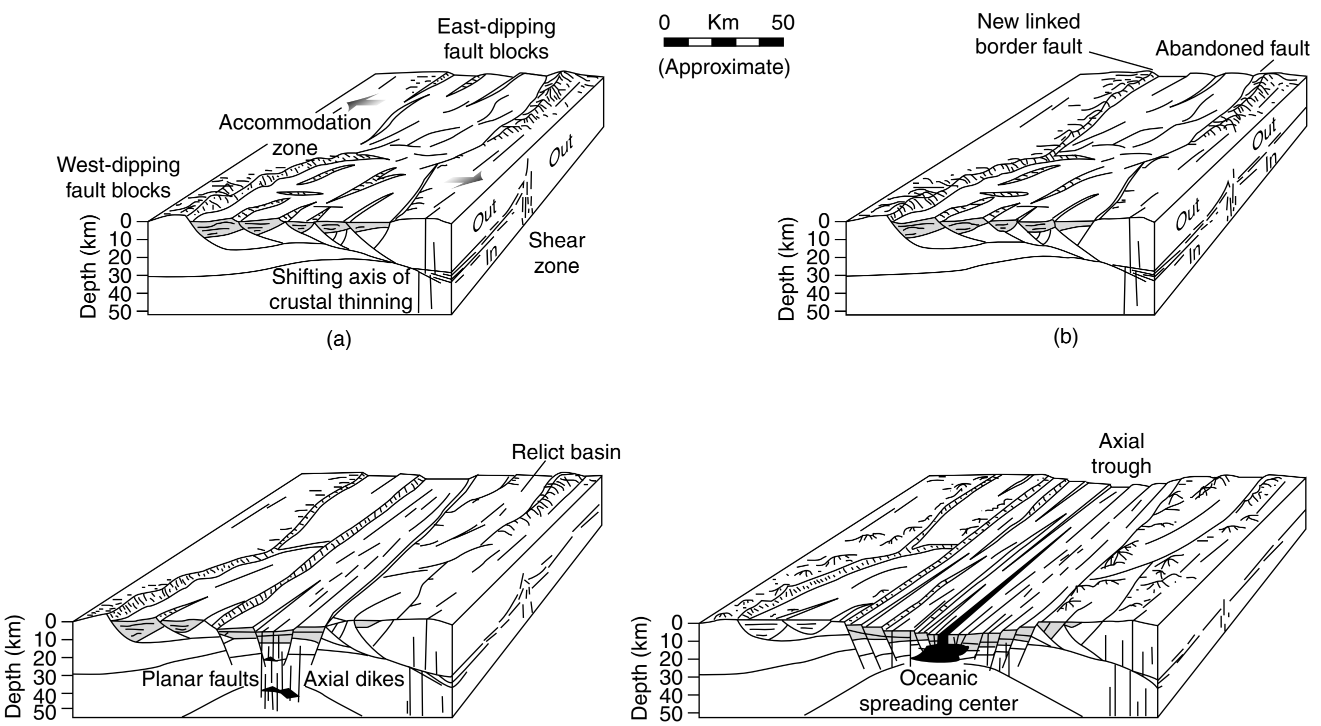


FIGURE 16.16 Rift evolution. (a) After initial rifting, two north-south-trending rift segments that face in opposite directions begin to interact. (b) A new breakaway propagates from the southern rift segment northwards, across the less active segment of the other rift. The original breakaway of the northern fault becomes inactive. (c) A swarm of dikes begins to intrude along the axis of the rift, and the rift–drift transition begins. (d) A new midocean ridge initiates, and seafloor spreading begins.

detachment also grows during this time, and propagates to greater depths in the crust.

As a rift evolves, and the lithosphere continues to stretch, the **rift–drift transition** takes place (Figure 16.16c). When this happens, discrete elongate troughs, in which heat flow is particularly high, form in places along the rift. These troughs are sites of nascent seafloor spreading. As stretching continues, the troughs widen until the continental lithosphere breaks apart, a mid-ocean ridge forms (Figure 16.16d), and a new ocean is born. Geologists can examine the evolution of rifting to seafloor spreading in the Red Sea. The northern part of the Red Sea is a mature continental rift, the central part of the sea is currently undergoing the rift–drift transition, while the southern part of the sea is a narrow ocean underlain by young oceanic crust.

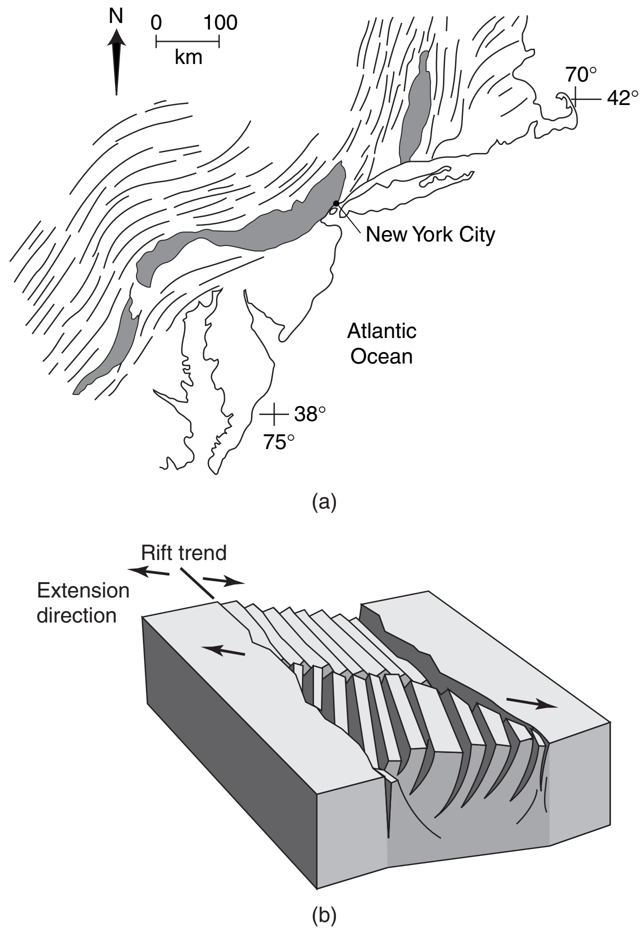


FIGURE 16.17 (a) Map of Triassic/Jurassic rift basins along the east coast of the United States. These basins formed during the early stage of rifting that separated North America from Africa, as Pangaea broke up. Note how the axes of these basins parallel the Paleozoic structural grain (defined by the trends of thrusts, folds, and fabrics) of the Appalachians. (b) Oblique rifting, as modeled in a sandbox. Here, the extension direction is oblique to the rift axis.

16.5 CONTROLS ON RIFT ORIENTATION

Two fundamental factors, preexisting crustal structure (e.g., preexisting fabrics, faults, and sutures) and the syn-rift stress field, appear to control the regional orientation of a rift. Examination of rift geometry worldwide suggests local correlation between rift-axis strike and/or accommodation-zone strike and the orientation of basement foliations, faults, and shear zones. In the Appalachian Mountains, for example, the trend of Mesozoic rift basins (Figure 16.17a) closely follows the trend of Paleozoic fabrics and faults, and in the South Atlantic Ocean, the Romanche Fracture Zone aligns with a major Precambrian shear zone in the adjacent Brazilian Shield. Your intuition probably anticipates such relationships, because planes of foliation are weak relative to other directions in a rock. Preexisting orogens are particularly favored sites for later rifting, for not only do orogens contain preexisting faults that can be reactivated, but orogens are warmer than cratonic areas of continents, and thus are weaker. Orogens are, effectively, the weakest link between continental blocks that have been connected to form a larger continent.

Though the extension direction is commonly perpendicular to the rift axis, **oblique rifting** takes place where the direction of regional stretching lies at an acute angle to the rift axis (Figure 16.17b). In such rifts, oblique-slip displacement occurs on faults. Oblique rifting may happen where fault geometry was strongly controlled by preexisting fabric and the faults happen to be at an angle to the regional stretching direction. It may also develop where stretching directions change once a rift has initiated, so that the rift axis no longer parallels a principle axis of stress.

Preexisting structures do not always control the geometry of rifting. For example, in the interior of Africa, contemporary faults of the East African Rift system obliquely cut across a preexisting rift system. At this location, it appears that the syn-rifting stress field plays a more important role in controlling rift geometry than does preexisting structure. In East Africa, stress measurements suggest that faults in the rift are perpendicular to the contemporary σ_3 direction, as predicted from the theory of brittle failure (Chapter 6).

16.6 ROCKS AND TOPOGRAPHIC FEATURES OF RIFTS

Perhaps the first geologic feature that comes to mind when you think about a rift is the system of normal faults that crack the crust and accommodate regional extensional strain. But faults are not the only geologic features of rifts. Rifts also contain distinct assemblages of sedimentary and igneous rock, and they have a distinct topography.

16.6.1 Sedimentary-Rock Assemblages in Rifts

Rifts are low areas, or "depressions," relative to the **rift margins** (the borders of the rift). Within the regional depression, movement on individual normal faults results in the development of numerous narrow, elongate **basins** (or **rift basins**), each separated from its neighbor by a **range.** These basins are grabens or half grabens, which fill with sediment, while the ranges are horsts or tilted fault blocks. Of note, the Basin and Range Province, which we noted earlier is a broad rift in the western United States, lies between two high rift margins (the Sierra Nevada Mountains on the west and the Colorado Plateau on the east); it derives its name from the many basins and ranges it contains. Geographers refer to the central portion of the Basin and Range Province as the "Great Basin;" this portion has **interior drainage,** meaning that streams from surrounding highlands flow into it, but it has no outlet.

Subsidence (sinking) of rifted crust, overall, can be relatively rapid, so that a succession of sediment several kilometers thick can accumulate in the basins of a rift during just a few million years. The depositional sequence in a rift basin reflects stages in the evolution of the basin. When a rift first forms, the floor of the rift lies above sea level (Figure 16.18a). At this stage, sediment deposited in the rift basin consists entirely of nonmarine clastic debris supplied by the erosion of rift margins or by the erosion of ranges in the rift. This sediment accumulates in overlapping alluvial fans fed by streams draining ranges or rift margins. Streams may flow down the axis of the basin, providing an environment in which fluvial deposits accumulate. Water may collect in particularly low areas of basins to form a lake; examples of such lakes include Lake Victoria in Africa, and the Great Salt Lake of Utah. If the rift lies in a desert climate, the water of the lake may evaporate to leave a salt-encrusted playa, but in more humid climates the water lasts all year. Fine-grained mud settles out in the quiet water of such lakes. Turbidites (graded beds) may also accumulate, if the lake is deep enough, for rift-related earthquakes trigger subaqueous avalanches. Thus, the sedimentary sequence deposited during the early stages of rifting typically contains coarse gravels interstratified with red sandstone and siltstone, and lacustrine sediments.

As a potentially successful rift continues to get wider, its floor eventually subsides below sea level (Figure 16.18b). At this stage, the seawater that covers the rift floor is very shallow, so evaporation rates are high. In fact, if the basin floor temporarily gets cut off from the ocean (either due to tectonic uplift at the ends of the rift or to global sea level drop) seawater in the rift basin becomes isolated from the open ocean. The trapped seawater evaporates and deposits of salt (dominantly halite ± gypsum ± anhydrite) precipitate, burying the continental clastic deposits that had formed earlier. The resulting evaporite sequence may become very thick, because global sea level may rise and fall several times, so that the rift repeatedly floods and then dries up. At this stage, the crests of ranges may be buried by sediment, so the whole rift contains only a single, wide rift basin.

With continued extension, a successful rift broadens and deepens, and evolves into a narrow ocean (Figure 16.18c). At this stage, marine strata (carbonates, sandstones, and shales) bury the evaporites. As we will see in our later discussion of passive margins, if the rift is successful, this marine sequence continues to accumulate long after rifting ceases and seafloor spreading begins, forming a very thick passive-margin sedimentary wedge. If the rift is unsuccessful, a very thick marine sequence does not cover the rift. However, as the lithosphere beneath the central part of the rift cools, thickens, and subsides, the rift margins eventually sink. As a result, a thin tapering wedge of sediment covers the rift margins, so that a vertically exaggerated profile of the rift basin and its margins resembles the head of a longhorn bull. Basins with this geometry are informally called **steerhead basins** (Figure 16.19).

16.6.2 Igneous-Rock Assemblage of Rifts

Stretching during rifting decreases the thickness of the lithosphere, and thus decreases lithostatic pressure in the asthenosphere directly beneath the rift. The asthenosphere is so hot (>1280° C) that, when such decompression occurs, partial melting takes place. Partial melting of the ultramafic rock (peridotite) comprising the asthenosphere, yields a mafic (basaltic)

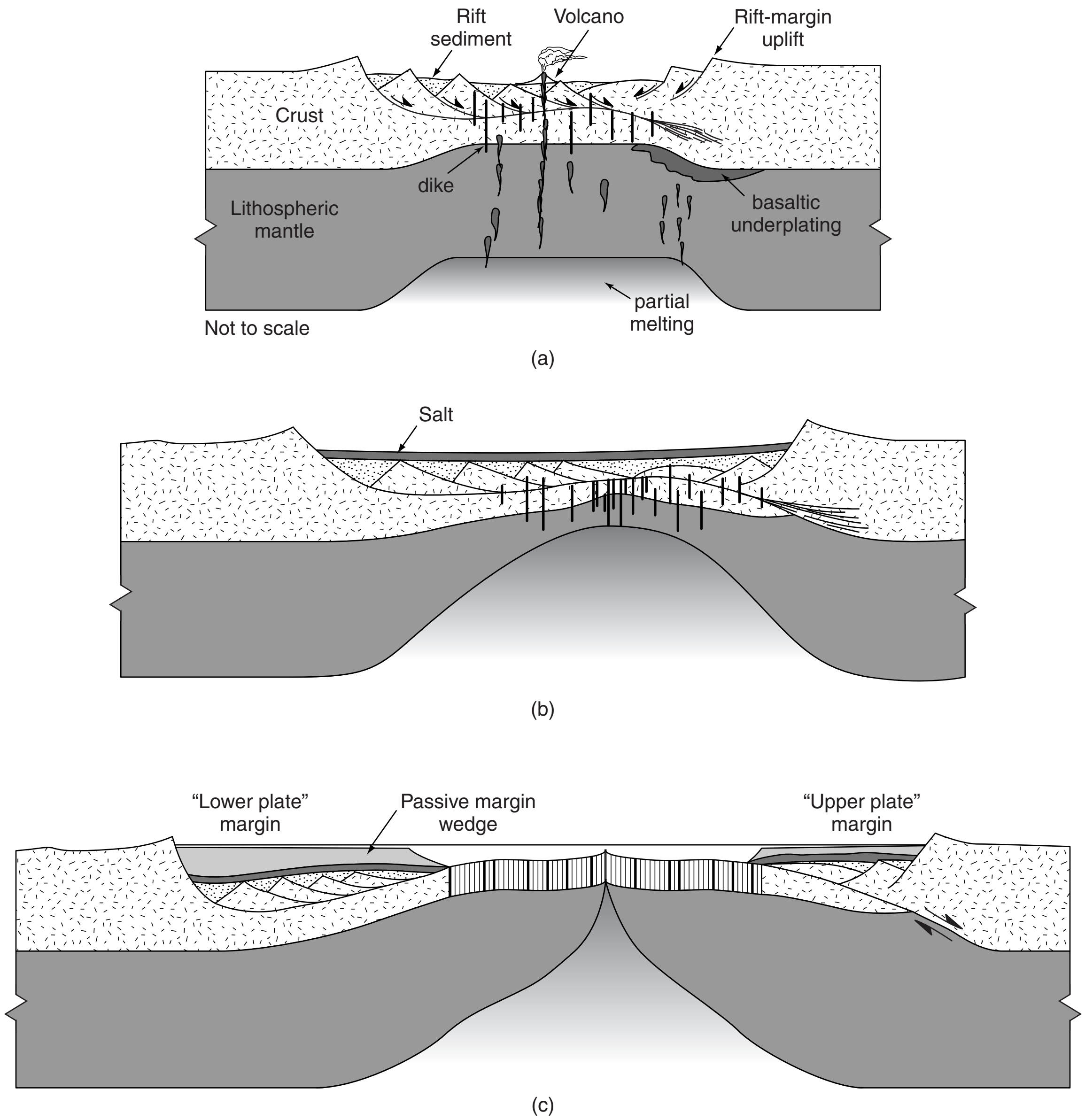


FIGURE 16.18 Evolution of a rift system into paired passive margins and sedimentary environment. (a) Rift stage with nonmarine basins; (b) rift–drift transition with evaporite deposition; (c) drift stage, with seafloor spreading occurring and passive-margin basins evolving. Marine deposition occurs in the basins.

magma.[2] The amount of partial melting at a given pressure depends on temperature, so if there is a mantle plume beneath the rift, making the asthenosphere hotter than normal, large amounts of magma will form.

[2]By **partial melting,** we mean that only the minerals with lower melting temperatures in the rock melt. Minerals with a higher melting temperature remain in the solid state. Typically, only 2–4% of the asthenosphere actually melts during the formation of rift magma. Partial melting results in a magma that is less mafic than its source, because mafic minerals (such as olivine and pyroxene) have higher melting temperatures than felsic minerals (such as quartz and K-feldspar).

Magma is less dense than the overlying lithosphere, and thus rises into the lithosphere. In some cases, a portion of the magma gets trapped and solidifies at the base of the crust, a process called **magmatic underplating.** Magmatic underplating is an important process that thickens the crust by adding mass to its base. The magma that does not solidify at the base of the crust rises higher, through cracks and along faults, and intrudes into the crust. Some magma even makes it to the Earth's surface and erupts at volcanoes. Mt. Kilimanjaro in east Africa is an example of a rift volcano.

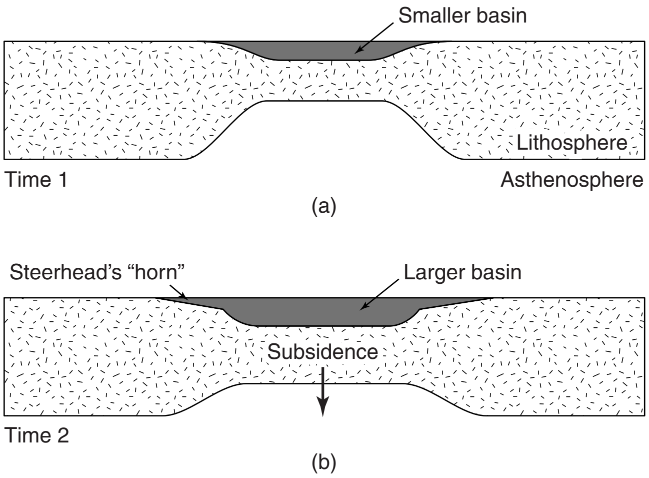


FIGURE 16.19 Thermal subsidence of a rift basin. (a) Just after rifting, the lithosphere is very thin, and the surface of the lithosphere has sunk to form a rift basin that rapidly fills with sediment. The details of the rift (faults) are not shown. (b) After time has passed, the lithosphere cools and thickens, and therefore sinks to maintain isostasy. The basin therefore gets deeper. The margins of the basin may warp down, so that sediments lap onto the unthinned lithosphere. This geometry is called a steerhead basin, because it resembles a head-on view of a bull with horns.

In rifts, which are zones of horizontal extension, the maximum principal stress (σ_1) in the crust is due to gravity and must be vertical, while the least principal stress (σ_3) trends roughly perpendicular to the rift axis (except where oblique rifting occurs). Thus, basalt that rises into the crust in rifts forms subvertical **dike swarms,** arrays of many parallel dikes, that strike parallel to the rift axis. At very shallow crustal levels, however, magma pressure may be sufficient to lift up overlying layers of rock. Here, basalt intrudes between layers and forms **sills.**[3] Basalt erupting at the ground surface in rifts has low viscosity and thus spreads laterally to create flows covering a broad area.

In a few locations, such as the Columbia Plateau in Washington and the Parana Basin in Brazil, immense quantities of basaltic lava erupted burying the landscape in immense sheets; the resulting rock is aptly named **flood basalt.** The formation of flood basalts probably occurs where a mantle plume underlies the rift, for the peridotite in the plume is significantly hotter than the peridotite comprising normal asthenosphere, and thus undergoes a greater degree of partial melting than does normal asthenosphere during decompression. Because there is more melting, more magma is produced. Further, this lava is hotter and less viscous than normal rift lava. When rifting opens conduits, the magma formed at the top of the mantle plume rushes to the surface and spews out in flows up to several hundred kilometers long.

Mafic magma is so hot (>1000° C) that when it becomes trapped in the continental crust, it conducts enough heat into the adjacent crust to cause partial melting of this crust. This melting takes place because the rock comprising the continental crust contains minerals with relatively low melting temperatures (<900° C). Partial melting of continental crust yields silicic magma that then rises to form granite plutons and rhyolite dikes at depth, or rhyolite flows and ignimbrites (sheets of welded tuff formed when hot ash flows blast out of volcanoes) at the surface. The common association of silicic and mafic volcanism in rifts is called a **bimodal volcanic suite** (Figure 16.20).

16.6.3 Active Rift Topography and Rift-Margin Uplifts

Because of their fault geometry, continental rifts typically display **basin-and-range topography** (such as occurs in the Basin and Range Province of the western United States). As noted earlier, some of the **ranges** (high ridges) are horsts, but most are the unburied tips of **tilted fault blocks** (i.e., hanging-wall blocks). Similarly, some of the **basins** (low, sediment-filled depressions) are grabens, but most are **half grabens,** depressions bounded on one side by a normal fault and on the other side by the surface of a tilted fault block.

Figure 16.21 shows profiles across two active rifts, the Basin and Range Province in Utah and Nevada, and the Red Sea rift in Egypt. Note that in both examples, the axis of the rift is a low area relative to the rift margin. As we mentioned earlier, the highlands bordering the rift are known as **rift-margin uplifts,** and they can reach an elevation of 2 km above the interior of the rift. In the Basin and Range, an early- to intermediate-stage rift, the interior of the basin is itself substantially above sea level, whereas in the Red Sea, which is a late-stage rift, the interior of the basin is below sea level. Clearly, the topography of rifts evolves through time in concert with the development of strain in the rift. During most of its evolution, a rift is actually relatively high. It is only during its old age, long after it has ceased being active, or when it enters the rift–drift transition, that rift floors subside to sea level or deeper.

Why are many rifts high during the early stages of their evolution? One cause may be preexisting elevation

[3]A large sill, called the Palisades Sill, intruded the redbeds of the Mesozoic Newark rift basin in New Jersey. Rock of this sill forms the high cliffs that border the west bank of the Hudson River, opposite New York City.

FIGURE 16.20. Photograph of a basalt flow on top of a rhyolitic ignimbrite, in the Basin and Range Rift of the western United States.

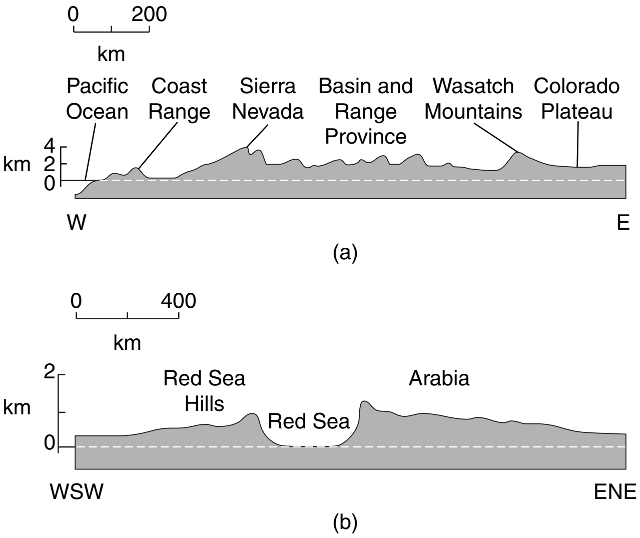


FIGURE 16.21 Topographic profiles across two rifts. (a) The Basin and Range Rift of western United States; (b) the Red Sea Rift of northeastern Africa.

of the region undergoing extension. Rifts forming in regions that previously were orogenic belts are high to start with. For example, rifting is currently occurring in Tibet, a region that is several kilometers above sea level (Chapter 21). But even if rifting begins in a region at relatively low elevation, the process of rifting can cause uplift, because rifting heats the lithosphere and makes it more buoyant. Let's see how this works.

Heating in rifts occurs for two reasons. First, the magma that rises through the lithosphere during rifting carries heat, and this heat radiates into the surrounding rock. Second, thinning of the lithosphere that accompanies rifting causes the 1280° C isotherm defining the lithosphere-asthenosphere boundary to move closer to the Earth's surface. Bringing the isotherm closer to the surface is much like placing a hot plate beneath the lithosphere, so the lithosphere at shallower depths gets warmer.[4] Heating the lithosphere causes it to become less

[4]Heat-flow studies demonstrate that heating of the lithosphere does occur in rifts. For example, the average heat flow for the Canadian shield is ~1 HFU whereas the average heat flow in the East African Rift is about 2.5 HFU. (The abbreviation "HFU" means "Heat Flow Unit." It is a measure of the heat passing through an area in a given time. 1 HFU = $40 mW/m^{-2}$ where mW = megawatt).

dense because rock expands when heated; so, to maintain regional isostatic compensation, the surface of the lithosphere in the rift rises when stretching takes place.

As rifting progresses, stretching leads to normal faulting and graben and half-graben formation. This deformation causes the interior of the rift to sink, relative to its margins. In addition, normal faulting decreases the vertical load (weight) pressing down on the rift margins, so they rebound elastically and move up even farther, much like a trampoline surface moves up when a gymnast steps off of it. The combination of subsidence within the rift and elastic rebound of the rift margins can lead to a situation in which rift margins evolve into mountain ranges that tower over the internal portion of the rift.

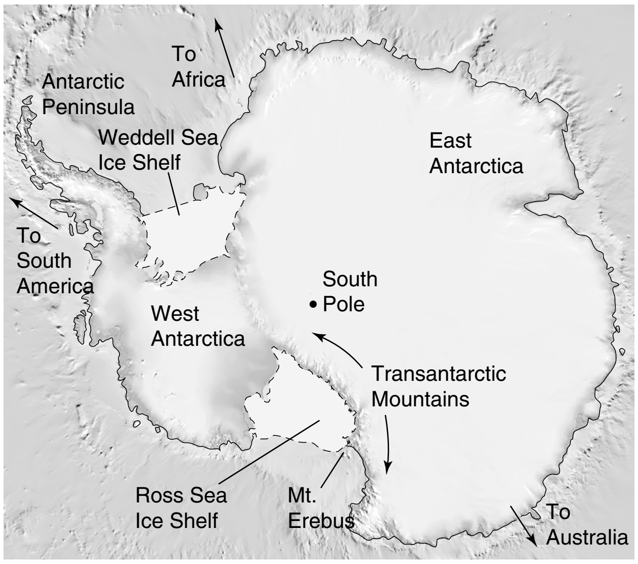
After rifting ceases, rifts eventually subside. This subsidence happens because, when rifting ceases (i.e., lithospheric stretching ceases), the stretched lithosphere begins to cool. As a consequence, the 1280° C isotherm migrates down to lower depth in the mantle that, by definition, means that the lithosphere thickens. As the lithosphere cools and becomes thicker, its surface sinks in order to maintain isostatic equilibrium. In some cases, the rift margins also subside.

Significantly, not all rift margins sink after rifting ceases—some remain elevated long afterwards. Examples of **long-lived rift-margin uplifts** include the Serra do Mar along the southeast coast of Brazil (Figure 16.22a), the Blue Mountains of Australia, and the Transantarctic Mountains of Antarctica (Figure 16.22b). Note that the elevation of such uplifts is not a relict of collisional orogeny. For example, the rocks that comprise the famous high peaks of Rio de Janeiro formed during a collisional orogeny in the Proterozoic, but the present uplift developed during the Late Mesozoic and Cenozoic initially in association with opening of the South Atlantic. Similarly, metamorphism of rocks in the Transantarctic Mountains took place during a collisional orogeny in Late Proterozoic to Early Paleozoic times, but this orogen was beveled flat and was covered by Jurassic strata and volcanics. The uplift we see today developed in Late Mesozoic and Cenozoic times, in association with the rifting that formed the Ross Sea. Jurassic strata of the Transantarctic Mountains are still nearly flatlying, even though they now lie at elevations of over 4 km.

The cause of long-lived rift-margin uplifts remains a matter of debate. Two processes may contribute to their development (Figure 16.23):

- First, the uplifts may have gone up because of basaltic underplating. This process thickens the less-dense crust relative to denser (ultramafic) mantle lithosphere and leads to isostatic uplift, which may be long-lived. Thickening the crust, relative to the lithospheric mantle, is like adding a float to a block of wood floating in a bathtub.
- Second, the uplifts may have gone up because rifting decreased the strength of the lithosphere. When this happened, the presence of isostatically uncompensated buoyant loads (e.g., large granitic plutons from an earlier episode of intrusion) could have caused the lithosphere containing the loads to rise to attain isostatic equilibrium. Prior to rifting, these loads had been held down by the strength of the prerift lithosphere. As an analogy, picture a balloon trapped under a floating piece of plywood. The strength of the plywood can hold the balloon completely under water. But if the wood is replaced by

(a)



(b)

FIGURE 16.22 (a) Photo of the Serra do Mar, along the east coast of Brazil. Here, mountains exposing Precambrian granite and gneiss rise directly out of the ocean, along South America's passive margin. (b) Shaded relief map of Antarctica, showing the Transantarctic Mountains along the edge of the East Antarctic craton. The Ross Sea and Wedell Sea were formed by Mesozoic rifting.

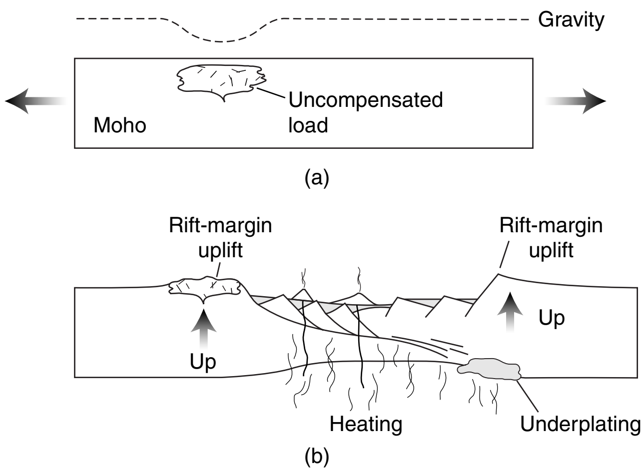


FIGURE 16.23 Hypotheses to explain long-lived rift-margin uplifts. (a) The crust, before rifting, containing an uncompensated buoyant mass (e.g., a large pluton); (b) during rifting, the strength of the lithosphere decreases, so the buoyant crust rises to achieve isostatic equilibrium. Underplating of basalt thickens the crust and can also cause uplift.

a thin sheet of rubber, a much weaker material, the buoyancy of the balloon can cause the surface of the rubber to rise and stay high.

16.7 TECTONICS OF MIDOCEAN RIDGES

In December, 1872, the H.M.S. *Challenger* set sail from England with a staff of researchers to begin the world's first oceanographic research voyage. Over the next four years, the ship traversed 127,500 km of the Atlantic and Pacific Oceans, during which, among other tasks, it made depth soundings at 492 different locations. Each measurement took many hours, for it required sailors to let out enough cable for a lead weight to reach the seafloor. Nevertheless, the crew obtained enough measurements to suggest that elongate submarine mountain ranges, known as midocean ridges, rose from the abyssal plains of the oceans. Sonar surveys in the twentieth century provided much more data and, by the 1950s, had provided our modern image of midocean ridges. Midocean ridges occur in all ocean basins, rising about 2 km above the floor of adjacent abyssal plains. Stretched out end-to-end, the midocean ridges of Earth today would make a chain 40,000 km long (Figure 16.24). Ridges are not continuous, but rather consist of segments, ranging from ten to several hundred kilometers in length, linked at their ends by transform faults. (We'll discuss these transform faults in Chapter 19). The transform faults are roughly perpendicular to the ridge axis.

In 1960, Harry Hess suggested that midocean ridges mark divergent plate boundaries at which new oceanic lithosphere forms as seafloor spreading occurs. The new lithosphere moves away from the ridge axis, and the ocean basin becomes wider. At the ridge axis, seafloor spreading involves extensional deformation, producing structural features that are similar to those found in continental rifts.

Oceanic lithosphere consists of two components (Chapter 14). The upper part of the lithosphere, the oceanic crust, ranges from 6 to 10 km in thickness and consists of five distinct layers named, from bottom to top: the cumulate layer, the massive-gabbro layer, the sheeted-dike layer, the pillow-basalt layer, and the pelagic-sediment layer. The lower part of the oceanic lithosphere consists of lithospheric mantle; this layer varies in thickness from 0 km, beneath the ridge, to over 90 km, along the margins of a large ocean. In Chapter 14, we noted that oceanic crust forms by magmatic processes at the midocean ridge. Here, we delve into these processes in greater detail, and add information about bathymetry and the structures of midocean ridges.

Figure 16.25 provides a simplified cross section of a midocean ridge. Hot asthenosphere rises beneath the ridge axis. The resulting decrease in pressure (decompression) in the asthenosphere at this location causes partial melting, producing basaltic magma, which rises to fill a large magma chamber in the crust beneath the ridge axis. As time passes, the magma cools, and crystals of olivine and pyroxene begin to grow within it. These crystals are denser than the magma and sink to the bottom of the magma chamber to form a layer of cumulate (**cumulate,** by definition, is a rock formed from dense minerals that sank to the bottom of a magma chamber). Meanwhile, crystallization at the sides of the magma chamber produce gabbro, a coarse-grained mafic igneous rock. As seafloor spreading progresses, this gabbro moves away from the axis of the chamber, creating more room for magma. Spreading also cracks the rock overlying the magma chamber, creating conduits for magma to rise still further. Some of this magma freezes in the cracks to form the dikes of the sheeted dike layer, while some reaches the surface of the Earth and extrudes on the seafloor to form the pillow basalt layer. Continued seafloor spreading moves new-formed crust away from the ridge axis. Eventually, sediment (plankton shells and clay) snow-

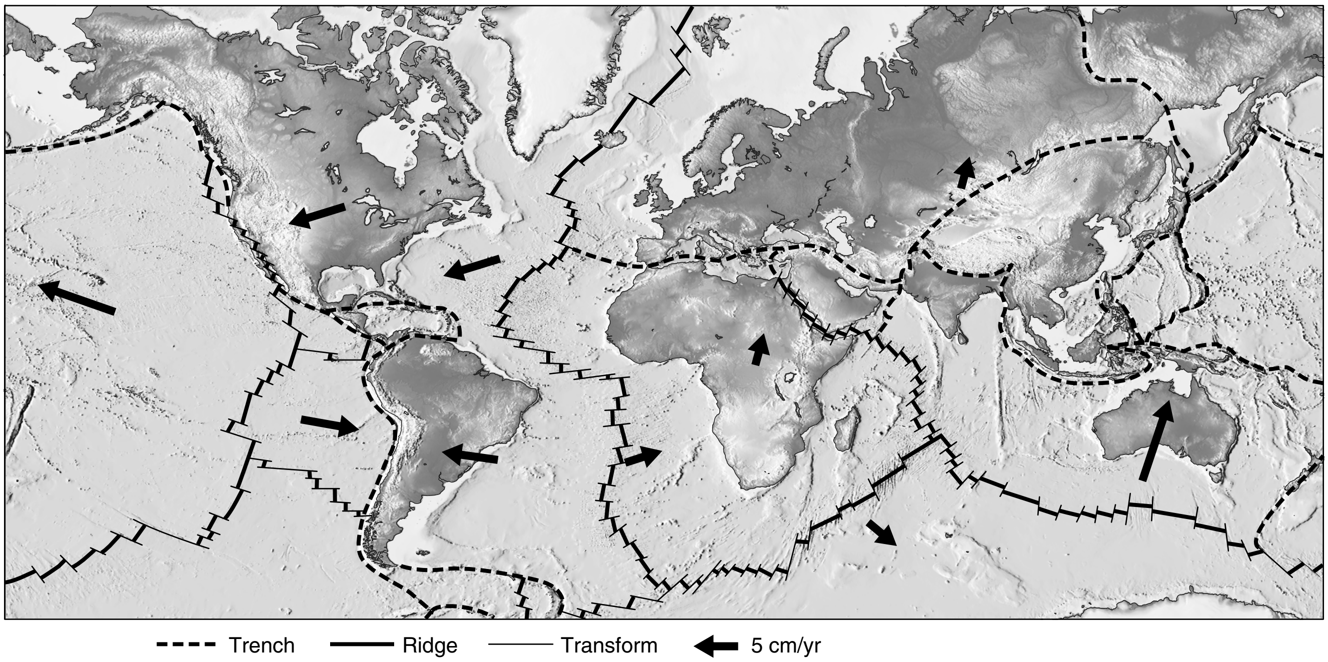


FIGURE 16.24 The midocean ridges of Earth. Note that ridges, in map view, consist of short segments, linked by transform faults. The lengths of arrows are proportional to the velocity.

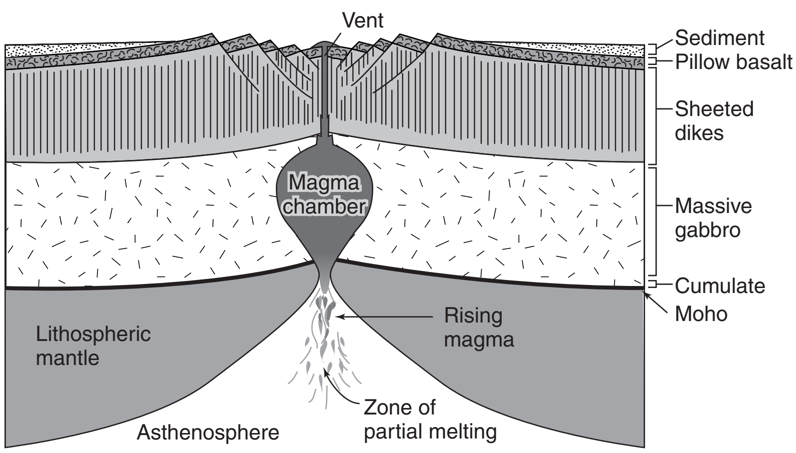
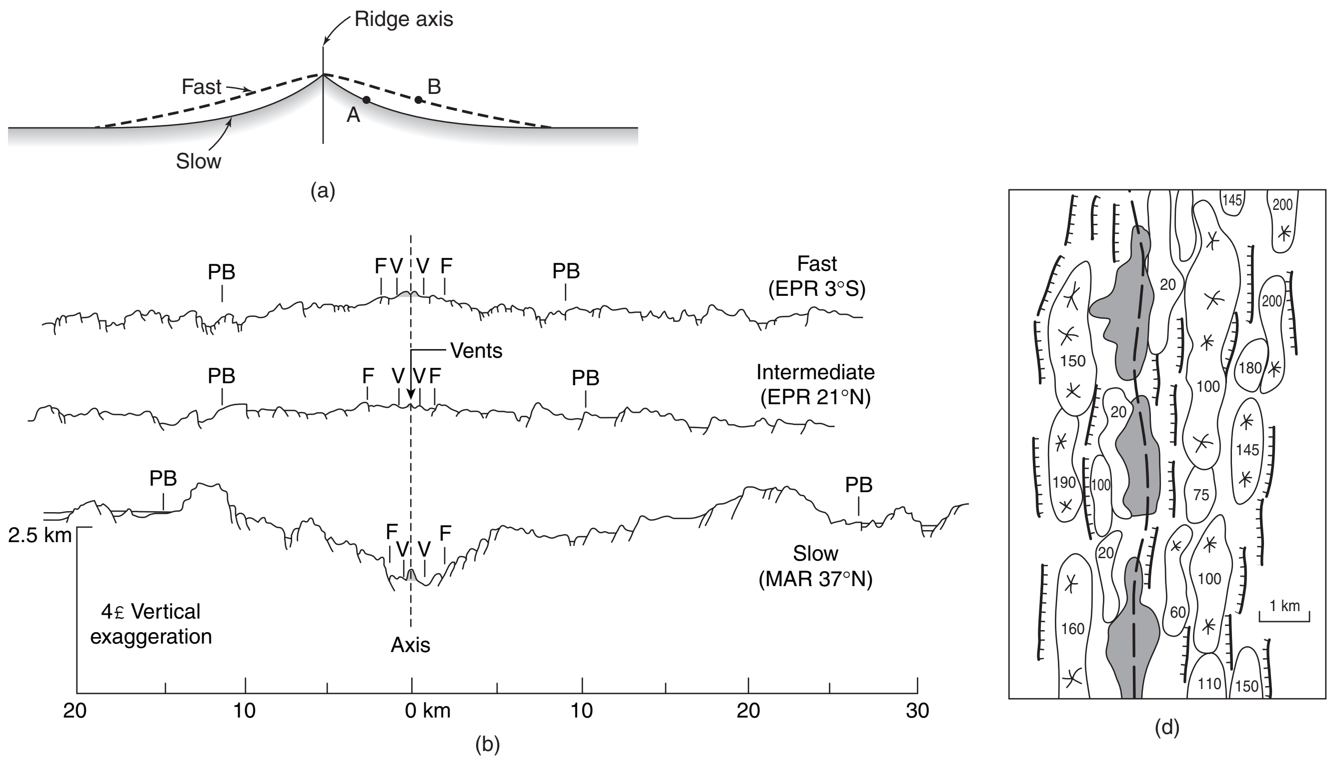


FIGURE 16.25 Schematic cross section (not to scale) through the oceanic crust and upper mantle at a fast-spreading ridge. This illustrates a model for the formation of the distinct layers of oceanic crust.

ing down from the overlying sea buries the pillow basalt.

The bathymetry and structure that occur at a midocean ridge reflect the rate of seafloor spreading (Figure 16.26a). Geoscientists distinguish between **slow ridges** (e.g., the Mid-Atlantic Ridge), where plates move apart at rates of <4 cm/y, and **fast ridges** (e.g., the East Pacific Rise), where plates move apart at rates of > 8 cm/y. (There are also intermediate-spreading ridges, with spreading rates of between 4 and 8 cm/y). Slow ridges are relatively narrow (hundreds of kilometers wide), and have deep (>500 m) axial troughs bordered by steplike escarpments (Figure 16.26b and c). In contrast, fast ridges are broad (up to 1500 km wide), relatively smooth swells, with no axial trough.

Why are fast ridges wider than slow ridges? The answer comes from a consideration of isostasy and how it affects the level at which oceanic lithosphere floats in the sea (Figure 16.26a; see Chapter 14). Beneath a midocean ridge, the lithospheric mantle is



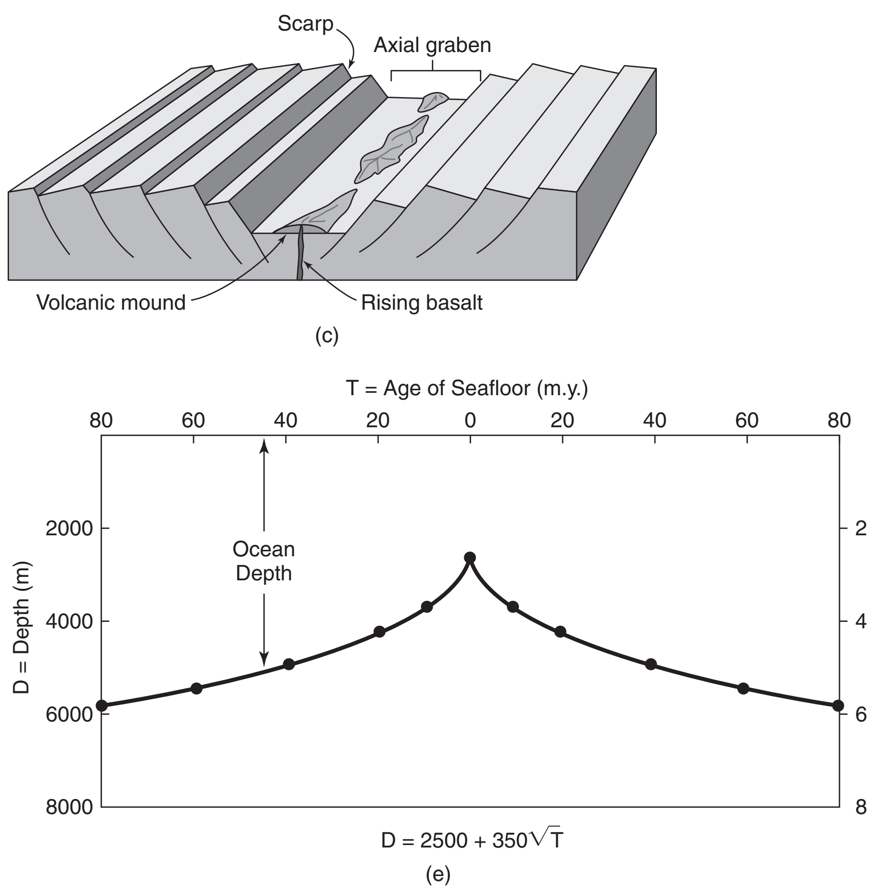


FIGURE 16.26 Morphology of a midocean ridge axis. (a) Simplified profiles of a fast ridge versus a slow ridge. Notice that fast ridges are wider. That's because the depth to which the seafloor sinks depends on its age. The slow plate at point *A* is the same age as the fast plate at point *B,* but both are at the same depth. (b) Profiles contrasting fast and slow ridge axes. The top profile is across the East Pacific Rise (3° S latitude), while the lower profile is across the Mid-Atlantic Ridge (37° N latitude). Note that an axial trough only occurs at slow ridges, not at fast ridges. "V" marks the position of volcanic vents. (c) Three-dimensional block diagram emphasizing the axial graben of a ridge. Mounds of pillow basalt build up over individual vents. Fault scarps border the graben. (d) A detailed map of the axial graben of the Mid-Atlantic Ridge, showing dated mounds of basalt. Note that the entire ridge is not active at any given time. The barbed lines are faults, and the stars are vents. The heavy dashed line is the plate boundary. Ages are in hundreds of thousands of years. (e) The age versus depth curve for seafloor.

very thin (in fact, at the axis of the ridge, hot asthenosphere lies just below the crust, so there is no lithospheric mantle). As the seafloor drifts away from the ridge axis, cooling takes place, and the 1280° C isotherm sinks, making the lithospheric mantle thicker. As a consequence, the ratio of less-dense crust to denser lithospheric mantle decreases as the lithosphere gets older. This makes the base of the older lithosphere sink deeper into the asthenosphere, relative to the base of the younger lithosphere, to maintain isostatic compensation.[5] Thus, the depth of the seafloor increases with age, as shown by Figure 16.26e. With this concept in mind, we see that fast ridges are wider than slow ridges because young seafloor (with a thin lithospheric mantle) exists at a greater distance from the axis of a fast ridge than it does from the axis of a slow ridge. Put another way, because seafloor spreading takes place more slowly at a slow ridge, lithosphere at a given distance from the ridge axis, has aged more, and thus has subsided more, than lithosphere at the same distance from a fast-ridge axis.

The morphology at the axis of a midocean ridge reflects the faulting that accompanies horizontal extension. Specifically, the stretching of newly formed crust breaks the crust and forms normal faults, which generally dip toward the ridge axis. Studies using research submersibles, like the *Alvin,* demonstrate that fault escarpments expose fault breccias whose matrix includes the mineral serpentine; this serpentine forms when the olivine of ocean-crust basalt reacts with hydrothermal fluids circulating in the crust of the ridge.[6] Submarine debris flows, formed from broken-up rock that breaks free and falls down fault scarps, accumulate at the base of the scarps. Submarine mapping shows that, on a timescale of thousands of years, the entire length of a ridge is not active simultaneously, for geologists can identify different mounds of lava formed at different times (Figure 16.26d).

The contrast between the morphology of the ridge-axis zone at fast ridges and that of the ridge-axis zone at slow ridges may reflect, in part, a balance between the amount of magma rising at the axis and the rate at which stretching of the crust occurs. At slow ridges, relatively little magma forms, and the magma chamber beneath the ridge axis (see Chapter 15) periodically freezes solid. As seafloor spreading at slow ridges takes place, the brittle crust stretches and breaks, and a graben develops over the axis. This graben comprises the axial trough. At fast ridges, in contrast, a nearly steady-state magma chamber may exist in places beneath the ridge axis, meaning the supply of new magma added to the chamber from below roughly balances the volume of solid gabbro formed along the margins of the chamber in a given time. Thus, the magma chamber is always inflated with magma. Magma pressure in this chamber may keep the crust high at the axis, preventing formation of an axial graben. Also, the voluminous extrusion of basalt may fill what would have been the central graben. At a distance from the axis of any midocean ridge, fault scarps disappear, at least in part because pelagic sediment progressively buries the structures once they become inactive. Some researchers have suggested, however, that slip on faults reverses away from the ridge axis, so that hanging-wall blocks rise back up to higher elevation. The nature of this movement, if it occurs, is not well understood.

[5]As an analogy, think of a tanker ship being loaded with crude oil. Before loading, the tanker contains air and floats high, with its deck well above sea level. As the tanker fills, it sinks into the water, and the deck becomes lower.

[6]Hydrothermal fluids cause extensive alteration of ocean-crust basalt. These fluids form when seawater sinks into the crust via cracks, is warmed by heat brought into the crust by magma, and then rises back to the seafloor. Magmatic heat effectively drives convection of water through the crust. The vents at which hot seawater is released back into the ocean are called **black smokers,** because the hot water contains dissolved sulfide minerals that precipitate when the hot water mixes with cold seawater. Chimneys of sulfide minerals accumulate around the vents.

16.8 PASSIVE MARGINS

As noted earlier, geologists distinguish between two basic types of continental margins. **Active margins** are continental margins that coincide with either transform or convergent plate boundaries, and thus are seismically active, while **passive margins** are not plate boundaries and thus are not seismically active. Passive margins develop over the edge of the inactive rift relict that remains after the rift–drift transition has taken place and a new midocean ridge has formed. Examples of present-day passive margins include the eastern and Gulf Coast margins of North America (Figure 16.27), the eastern margin of South America, both the eastern and the western margins of Africa, the western margin of Europe, the western, southern, and eastern margins of Australia, and almost all margins of Antarctica.

Once the drift phase begins, the rift relict (i.e., thinned continental lithosphere) underlying a passive margin gradually subsides. This subsidence happens because, when stretching ceases, thinned lithosphere, which had been heated during rifting, cools. As cooling takes place, the lithospheric mantle thickens, and thus the lithosphere as a whole sinks to maintain isostatic

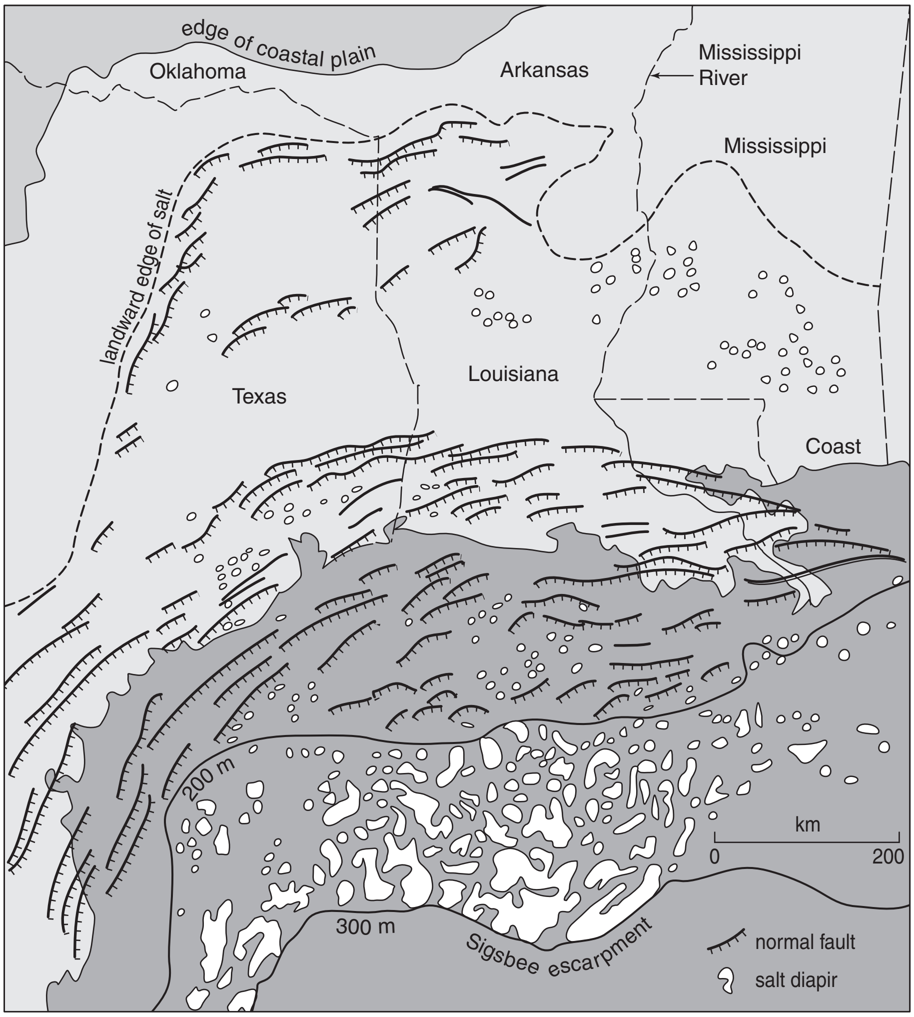


FIGURE 16.27 Simplified map of the Gulf Coast region, off the coast of Louisiana and Texas (USA), illustrating the normal faults that dip down to the Gulf of Mexico, and the abundant salt diapirs. Part of the region lies within the Coastal Plain Province, a region submerged by the sea in the Cretaceous and early Tertiary. The edge of the salt (zero isopach) is shown. Offshore, we show the 200-m and 3000-m bathymetric contours. Thrusts at the toe of the passive-margin wedge emerge at the Sigsbee Escarpment.

compensation. The process of sinking to maintain isostatic equilibrium during cooling is called **thermal subsidence.**

Effectively, thermal subsidence of a passive margin creates a space that fills with sediment eroded off of the adjacent continent and carried into the sea by rivers. This "space" is a **passive-margin basin,** and the sediment pile filling the basin is a **passive-margin sedimentary wedge.** The surface of the landward portion of a passive-margin sedimentary wedge comprises the **continental shelf,** a region underlain by an accumulation of shallow-water marine strata. The **continental slope** and **continental rise** form the transition between the shelf and the abyssal plain, and are underlain by strata deposited in deeper water.[7] Passive-margin basins are typically segmented along strike into discrete sub-basins separated by basement highs; sediment is thinner over the basement highs. Individual sub-basins probably correspond to discrete rift segments, while the highs correspond to accommodation zones. Note that loading by sediment causes the floor of the basin to sink even more than it would if no sediment was present. Specifi-

[7]In pre–plate tectonic literature, a passive-margin sedimentary wedge was referred to as a "geosyncline." Geosynclines, in turn, were subdivided into a "miogeosyncline," which consisting of shallower-water facies of the continental shelf, and a "eugeosyncline," consisting of deep-water facies of the slope and rise.

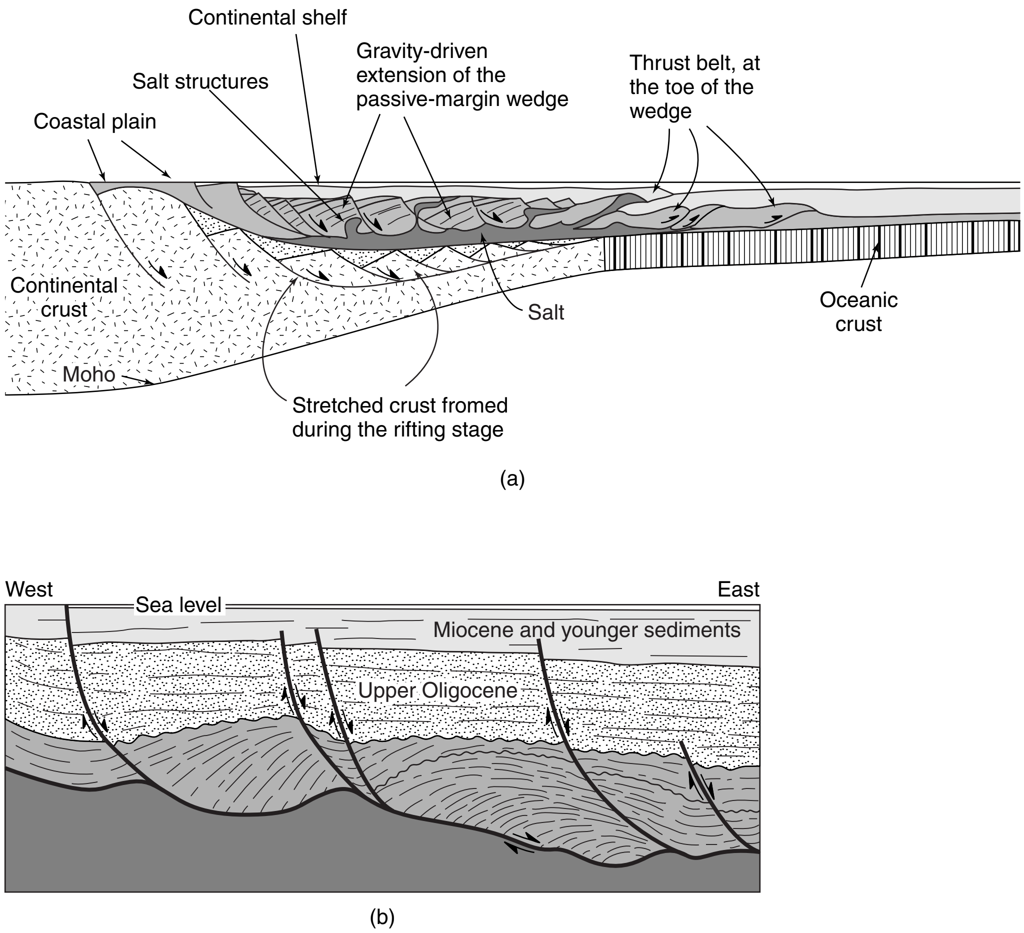


FIGURE 16.28 Cross section of a passive-margin wedge. This cross section resembles that of the Gulf Coast (USA), but has been simplified, and is not an exact representation. Note that there are two levels of extensional faulting. The lower level formed during the original rifting of the continental crust that led to the formation of this passive margin. The upper level developed within the passive-margin wedge sediments and is due to seaward gravity-driven slip of the wedge. The detachment for this upper level probably lies in the salt near the base of the section. The salt has risen to form diapirs, especially in the footwalls of normal faults, due to buoyancy and to differential vertical loads. A thrust belt has developed at the toe of the wedge. The inset shows a detail of bedding in Cenozoic strata. Note that the faults are listric growth faults and that strata in the hanging-wall blocks thicken toward the fault.

cally, the weight of 1 km of sediment causes the floor of the basin to sink by about 0.33 km.

Stretching and thinning of continental lithosphere ceases after the rift–drift transition. Normal faults, however, do develop in a passive-margin sedimentary wedge, long after the drift stage has begun (Figure 16.28). This faulting, which involves only the strata deposited above the older, rifted continental crust, is a consequence solely of gravitational force. It can occur because the seaward face of the passive-margin wedge is a sloping surface separating denser sediment below from less-dense water above. Thus, strata of the passive margin are able to slowly slip seaward, much like a slump in weak sediment of a hill slope. As it slips, the sedimentary wedge stretches; the stretching is accommodated by normal faulting. Typically, the evaporite horizon near the base of the sedimentary section serves as the sliding surface (i.e., basal detachment) for this normal faulting (Figure 16.28).

Seaward sliding of passive-margin wedge sediment leads to the development of a system of half grabens, grabens, and rollover folds in the wedge. Many of these faults can be called **growth faults,** in the sense that they are growing (displacement is taking place and the fault cuts updip) as sedimentation continues. The geometry of the normal-fault system that forms in a passive-margin wedge may be modified substantially by salt tectonics (see Chapter 2). Salt diapirs, salt pillows, and salt walls rise in response to the unloading that accompanies movement on normal faults. In some cases, rising salt flows spread laterally to form an **allochthonous** salt sheet at a higher stratigraphic level, and this sheet acts as a basal detachment for a new generation of normal faults, affecting only strata above the allochthonous salt sheet. Because of the complex interplay between gravity-driven faulting, deposition, and salt movement, the structural architecture of passive-margin basins tends to be extremely complex. Oil exploration companies have invested vast sums in trying to understand this architecture because passive-margin basins contain large oil reserves.

Just as the body of a slump on a hill slope rises up and thrusts over the land surface at the downhill toe of the slump, the seaward-moving mass of sediment in a passive-margin wedge rides up on a system of submarine thrusts at the base of the continental slope. Thus, a small fold-thrust belt develops along the seaward toe of the passive-margin wedge, while normal faults continue to develop in the landward part of the wedge (Figure 16.28).

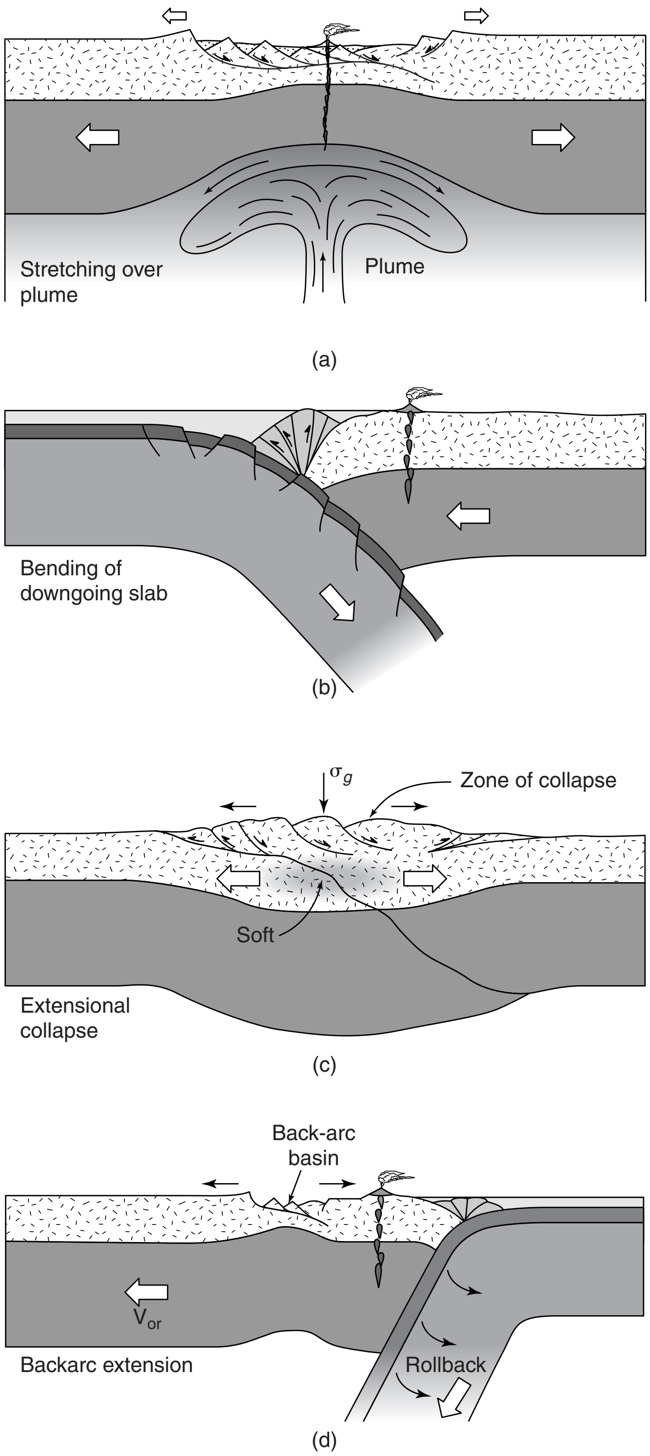
Considering that passive-margin basins evolve from continental rifts, and that rifts tend to be asymmetric, we find two different endmember classes of passive margins. These classes depend on whether the margin evolved from the upper-plate side of the rift or the lower-plate side of the rift (see Figure 16.10). In **upper-plate margins,** which evolve from the upper-plate portion of a rift, the lithosphere has not been stretched substantially, and there is relatively little subsidence; upper-plate margins typically develop small passive-margin basins and correspond to narrow continental shelves. In **lower-plate margins,** which evolve from the lower-plate portion of a rift, the lithosphere has been stretched substantially, so extreme subsidence takes place, leading to wide passive-margin basins that may contain a succession of sediment up to 20 km thick; these correspond to wide continental shelves. Figure 16.18c schematically illustrates the basic differences between these two classes. Researchers have been able to document a few examples in which a lower-plate margin lies on the other side of an ocean from an upper-plate margin. Notably, both upper-plate and lower-plate margins may occur along the same continental margin. In such cases they link along a strike at accommodation zones.

We conclude our brief description of passive margins by noting that not all of these form by stretching perpendicular to the margin. Some margins have evolved from regions where transtensional faulting took place. For example, the northeast margin of South America began as a transtensional system as South America began to pull away from Africa by rifting of the South Atlantic. Eventually, the extensional component on this system was sufficient to break South America and Africa apart. As a consequence of the transtensional phrase of its history, the northeast margin of South America has a fairly narrow continental shelf, locally underlain by “flower structure” (see Chapter 19).

16.9 CAUSES OF RIFTING

Up to this point, we’ve focused on the architecture and evolution of rifts. Now we turn our attention to the question of why rifting occurs in the first place. Given that convection occurs in the asthenosphere, is it correct to picture all rifts as places where the continent lies above the upwelling part of large mantle convective cell? No—because it is impossible to devise a geometry of simple convective cells that is compatible with the present-day geometry of rifts. (For example, if upwelling takes place at the South Atlantic Ridge and at the Indian Ocean Ridge, how can there also be upwelling along the East African Rift?) Shear stresses applied to the base of plates by convective flow of the asthenosphere may contribute to rifting in some places, but several additional processes appear to play a role as well. Let’s consider various reasons for rifting (Figure 16.29):

- In places where mantle plumes rise, forming hot spots at the base of the lithosphere, the lithosphere heats up and rises. Thus, it must undergo stretching. As a consequence of this stretching, normal faults form in the upper part of the lithosphere. Rifts that form in response to the rise of hot mantle are called **thermally activated rifts.** Three rift arms, each at an angle of 120° to its neighbor, may nucleate above a single hot spot. In classic models of thermally activated rifting, a long continuous rift develops when the arms radiating from one hot spot link with the arms radiating from a neighboring hot spot. In such situations, one of the arms of the original three-armed rift shuts off. This unsuccessful rift, called a **failed arm,** becomes an aulacogen. Whether thermally activated rifts can be successful, and whether large rifts really form by linkage of rift arms radiating from hot spots, remains unclear.
- Rifting may be caused by changing the radius of curvature of a plate. Such **flexure-related rifting** occurs where the lithosphere bends just prior to descending beneath a collisional boundary. As a consequence of the lithosphere’s bending, a series of normal faults that parallel the boundary develops in the descending plate. Flexure-related rifting may also occur because the Earth is not a perfect sphere.



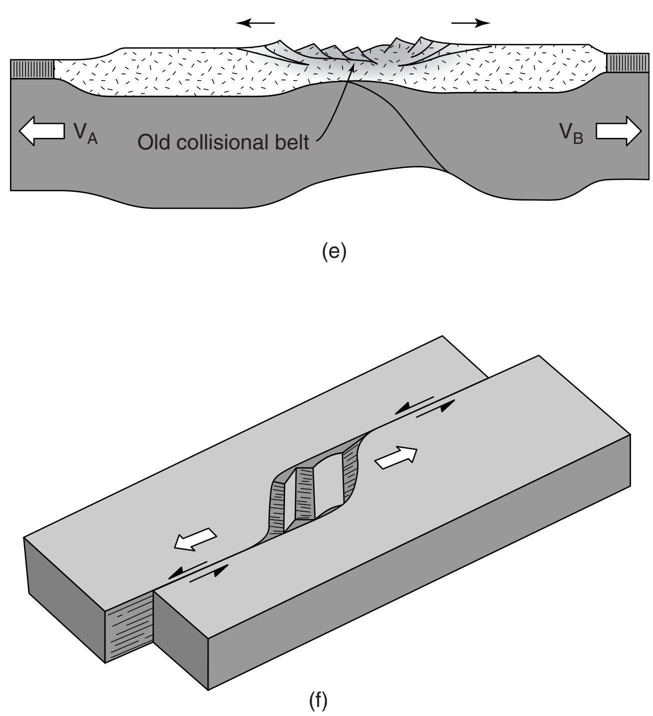


FIGURE 16.29 Causes of rifting. (a) Rifting above a thermal plume. The rising plume uplifts the lithosphere, domes up the crust, and causes it to stretch. (b) Outer-arc extension of a bending slab at a subduction zone; as the downgoing (subducting) plate bends to descend into the mantle, the top surface of the plate stretches to form a series of grabens. (c) Gravitationally driven extensional collapse of thickened crust at an orogen; the crust becomes soft at depth, and gravitational potential energy causes it to spread laterally, even if convergence is continuing. The upper crust breaks up by normal faulting. (d) Backarc extension associated with convergence. If the overriding plate is not moving in the same direction as the rollback of the subducting plate, the overriding plate breaks up at, or just behind the volcanic arc. (e) If plates are moving apart, perhaps dragged by slab pull, an intervening continent may be stretched and broken apart, like a loaf of bread that you break apart in your hands. Rifting localizes at a weak old orogen; (f) a pull-apart basin formed at a releasing bend along a strike-slip fault.

Specifically, the radius of curvature of an elastic plate changes as the plate moves from one latitude to another. The issue of whether such **membrane stresses** are sufficiently large to break plates also remains controversial.

- As discussed in Chapter 14, plates feel ridge-push and slab-pull forces, and these contribute to driving plates, independent of asthenosphere flow. Some rifts may form when the two ends of a continent are pulled in opposite directions by such plate-driving forces. If the continent contains a weak zone (e.g., a young orogen), these forces may be sufficient to pull the continent apart.
- Some rifts develop in regions of thickened and elevated crust in convergent or collisional orogens, even as contractional deformation continues. This observation implies that zones of extreme crustal thickening become zones of extension. Researchers

suggest that this relationship makes sense because the quartz-rich rocks of the continental crust are not very strong, especially where heated in an orogenic belt. So, when continental crust thickens and rises relative to its surroundings during a collisional or convergent orogeny, gravitational potential energy causes the thickened and elevated zone to spread laterally under its own weight. This process is known as **extensional collapse** (or **gravitational collapse,** or **orogenic collapse**). To visualize extensional collapse, take a block of soft cheese that has a hard rind. When you put the block in the sun, the cheese warms, weakens and flows slowly outwards. Eventually the rind splits along discrete "faults" to accommodate the overall displacement.

- Rifting occurs at "releasing bends" along continental strike-slip faults. Here, the strike-slip fault trace makes a jog in map view creating a geometry that requires the crust to pull apart in order to accommodate regional movement. Normal faults develop parallel to the jog, and thus oblique to the regional trend of the strike-slip system. Normal faulting along a releasing bend can lead to formation of a pull-apart basin (see Chapter 19).
- As we will see in Chapter 17, a zone of extension may develop in back-arc regions at convergent margins. In the case of continental arcs, the resulting backarc extensional zone becomes a rift.
- Finally, rifting may develop in the foreland of collisional orogens as a consequence of indentation. As further discussed in Chapter 17, if the lateral margins of the foreland are unconstrained, the collision of a rigid continental block with another continent may cause portions of the continent to squeeze sideways, in order to get out of the way, a process called lateral escape. In the region between the escaping blocks, the lithosphere stretches and rifts develop. These rifts trend roughly perpendicular to the trace of the orogen.

16.10 CLOSING REMARKS

This chapter was devoted to a discussion of rifting, the process by which continental breakup occurs. In some cases, rifts evolve into divergent plate boundaries, at which a new oceanic basin, bordered by passive margins, begins to grow (Figure 16.28). Rifting and seafloor spreading comprise the divergent end of the plate-tectonic conveyor. In the next chapter, we jump to the other end of the plate-tectonic conveyor and consider convergent tectonics, which is the process by which the oceanic lithosphere sinks back into the mantle, and collisional tectonics, which is the process by which buoyant crustal blocks merge and squeeze together.

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