CHAPTER SEVENTEEN

Convergence and Collision

17.1 INTRODUCTION

The Andes Mountains, a 6,000-km long rampart of rugged land speckled with several peaks over 6 km high, rim the western edge of South America (Figure 17.1a). Along the range, powerful volcanoes occasionally spew clouds of ash skyward. Halfway around the world, Mt. Everest, the highest mountain in the Himalayan chain (and in the world), rises 8.5 km above sea level (Figure 17.1b). At its peak, air density is so low that climbers use bottled oxygen to stay alive. Why did the immense masses of rock comprising such mountains rise to such elevations? Before the 1960s, geologists really didn't know. But plate tectonics theory provides a ready explanation—the Andes Mountains formed where the Pacific Ocean floor subducts beneath South America along a convergent plate boundary, while the Himalayan Mountains rose when India rammed into Asia, forming a collisional orogen.

Complex suites of structures (involving thrust faults, folds, and tectonic foliations) develop at convergent plate boundaries and collisional orogens. As a consequence, the crust shortens and thickens. In the process, metamorphism and, locally, igneous activity takes place. And, though it may seem surprising at first, gravity can cause the high regions of convergent and collision orogeny to collapse and spread laterally, yielding extensional faulting. In this chapter, we describe both the structural features and the rock assemblages that develop during convergent-margin tectonism and continental collision.

(a)

(b)

FIGURE 17.1 (a) Photo of the southern Andes Mountains of Chile. Rocks exposed on these peaks include relicts of old accretionary prisms, as well as granitic intrusions of a continental volcanic arc. (b) Photo of the central Himalaya Mountains, Nepal. The highest peak, which appears to be nucleating a cloud, is Mt. Everest, the highest mountain on Earth.

17.2 CONVERGENT PLATE MARGINS

When oceanic lithosphere first forms at a mid-ocean ridge, it is warm and relatively buoyant. But as lithosphere moves away from the ridge axis, it cools and the lithospheric mantle thickens, so that when lithosphere has aged more than 10 or 15 million years, it becomes negatively buoyant. In other words, old oceanic lithosphere is denser than underlying hot asthenosphere, and thus can **subduct** or sink into the asthenosphere, like an anchor sinks through water. Such subduction occurs at a **convergent plate boundary** (which may also be called a **subduction zone** or a **convergent margin**). Here, oceanic lithosphere of the **downgoing plate** (or **downgoing slab**) bends and sinks into the mantle beneath the **overriding plate** (or **overriding slab**). An overriding plate can include either oceanic crust or continental crust, but a downgoing plate can include only oceanic crust, because continental crust is too buoyant to subduct.

Exactly how the subduction process begins along a given convergent plate boundary remains somewhat of a mystery. Possibly it is a response to compression across a preexisting weakness such as may occur at a contact between continental and oceanic lithosphere along a passive continental margin, at a transform fault, or at an inactive mid-ocean ridge segment. Conceivably, compression causes thrusting of the overriding plate over the soon-to-be subducting plate (Figure 17.2a and b). Once the subducting plate turns down and enters the asthenosphere, it begins to sink on its own because of its negative buoyancy. The subducting plate pulls the rest of the oceanic plate with it and gradually draws this plate into the subduction zone. In other words, because of its negative buoyancy, the subducted plate exerts a **slab-pull force** on the remaining plate and causes subduction to continue.

Note that the process of subduction resembles the peeling of a wet piece of paper off the bottom of a table when you pull on one end. During the process of subduction, the position of the bend in the downgoing slab migrates seaward with time, relative to a fixed reference point in the mantle; this movement is called **rollback** (Figure 17.3). When the subducting slab reaches a depth of about 150 km, it releases volatiles (H_2O and CO_2) into the overlying asthenosphere, triggering partial melting of the asthenosphere. The melt rises, some making it to the surface, where it erupts to form a chain of volcanoes called a **volcanic arc.**

Convergent plate boundaries presently bound much of the Pacific Ocean. In fact, the volcanism along these boundaries led geographers to refer to the Pacific rim as the "ring of fire." Other present-day convergent plate boundaries define the east edge of the Caribbean Sea, the east edge of the Scotia Sea, the western and southern margin of southeast Asia, and portions of the northern margin of the Mediterranean Sea (see Figure 14.14d). In the past, the distribution of convergent plate boundaries on the surface of the Earth was much different. For example, during most of the Mesozoic, the west coast of North America and southern margins of Europe and Asia were convergent plate margins, but convergence at these localities ceased during the Ceno-
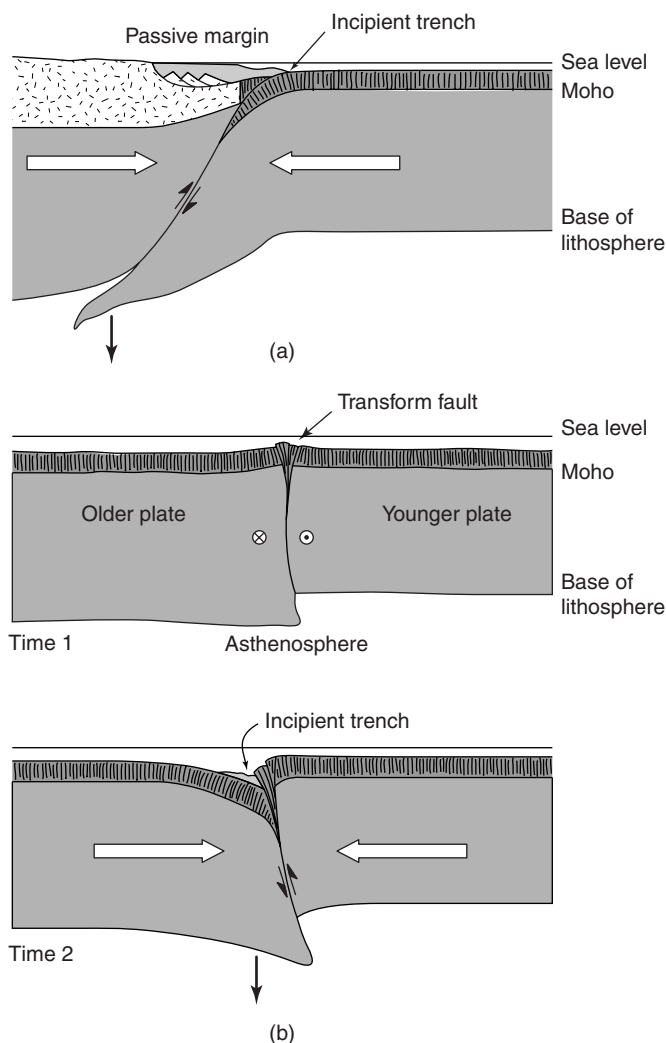
FIGURE 17.2. (a) Cross section that illustrates how a convergent margin may initiate along a passive continental margin. Here, thrusting of the margin over the denser oceanic plate has just begun. (b) Two stages during the evolution of an oceanic transform fault into a convergent plate boundary. At Time 1, two plates of different age are in contact along a transform fault. At Time 2, compression has developed across the fault, and it has become a thrust fault. The older plate has just begun to bend and sink into the asthenosphere.

zoic. Some convergent plate boundaries mark localities where oceanic lithosphere subducts under oceanic lithosphere (e.g., along the Mariana Islands and Aleutian Islands), and others mark localities where oceanic lithosphere subducts under continental lithosphere (e.g., along the Andes).

If you were to make a traverse from the abyssal plain of an ocean basin across a convergent plate boundary, you would find several distinctive tectonic features (Figure 17.4). A deep trough, the **trench,** marks the actual boundary between the downgoing and overriding plates. An **accretionary wedge** or **accretionary prism,** consisting of a package of intensely

FIGURE 17.3 The concept of rollback. As a subducting slab sinks into the asthenosphere, the position of the trench relative to the a fixed point in the mantle migrates. This movement is called rollback.

deformed sediment and oceanic basalt, forms along the edge of the overriding plate adjacent to the trench. Undeformed strata of the **forearc basin** buries the top of the accretionary prism and, in some cases, trapped seafloor or submerged parts of the volcanic arc. This basin lies between the exposed accretionary prism and the chain of volcanoes that comprises the **volcanic arc.** For purposes of directional reference, we refer to the portion of a convergent margin region on the trench side of a volcanic arc as the **forearc region,** while we refer to the portion behind the arc as the **backarc region.** These and related terms are summarized in Table 17.1.

To get an overview of what a convergent plate margin looks like, we now take you on a brief tour from the ocean basin across a convergent plate margin. We start on the downgoing slab, cross the trench, climb the accretionary wedge, and trundle across the forearc basin and frontal arc into the volcanic arc itself. We conclude our journey by visiting the backarc region.

17.2.1 The Downgoing Slab

The first hint that oceanic lithosphere is approaching a subduction zone occurs about 250 km *outboard* of the trench (i.e., in the seaward direction, away from the trench). Here, the surface of the lithosphere rises to form a broad arch called the **outer swell** or **peripheral bulge** (Figure 17.5a). The elevation difference between the surface of an abyssal plain of normal depth and the crest of the swell itself, is about 500–800 m. Outer
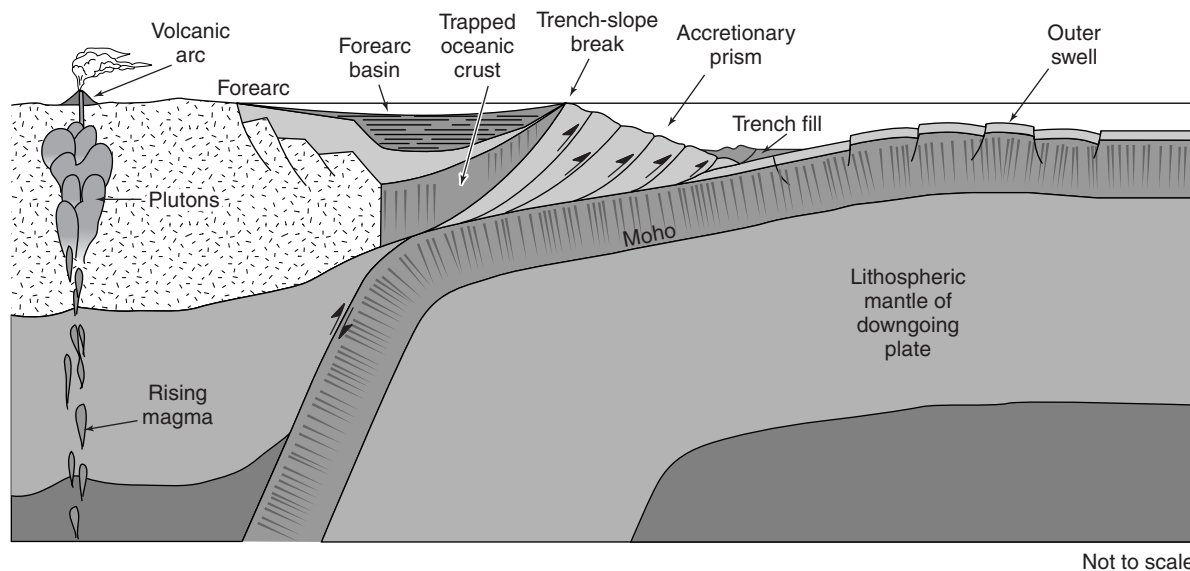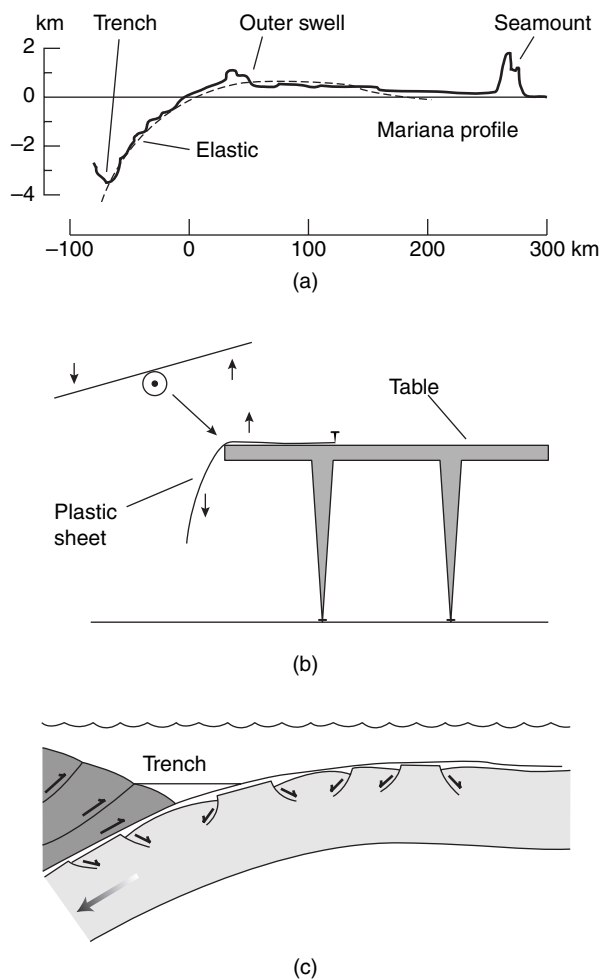

FIGURE 17.4 Idealized cross section of a convergent plate margin and related terminology. In this case, the margin occurs along the edge of a continent, bordered by a sliver of trapped oceanic crust.

TABLE 17.1	TERMINOLOGY OF CONVERGENT PLATE BOUNDARIES

Accretionary prism	A wedge of deformed sediment, and locally deformed basalt, that forms along the edge of the overriding slab; the material of the accretionary prism consists of pelagic sediment and oceanic basalt scraped off the downgoing plate, as well as sediment that has collected in the trench.
Arc-trench gap	The horizontal distance between the axis of the volcanic arc and the axis of the trench.
Backarc basin	A narrow ocean basin located between an island arc and a continental margin.
Backarc region	A general term for the region that is on the opposite side of the volcanic arc from the trench.
Continental arc	A volcanic arc that has been built on continental crust.
Convergent plate boundary	The surface between two plates where one plate subducts beneath another; as a consequence of subduction, oceanic lithosphere is consumed.
Coupled subduction	A type of subduction in which the overriding plate is pushing tightly against the downgoing plate.
Décollement	Synonym for detachment; the term is French.
Detachment	A basal fault zone of a fault system; in accretionary prisms it marks the top of the downgoing plate.
Downgoing slab (plate)	Oceanic lithosphere that descends into the mantle beneath the overriding plate.
Forearc basin	A sediment-filled depression that forms between the accretionary prism and the volcanic arc. The strata of a forearc basin buries the top of the accretionary wedge and/or trapped oceanic crust, and in general, are flat-lying.
Forearc region	A general term for region on the trench side of a volcanic arc.
Island arc	A volcanic arc built on oceanic lithosphere; the arc consists of a chain of active volcanic islands.
Marginal sea	Synonym for a backarc basin that is underlain by ocean lithosphere.
Mélange	A rock composed of clasts of variable origin distributed in a muddy matrix.
Oceanic plateau	A broad region where seafloor rises to a shallower depth. It is underlain by anomalously thick oceanic crust, probably formed above a large mantle plume.
Offscraping	The process of scraping sediment and rock off the downgoing slab at the toe of the accretionary prism.
Outer swell	A broad arch that develops outboard of the trench, in response to flexural bending of the lithosphere at the trench.
Overriding plate (slab)	The plate beneath which another plate is being subducted at a convergent plate boundary.
Peripheral bulge	Synonym for outer swell.
Retroarc basin	Synonym for backarc basin.
Rollback	The seaward migration of the bend in the downgoing plate as subduction progresses.
Subduction	The process by which one plate sinks into the mantle beneath another.
Subduction complex	Synonym for accretionary wedge.
Underplating	In the context of subduction, this is the process of scraping of material from the downgoing slab beneath the accretionary prism, so that the material attaches to the base of the prism.
Trench	A deep marine trough that forms at the boundary between the downgoing and overriding slabs; the trench may be partially or completely filled with sediment eroded from the volcanic arc or continental margin.
Trench slope break	Topographic ridge marking a sudden change in slope at the top of the accretionary wedge.
Uncoupled subduction	A process of subduction in which the downgoing plate is not pushing hard against the overriding plate, so the subduction system is effectively under horizontal tension.
Volcanic arc	A chain of subduction-related volcanoes.

km   Trench   Outer swell   Seamount
2
0
                              Mariana profile
−2
      Elastic
−4
  −100   0   100   200   300 km
            (a)

Table

Plastic
sheet

            (b)

Trench

            (c)

FIGURE 17.5 The peripheral bulge of the downgoing plate.
(a) Example of the peripheral bulge just east of the Mariana
Trench in the western Pacific. The solid line is a profile of the
actual surface of the downgoing slab, while the dashed line
is a calculated profile assuming the plate behaves like a sheet
with flexural rigidity. (b) Table-top model of a peripheral bulge,
with the inset showing lever concept. (c) Stretching along the
outer swell produces horsts and grabens at the surface of
the downgoing slab.

swells form because of the flexural rigidity of the lithosphere. Specifically, downward bending of the lithosphere at a convergent plate boundary levers up the lithosphere outboard of the trench and causes it to rise; you can illustrate this phenomenon by bending a sheet of plastic over the edge of a table (Figure 17.5b). Oceanic crust stretches to accommodate development of an outer swell and, as a result, an array of trench-parallel normal faults develops along the crest of the outer swell (Figure 17.5c).
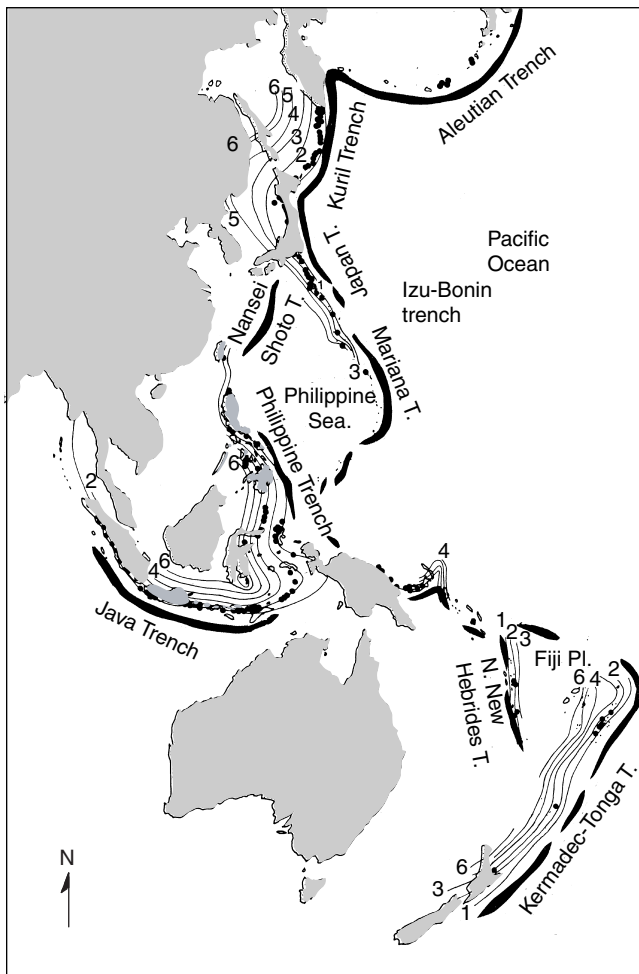
Not surprisingly, as a downgoing plate shears along the base of the overriding plate, rock ruptures abruptly, to cause earthquakes. But even after it has descended into the mantle, the downgoing plate remains seismi-

cally active. Earthquakes in subducting lithosphere define an inclined belt, called the **Wadati-Benioff zone,**[1] that reaches a maximum depth of around 670 km (Figure 17.6a, b, and c). In fact, it is the distribution of earthquakes in the Wadati-Benioff zone that defines the location of the subducted plate. Based on the shape of the Wadati-Benioff zone, researchers find that not all subducted plates dip at the same angle. In fact, dips vary from nearly 0°, meaning that the slab shears along the base of the overriding slab, to 90°, meaning that it plunges straight down into the mantle. Subducted-slab dip may be controlled, in part, by the age of the subducting lithosphere, for older oceanic plate is denser and may sink more rapidly. It may also be controlled by **convergence rate,** the horizontal rate at which plates are converging across the trench, for if we assume that sinking velocity is constant, an increase in convergence velocity decreases the dip of the subducting plate. The angle may also be affected by the flow direction and velocity of the asthenosphere into which the lithosphere sinks.
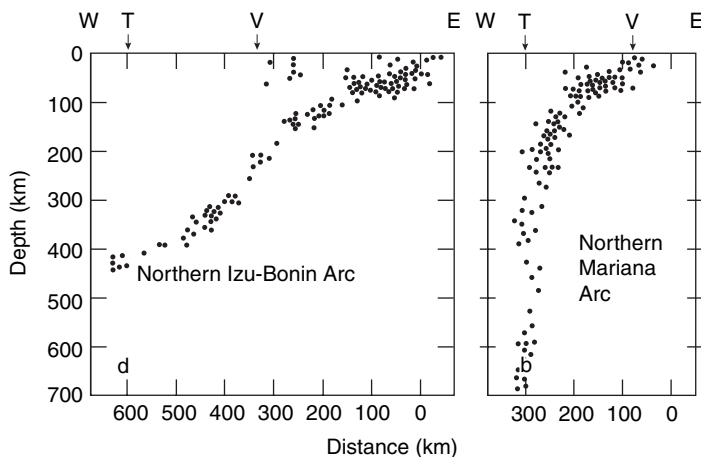
Because the lithosphere is cooler then the asthenosphere, the subducting plate perturbs the thermal structure of the mantle (Figure 17.7). The internal part of the downgoing plate remains relatively cool down to significant depths because rock has such low thermal conductivity. Under the pressure and temperature conditions found in the subducting plate, basalt of the oceanic crust undergoes a phase transition to become a much denser rock called **eclogite;** formation of eclogite may increase the slab-pull force.

The type of stress associated with earthquakes changes character with depth along the Wadati-Benioff zone (Figure 17.8). Beneath the outer swell, earthquakes result from tension caused by plate bending, whereas in the region beneath the accretionary wedge, earthquakes result from compression; thrust movements are due to shear between the overriding and downgoing slab. Huge, destructive earthquakes, such as the 1964 "Good Friday" earthquake of southern Alaska, result from ruptures in this zone. At depths of about 150–300 km, earthquakes of the Wadati-Benioff zone occur in a tensional stress field. Perhaps slab pull by the deepest part of the subducting plate stretches the plate in this interval. At deep levels, earthquakes of the Wadati-Benioff zone indicate development of compression, perhaps caused by shear between the deep downgoing plate and the asthenosphere. Seismologists do not understand why deep-focus earthquakes of the

[1]Named after its discoverers, K. Wadati (in Japan) and H. Benioff (in the USA) who worked independently and several decades apart.

(a)

(b)

Wadati-Benioff zone can occur, because at great depths, the downgoing slab should be warm enough to be ductile. Some researchers suggest that deep-focus earthquakes happen when sudden mineralogical phase transition or sudden dehydration reactions take place in rock comprising the downgoing plate, and that these cause an abrupt change in the volume of the rock; this change generates vibrations.

The deepest earthquakes of the Wadati-Benioff zone occur near the boundary between the seismically defined transition zone of the mantle and the lower mantle. Earthquakes from depths greater than 670 km have not been detected. But the deepest earthquakes do not necessarily define the greatest depth to which the downgoing slab sinks. Seismic tomography studies show that downgoing plates, because they are relatively cool, show up as bands of anomalously fast velocity. Some bands continue downwards into the lower mantle, suggesting that the downgoing slab flows downwards into the lower mantle. In fact, subducted plates may eventually sink almost to the base of the mantle, accumulating in "slab graveyards" near the core-mantle boundary (Figure 17.9). If this image is correct, then the base of the Wadati-Benioff zone does not mark the base of subducted slabs, but merely the depth at which earthquakes no longer occur because the fracturing and/or phase changes that produce seismic energy in slabs can no longer take place.

17.2.2   The Trench

Trenches are linear or curvilinear troughs that mark the boundary, at the Earth's surface, between the downgoing slab and the accretionary prism of the overriding plate (Figure 17.4). Trenches exist because the subducted portion of the downgoing slab pulls the slab downwards to a
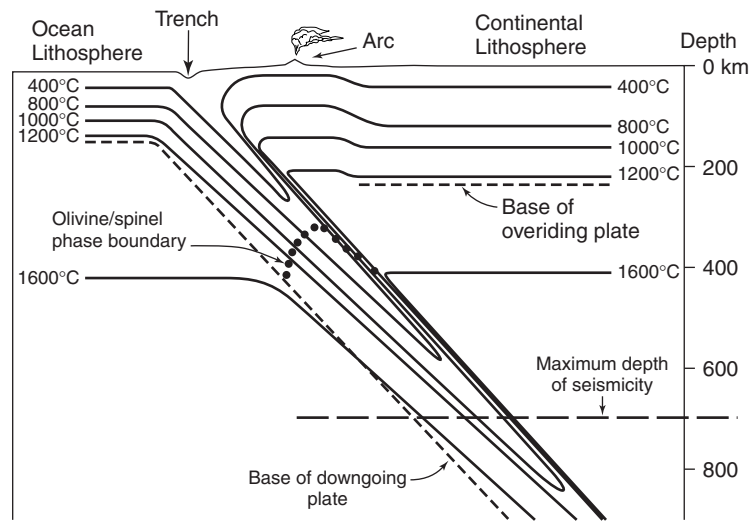
FIGURE 17.6   (a) Map of the western Pacific, showing trenches (heavy lines), and volcanoes (black dots) related to subduction in the western Pacific. The depth to Wadati-Benioff zones is shown by contour lines. The contours are given in multiples of 50 km (e.g., "2" means $2 \times 50 = 100$ km depth below the surface.) (b) Cross section showing earthquake foci defining the moderately dipping Wadati-Benioff zone of the northern Izu-Bonin arc. (c) Cross section showing earthquake foci defining the steeply dipping Wadati-Benioff zone of the northern Mariana Arc. T = location of trench; V = location of volcanic arc; the distance between T and V is the arc-trench gap.

FIGURE 17.7 A simplified model of the thermal structure of the downgoing plate. (The thermal effects of mineral phase changes are not shown.) Note that within the downgoing plate, relatively low temperatures are maintained to great depth. For example at a depth of 400 km, the asthenosphere is at about 1600°C, while the interior of the downgoing plate may be as cool as 750°C.
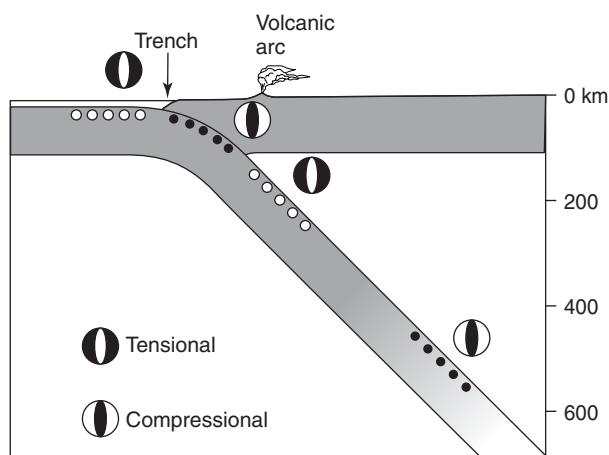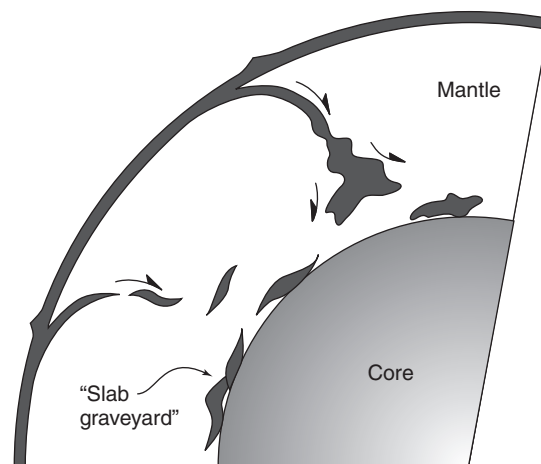

FIGURE 17.8 Schematic cross section illustrating the different types of earthquakes that occur at different depths within the downgoing plate.

FIGURE 17.9 Schematic cross section of the Earth illustrating the concept of a slab graveyard in which masses of subducted oceanic lithosphere may accumulate near the base of the mantle.

depth greater than it would be if the lithospheric plate were isostatically compensated.[2]

The deepest locations in the oceans occur at trenches. In fact, the floor of the Mariana Trench in the western Pacific (Figure 17.6a) reaches a depth of over 11 km, deep enough to swallow Mt. Everest (nearly 9 km high) without a trace. But not all trenches are so deep. For example, the Juan de Fuca Trench in the Pacific, off the coast of Oregon and Washington

(northwestern USA), is not much deeper than the adjacent abyssal plain of the ocean floor. Trench-floor depth reflects two factors: (1) the age of the downgoing slab (the floor of older oceanic lithosphere is deeper than the floor of younger oceanic lithosphere), and (2) the sediment supply into the trench (if a major river system from a continent spills into a trench, the trench fills with sediment). To see the effect of these parameters, let's compare the geology of the Mariana Trench and that of the Oregon-Washington Trench. The great depth of the Mariana Trench is a result of its location far from a continental supply of sediment and the fact that the plate being subducted at the Mariana Trench is relatively old (Mesozoic). In contrast, the

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[2]The resulting mass deficit from this depression at trenches produces a large negative gravity anomaly, which is a signature of subduction zones.
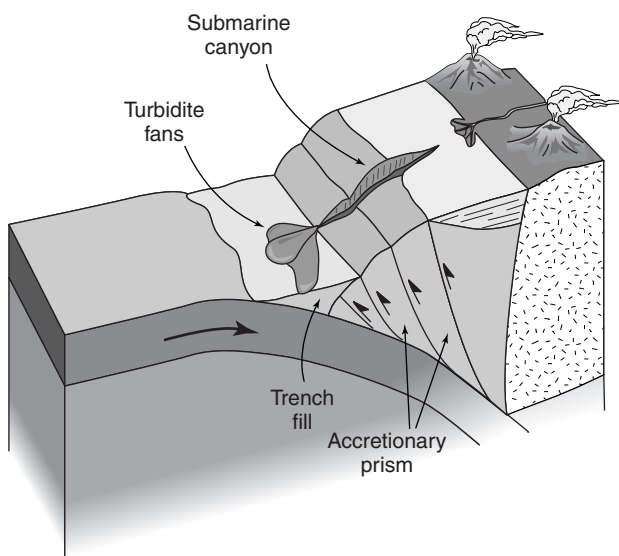
trench along the Pacific northwest margin of the United States has filled with sediments carried into the Pacific by the Columbia River, and the downgoing slab beneath the trench is quite young (Late Cenozoic).

Even though the thickness of sediments in trenches is variable, all trenches contain some sediment, called **trench fill.** Typically, trench fill consists of flat-lying turbidites and debris flows that decended into the trench via submarine canyons (Figure 17.10). The sediment comes from the volcanic arc and its basement, from the forearc basin, and from older parts of the accretionary wedge. Eventually, the trench fill becomes incorporated into the accretionary prism, where it becomes deformed.
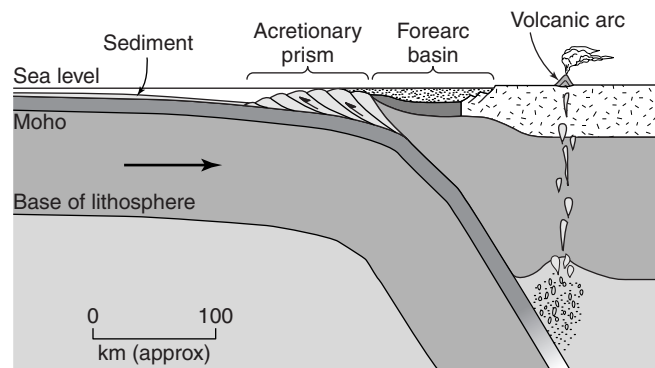
17.2.3 The Accretionary Prism

During the process of subduction, the surface of the downgoing plate shears against the edge of the over-riding plate. As we have already noted, the shear between the two plates produces an **accretionary prism** (or **accretionary wedge**). This is a wedge consisting of deformed pelagic sediment and oceanic basalt, which were scraped off the downgoing plate, and of deformed turbidite that had been deposited in the trench. Researchers have described two different accretionary prism geometries. Figure 17.11 illustrates these differences for the case of a convergent margin near a continent. In Figure 17.11a, the prism forms seaward of a trapped sliver of oceanic crust, whereas in 17.11b, the edge of the continent comes directly in contact with the surface of the downgoing plate.

Traditionally, the process of forming an accretionary prism has been likened to the process of forming a sand pile in front of a bulldozer (Figure 17.11c). The blade of the bulldozer can be called a **backstop,** in the sense that it is a surface that blocks the movement of material that had been moving with the downgoing

(a)

(b)

(c) Bulldozer analogy

(d) Sandbox model
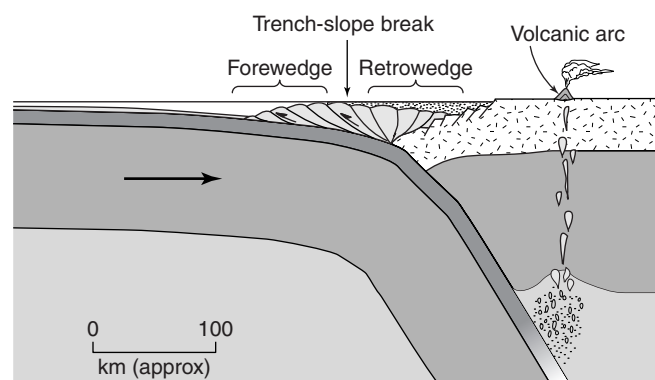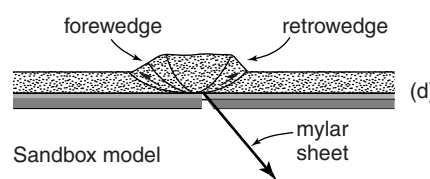
FIGURE 17.11 Two possible configurations of accretionary prisms (roughly to scale). (a) An accretionary prism caught beneath the lip of trapped ocean lithosphere. (b) An accretionary prism being scraped off the edge of a continent. Note that the prism, in this case, is bivergent. (c) The bulldozer analogy for the formation of an accretionary prism. The blade acts as the backstop. (d) Sandbox model showing the formation of a bivergent wedge.

FIGURE 17.10 Trenches fill with turbidites. Much of this sediment flows down submarine canyons and then accumulates in turbidite fans on the floor of the trench.
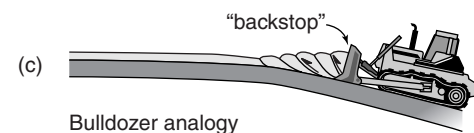
plate. Another way to visualize accretionary prism formation process comes from a simple sand-box model. In this model, a sand layer buries a sheet of mylar (thin plastic) that can be pulled through a slit in the base of the box; the slit represents the contact between the overriding and downgoing plates. As the mylar sheet moves down through the slit, the nonmoving sand on the "overriding plate" acts as a backstop, so the sand brought into the subduction zone piles up (Figure 17.11d). Note that in this configuration, a **bivergent wedge** (or **bivergent prism**) forms. This means that the prism consists of a forewedge and retrowedge. In the **forewedge,** the portion of the accretionary prism closer to the trench, structures verge toward the trench (i.e., toward the downgoing plate), while the **retrowedge,** the portion of the prism closer to the arc, structures verge toward

the arc (i.e., toward the overriding plate). Note that the material of the wedge itself serves as the backstop.

Compressional deformation in the accretionary prism produces thrust faults, folds, and cleavage. But tectonic compressional stress is not the only cause of strain in a prism. Gravity sliding causes slumping of rock and sediment down the slope of the prism toward the trench. And once the prism has become very thick, it begins to undergo **extensional collapse** under its own weight, like soft cheese. This means that gravitational energy overcomes the strength of material at depth in the internal part of the prism, so this material spreads sideways, leading to horizontal stretching in the above prism. As a consequence of this stretching, the region near the surface of the prism undergoes normal faulting (Figure 17.12a and b).

(a)

Vertical exaggeration = 1.15×

(b)
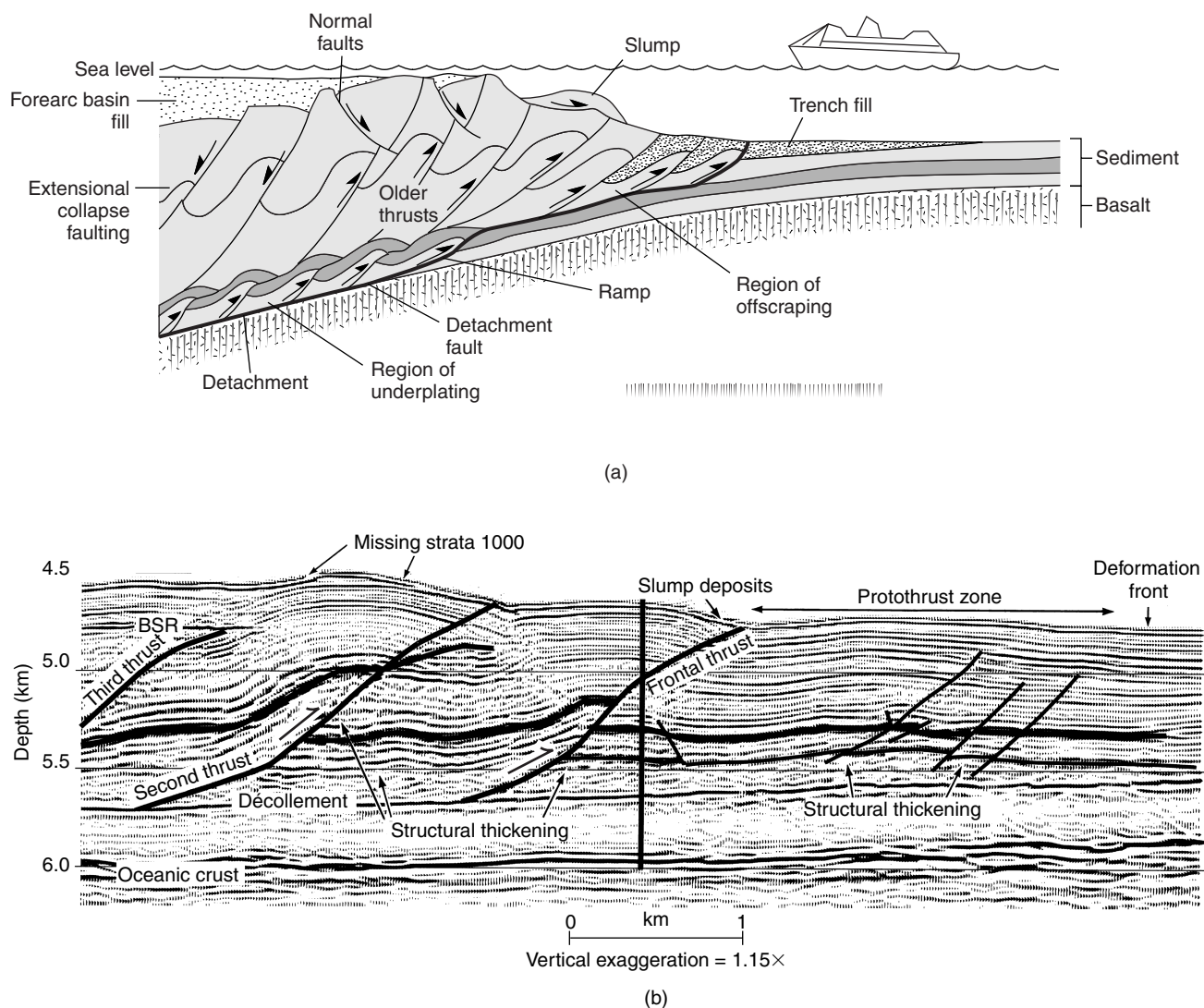
FIGURE 17.12 (a) Schematic detail of an accretionary prism, showing different regimes of deformation referred to in the text. (b) Interpreted seismic-reflection profile of the toe edge of an accretionary prism forming in the Nankai trough off Japan. Several faults can be imaged.

Let's first look a little more closely at the consequences of gravity-driven sliding down the slope of the prism. Such downslope movement events, which may be triggered by the relatively frequent earthquakes that occur in accretionary prisms, may lead to displacement of slump blocks (Figure 17.12), ranging from meters to tens of kilometers across. Very large (100s to 1,000s of meters long) slump blocks that remain semi-coherent during displacement are called **olistostromes.** The movement of slump blocks may lead to the formation of penecontemporaneous folding (see Chapter 1). Slope failure may also generate submarine debris flows in which muddy slurries flow downslope, and turbidity currents in which sediment disaggregates into an avalanche-like cloud that settles out in graded beds in the trench (Figure 17.10). Material that reaches the toe of the wedge will be recycled back into the wedge by offscraping. Thus, in some accretionary prisms, geologists find blocks composed of conglomerate containing clasts of conglomerate.

Now let's look more closely at the structural development of an accretionary prism. Effectively, an accretionary wedge is a fold-thrust belt that forms where the subducting plate slides under (underthrusts) an overriding plate (or, viewed from different perspective, where the overriding plate slides over the subducting plate). A **detachment fault** (i.e., **décollement**) delineates the boundary between the top of the subducting plate and the base of the accretionary prism and defines the actual plate boundary. This detachment ramps to progressively shallower stratigraphic levels toward the trench. At the toe of the prism the décollement grows and ramps up into trench strata or pelagic strata. As soon as strata becomes part of the hanging wall of the fault, it becomes, by definition, part of the overriding plate. The processes of detachment growth and ramping continue as subduction brings more material into the plate boundary, so with time, more and more material adds to the toe of the prism. This process is called **offscraping** (Figure 17.12).

Meanwhile, at the base of the accretionary prism, a duplex forms as the basal detachment cuts trenchward into strata of the downgoing plate and then ramps up to merge with a higher-level detachment. When a new **horse** (lens of rock or sediment surrounded by faults) forms, the material of the horse, by definition, becomes part of the overriding plate. Thus, material can be transferred from the downgoing plate to the overriding plate at the base of the accretionary prism, a process called **tectonic underplating**[3] (Figure 17.12). While faulting

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[3]Not to be confused with magmatic underplating, the process of adding basalt to the base of the crust along the margins of rifts or above mantle plumes.

takes place, the trench fill continues to be deposited over the subducting ocean floor. With continued deformation, faults in the internal part of the accretionary prism progressively steepen and rocks become penetratively strained. These processes lead to thickening within the prism.

Throughout most of an accretionary prism, thrusts and associated folds verge toward the trench, but along the arc side of the trench, thrusting may verge toward the arc. In this regard, accretionary prisms can be **bivergent** (Figure 17.11b). As noted earlier, you can simulate the development of a bivergent accretionary prism with a simple sand-box experiment (Figure 17.11d). Take a wooden box and cut a slit (representing a trench) in the floor of the box, about one-quarter of the distance in from one end of the box. Now, place a sheet of mylar (thin plastic) on the base of the box, and run the end through the slit. Bury the floor of the whole box (including the mylar) with a layer of sand. As you begin to pull the mylar sheet through the slit, to simulate subduction, the sand above the sheet moves toward the slit. Sand piles up above the slit, because the moving sand on the mylar collides with the stationary sand on the overriding side of the slit, and as this happens, thrust faults develop on both sides of the slit to accommodate the sand buildup. Those thrusts formed on one side of the slit verge in the direction opposite to the thrusts on the other side, creating a bivergent wedge. Note that the material of the wedge itself serves as the backstop that causes material to be scraped off the downgoing plate.

Recall that thrusting is not the only type of faulting to occur in an accretionary prism. When an accretionary prism reaches a substantial thickness, the upper part may undergo extensional collapse leading to the development of normal faults. Slip on these faults results in exhumation (uplift and exposure) of deeper rocks (Figure 17.12a).

The formation of an accretionary prism, as we discuss further in Chapter 18, can be described by the concepts of **critical taper theory.** To picture the essentials of this theory, imagine a plow blade moving into a layer of sand. The blade acts as a rigid **backstop** that transmits stress into the sand. (At a convergent plate boundary, the edge of the overriding plate and, later, the already accreted portions of the wedge act as a backstop with respect to new material being added to the wedge.) According to critical taper theory, a dynamic balance develops to maintain the **critical-taper angle,** the angle between the seaward surface of the prism and the surface of the downgoing plate. To maintain this angle, processes that cause the prism to grow wider, relative to its height (e.g., offscraping,
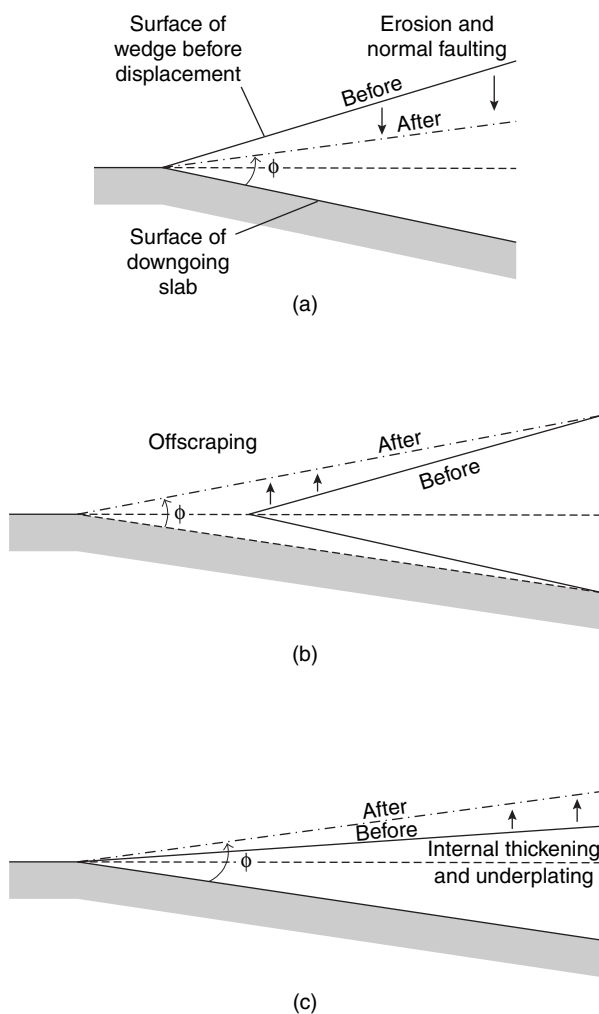
FIGURE 17.13 Cross sections illustrating maintenance of critical taper angle in an accretionary wedge. The solid line represents the wedge when it does not have a critical taper (i.e., before adjustment) and the dot-dash line represents the wedge that does have a critical taper. (a) If the wedge slope is too steep, steepness can be decreased by erosion and normal faulting in the upper part of the wedge. (b) If the slope is too steep, steepness can also be decreased by offscraping and building out of the wedge at the toe. Note the surface of the downgoing slab sinks. (c) If the slope is not steep enough, steepness can be increased by underplating and internal thickening of the wedge.

normal faulting, and trenchward slumping), compete with processes that cause the prism to grow thicker, relative to its width (e.g., underplating and penetrative shortening). The taper angle of the wedge can also be decreased as a consequence of trenchward slumping or by normal faulting in higher parts of the wedge (Figure 17.13a). If internal shortening and underplating cause the internal part of the wedge to thicken, so that the surface slope becomes steeper, then the taper angle of the wedge becomes too large. At this time, the wedge as a whole slides seaward and new material is
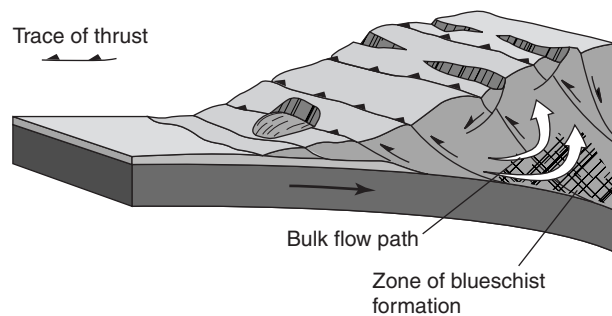
FIGURE 17.14 Bulk flow path of sediment in an evolving accretionary wedge. The large arrows trace the average movement of a grain of material. Note that during its movement, the grain may enter the region of blueschist metamorphism.

added to the toe of the wedge by offscraping, the net result being that the taper angle decreases below the critical value again (Figure 17.13b). If, alternatively, the taper angle becomes less than the critical value, then internal strain of the wedge and underplating occurs, thickening the wedge and resulting in an increase of the taper angle (Figure 17.13c).

The overall consequence of deformation and mass-wasting processes in accretionary wedges results in long-term internal "circulation" of material within the wedge. During this process material first moves down to the base of the wedge and then moves back up toward the surface (Figure 17.14). Thus, sediment that accreted to the base of the wedge by subduction may later end up exposed at the surface of the wedge. This movement reflects both internal thrusting in the wedge that pushes material up, and normal faulting and slumping that strips away the overlying material of the wedge (Figure 17.14). Such net material flow within an accretionary wedge explains how blueschist, formed at the base of the wedge, can eventually be brought to the surface of the wedge.

The combination of tectonic deformation, gravity-driven slumping, and extensional collapse that takes place in prisms makes them structurally complex and heterogeneous. In some places, accretionary prisms consist of coherent sequences of strata containing parallel arrays of folds and faults and an axial-planar cleavage. Elsewhere, prisms consist of **broken formation** in which beds can be traced for only a short distance before they terminate at another lithology, or **mélange,** a chaotic mixture of different rock types (Figure 17.15).[4] In mélange, bedding cannot be traced very far at all, and rocks of radically different lithology and metamorphic grade are juxtaposed.

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[4]From the French word for "mixture."

FIGURE 17.15   Photograph of Paleozoic mélange exposed in an outcrop in north-central Newfoundland. Note how layers are disrupted, strongly cleaved, and complexly folded, and that a variety of different rock types are visible.

On a traverse across an accretionary prism, you will find exposures of slate and lithic sandstone (derived from trench-fill turbidites), bedded chert (derived from pelagic silicic ooze), micrite (derived from pelagic carbonate ooze), greenstone (altered seafloor basalt), and **blueschist** (metamorphic rocks containing a blue amphibole called glaucophane). The origin of blueschist puzzled geologists for years, because glaucophane can only form under unusual conditions of very high pressure (as occurs at depths >20 km) and relatively low temperature. Because of the geothermal gradient (rate of increase of temperature with depth) characteristic of continental crust, glaucophane does not form in normal continental crust. But once geologists began to understand the process by which accretionary prisms formed, the location of blueschist formation became clear. Blueschists form at the base of the deepest part of the prism, where pressures are great, due to the overburden of 20 km of sediment, but temperatures are relatively low because the underlying oceanic lithosphere is relatively cold.

17.2.4   The Forearc Basin and the Volcanic Arc

As we continue our tour up the slope of the accretionary prism and toward the volcanic arc, we find that the top of the prism is defined by an abrupt decrease in slope. This topographic ridge is the **trench-slope break** (Figure 17.11). In a few locations around the world (e.g., Barbados, east of the Lesser Antilles volcanic arc along the east edge of the Caribbean), the trench-slope break emerges above sea level.

At many convergent margins, a broad shallow basin covers the region between the trench-slope break and the volcanic arc (Figure 17.11). This **forearc basin** contains flat-lying strata derived by erosion of the arc and the arc's substrate. Typically, strata of the forearc basin overlie older, subsided, portions of the prism. But locally, these strata overlie ocean lithosphere that had been trapped between the arc axis and the trench when subduction initiated. The strata may also overlie older parts of the volcanic arc and its basement.

The **volcanic arc** is the chain of volcanoes that forms along the edge of the overriding plate, about 100–150 km above the surface of the subducted oceanic lithosphere. As noted earlier, most of the magma that rises to feed the arc forms by partial melting in the asthenosphere above the surface of the downgoing slab. This partial melting takes place primarily because of the addition of volatiles (H_2O or CO_2) released from the downgoing plate into the mantle as the downgoing plate heats up. Some researchers argue that small amounts of melt may be derived locally from the downgoing plate.

Island arcs form where one oceanic plate subducts beneath another, or where the volcanic arc grows on a sliver of continental crust that had rifted from a continent; **continental arcs** grow where an oceanic plate subducts beneath continental lithosphere. Volcanism at island arcs formed on oceanic crust tends to produce mostly mafic and intermediate igneous rocks, whereas volcanism at continental arcs also produces intermediate and silicic igneous rocks, including massive granitic batholiths. The large volumes of silicic magmas in continental arcs form when hot mafic magmas rising from the mantle transfer heat into the surrounding continental crust and cause melting of the crust. While partial melting of mantle peridotite (an ultramafic rock) yields basaltic (mafic) magma, partial melting of mafic or intermediate continental crust yields intermediate to silicic magma.

The **arc-trench gap,** meaning the distance between the arc axis and the trench axis, varies significantly among convergent margins (Figure 17.6). Two factors control the width of the arc-trench gap at a given convergent margin: (1) *Dip of the downgoing slab:* Geometric principles dictate that if the downgoing slab dips very steeply, then the arc-trench gap must be narrow, but if the downgoing slab dips gently, then the arc-trench gap must be broad. (2) *Width of the accretionary prism:* Where subduction has continued for a long time, or where a large river fills the trench with sediment, the accretionary prism grows to be very large. When this happens, the prism acts as a weight that flexurally depresses the downgoing slab, and as the prism builds seaward, the trench location migrates seaward.

17.2.5 The Backarc Region

The **backarc region** refers to the region on the opposite side of the volcanic arc from the forearc basin. The structural character of backarc regions varies with tectonic setting (Figure 17.16). For the purpose of discussion, we define three types of backarc regions: (1) contractional, (2) extensional, and (3) stable.

In a **contractional backarc** (Figure 17.16a), a backarc basin does not form. Rather, crustal shortening generates a fold-thrust belt and/or a belt of basement-cored uplifts (see Chapter 18). Both types of deformation developed on the east side of the Andes during the Cenozoic, so a contractional backarc is commonly called an **Andean-type backarc.** The style of deformation that develops in a contractional backarc depends on the angle of subduction and on the nature of the crustal section in the overriding plate. If subduction angles are moderate to steep and the backarc region contains thick strata of a former passive margin, a fold-thrust belt develops. (Such a backarc fold-thrust developed during the Mesozoic Sevier Orogeny in western North America.) If, alternatively, subduction angles are shallow, so the subducted plate shears along the base of the overriding plate, stress activates preexisting basement-penetrating faults further to the foreland, and reverse-sence movement on these faults uplifts basement blocks. Overlying strata drape over the block uplifts and form monoclinal folds. (Such block uplifts, and associated monoclines, formed in the Rocky Mountain region of the United States during the 80–40 Ma Laramide Orogeny.)

In **extensional backarc** regions, crustal stretching takes place (Figure 17.16b). This stretching produces a **backarc basin.** If the stretching produces only a continental rift, then continental crust underlies the backarc basin, but if seafloor spreading takes place, then oceanic crust underlies the backarc basin. A clear example of a backarc basin formed by seafloor spreading occurs behind the Mariana Island Arc in the western Pacific Ocean (Figure 17.7), so extensional backarcs are commonly called **Mariana-type backarcs.** Backarc basins appear to initiate by rifting along the length of the volcanic arc. When the rift evolves into a new mid-ocean ridge, it splits off a slice of the volcanic arc, and then seafloor spreading separates this now-inactive slice of arc crust from the still active volcanic arc. A slice of arc crust, separated from the active arc by a new segment of seafloor, is called a **remnant arc.** Large backarc basins like the Philippine Sea contain several remnant arcs. In the Philippine Sea these appear to have been produced by a succession of separate rifting episodes, each of which yielded a short-lived mid-ocean ridge.

In a **stable backarc,** no strain accumulates (Figure 17.16c). Some stable backarcs may have once been contractional or extensional, but then later became stable when plate motions changed. Others are composed of oceanic lithosphere trapped behind the arc when a convergent margin developed far off the coast of the passive margin. The Bering Sea, for example, is

(a)

(b)
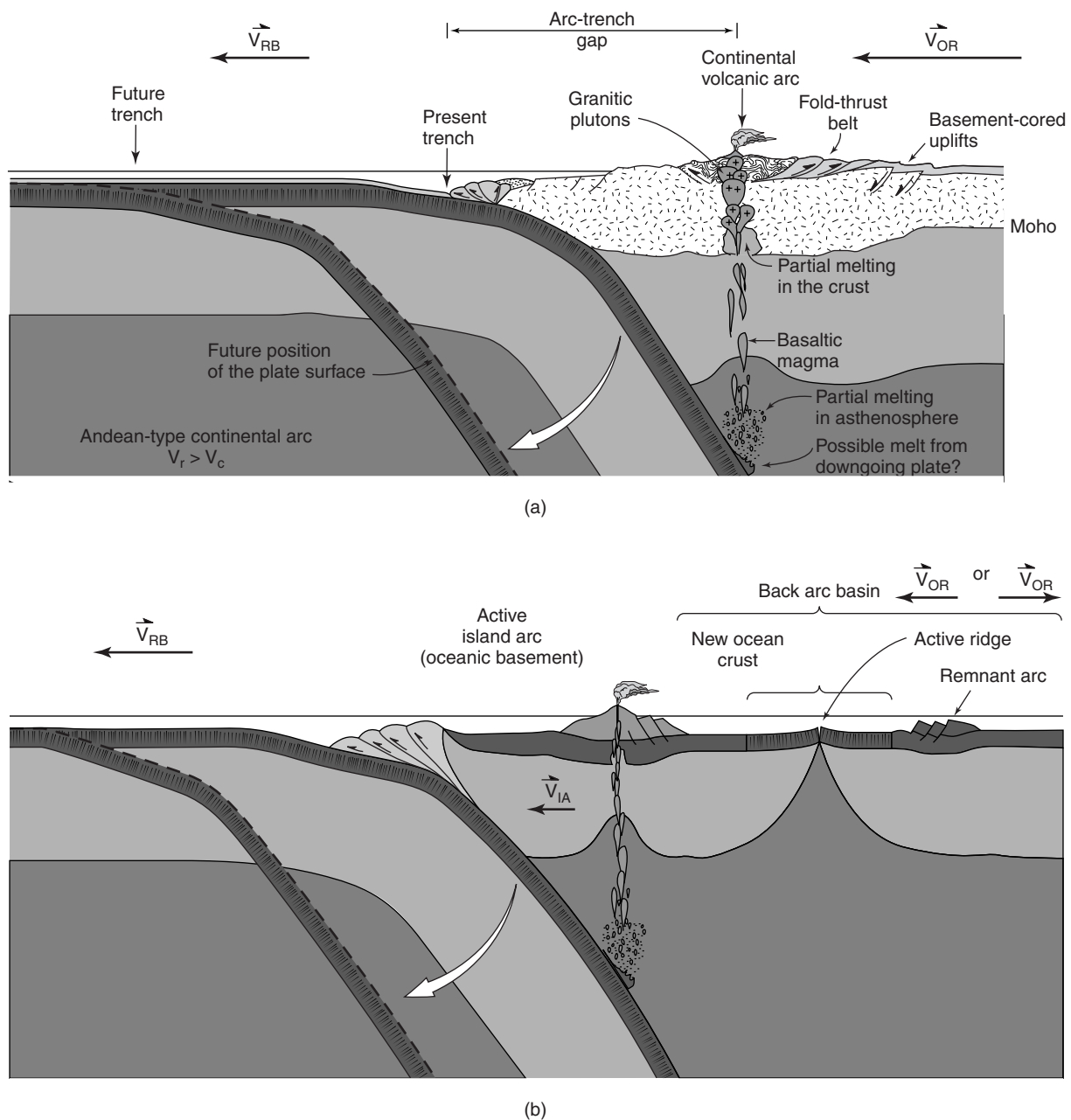
FIGURE 17.16 Different kinds of volcanic arcs and backarc regions. (a) An Andean-type continental arc, with a compressional backarc region. Here, the volcanic arc grows on continental crust, and compression has generated a fold-thrust belt and "Laramide-style" basement-cored uplifts. Large granitic plutons develop. This situation develops where the velocity of the overriding plate (v_{OR}) is in the same direction and exceeds the rollback velocity (v_{RB}). (b) A Mariana-type island arc, with an extensional backarc. Here, the volcanic arc grows on oceanic crust, and a backarc basin develops in which there is seafloor spreading. A remnant arc, composed of a rifted-off fragment of the arc may occur in the basin. This situation develops when $v_{OR} < v_{RB}$, or is in the opposite direction to v_{RB}. The island arc must move to keep up with the rollback. (c) A Japan-type volcanic arc, in which the island arc has continental basement that had rifted off a continent when a backarc basin grew. Here, the backarc spreading has ceased, and the backarc is stable. This is because $v_{OR} = v_{RB}$. A strike-slip fault could develop in the backarc region.
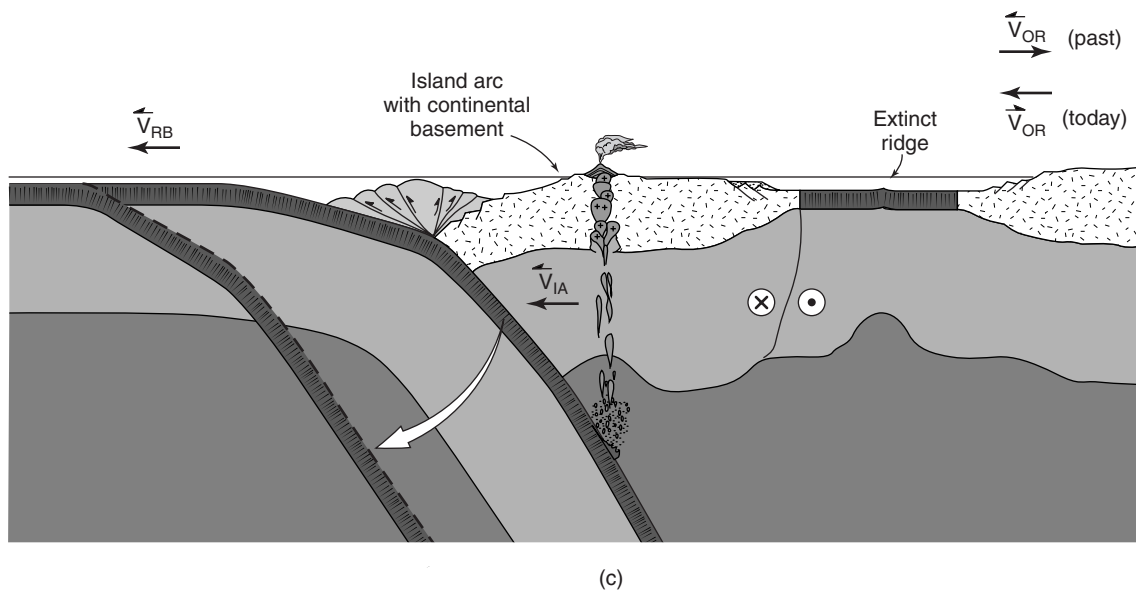
$\vec{V}_{OR}$ (past)

Island arc
with continental
basement

$\vec{V}_{RB}$

Extinct
ridge

$\vec{V}_{OR}$ (today)

$\vec{V}_{IA}$

(c)

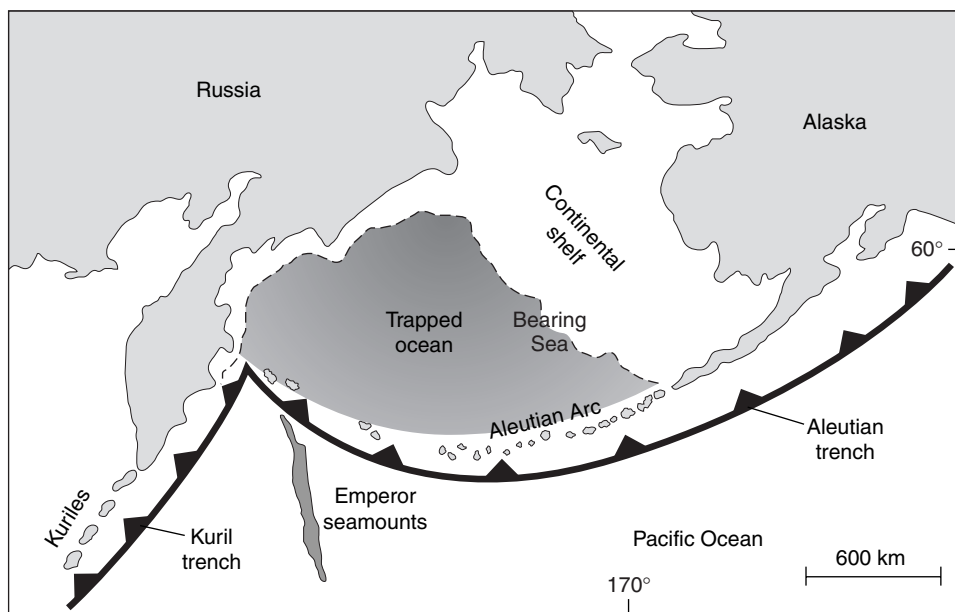**FIGURE 17.16** (Continued)

FIGURE 17.17 The Bering Sea is an example of a backarc region formed from trapped ocean. Here, the oceanic lithosphere that now lies behind the arc formed in the Mesozoic—it is not a consequence of Cenozoic backarc spreading.

underlain by Mesozoic-age trapped ocean floor that was isolated from the rest of the Pacific plate when the Aleutian volcanic arc formed (Figure 17.17).

Why do we observe such a wide range of kinematic behavior in backarc regions? The answer comes from examining the relative motion between the backarc region and the volcanic arc. As subduction progresses, the location of the bend in the downgoing plate rolls back, away from the backarc (Figure 17.18). The axis of the volcanic arc moves with the rollback. Thus, if the overriding plate is moving in the same direction but at a rate faster than rollback, a contractional backarc develops. If, however, the overriding plate is stationary or moves away from the trench, then rifting and a

backarc basin develop. And if the overriding plate moves in the direction of rollback at the same rate as rollback, then the backarc region is stable. If there is a component of lateral motion between the overriding and downgoing plates, then some strike-slip faulting may also occur in the backarc.

Backarc regions can evolve with time. For example, the Japan Sea (Figure 17.7) started as an extensional backarc. In fact, Japan's basement consists of continental crust that originally linked to eastern Asia; the islands separated from the rest of Asia when seafloor spreading in the backarc produced the Japan Sea. Presently, however, the character of earthquakes indicates that shortening, accompanied by strike-slip
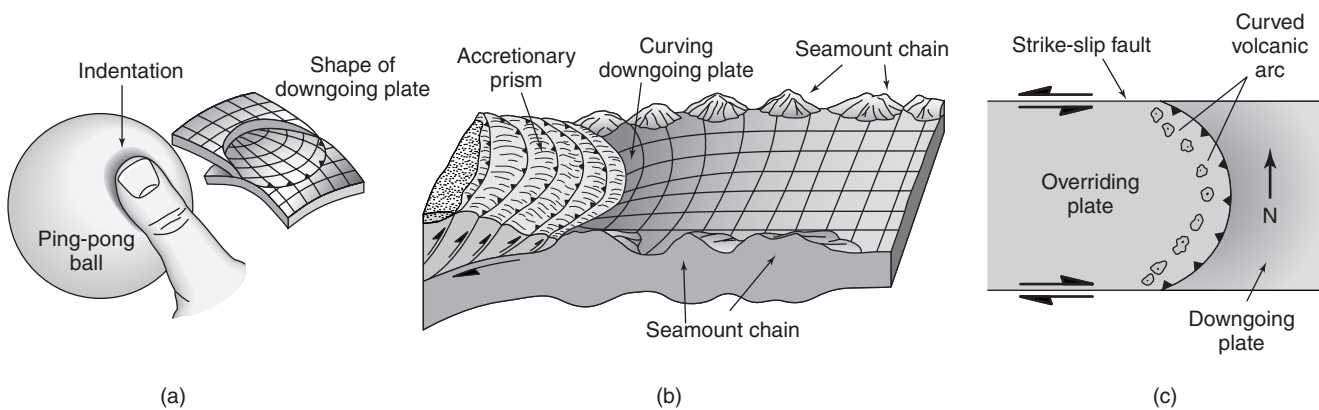
FIGURE 17.18   Reasons for the curvature of volcanic arcs. (a) If you make a thin slit on the surface of a ping-pong ball, and push it in, the piece must pucker into a curve, to maintain its surface area; (b) If a chain of seamounts collides with the arc, it may cause indentations because the lithosphere of the seamount is more buoyant and subducts at a shallower angle, and because the seamounts act as an obstacle to growth of the accretionary prism. (c) Strike-slip and drag origin for island-arc curvature. Shear on strike-slip faults may bend the volcanic arc in map view.

deformation, now takes place in the Japan Sea. Because of this motion, the Japan Sea may someday close, suturing Japan back to Asia.

17.2.6   Curvature of Island Arcs

The reason that island arcs are called *arcs* (e.g., Figure 17.17), is that many have a curved trace in map view. Let's now examine a few possible explanations for why such curvature exists.

First, a curve reflects the natural shape of an indentation on a sphere. Places where oceanic lithosphere subducts can be viewed as indentations on the surface of the spherical Earth; the downgoing plate must curve in order to maintain the same surface area for a given length of the convergent boundary, and if the downgoing plate curves in three dimensions, then the trench and volcanic arc must also curve on the Earth's surface (Figure 17.18a). To visualize this geometry, take a ping-pong ball and push in one side with your thumb—the trace of the indentation is a curve on the ball's surface.

Second, a curve forms where a seamount collides with an originally straight arc during subduction. To picture this geometry, look at a map of the western Pacific, and note that some of the major cusps in subduction zones coincide with sites of seamount subduction (Figure 17.18b). For example, the Emperor Seamount Chain subducts at the cusp between the Aleutian Trench and the Kuril Trench (Figure 17.17). This relationship may develop because a seamount acts as an obstacle that inhibits propagation of the accretionary wedge, and/or the buoyancy of a seamount decreases the rate of rollback, so that portions of the downgoing plate away from the seamount roll back faster than the plate under the seamount.

Third, strike-slip faulting causes map-view shear at the end of an arc (Figure 17.18c). To picture this geometry, look at a map of the region encompassing the southern end of South America, the Scotia Sea, and the northern tip of the Antarctic Peninsula. Transform faults delineate both the northern and the southern boundaries of the Scotia Plate. Conceivably, shear along the northern fault bent the southern end of South America and shear along the southern fault bent the northern end of the Antarctic Peninsula. Shear on these faults also caused the Scotia arc to bend into an arc. Alfred Wegener noted the shape of the Scotia Arc and used the shape as evidence for the westward drift of South America. A similar geometry of faulting may explain the curvature of the Lesser Antilles volcanic arc at the eastern edge of the Caribbean Sea.

17.2.7   Coupled versus Uncoupled Convergent Margins

Taking into account the description of convergent margins that we've provided earlier in this section, we see that not all convergent margins display the same suite of rocks and structures. Based on the contrasts among various convergent margins worldwide, geologists distinguish between two end-member types.

In a **coupled convergent margin,** the downgoing plate pushes tightly against the overriding plate, so the plate boundary overall is under compression. As a consequence, large shear stresses develop across the contact, causing efficient offscraping and tectonic underplating, and therefore buildup of a large accretionary
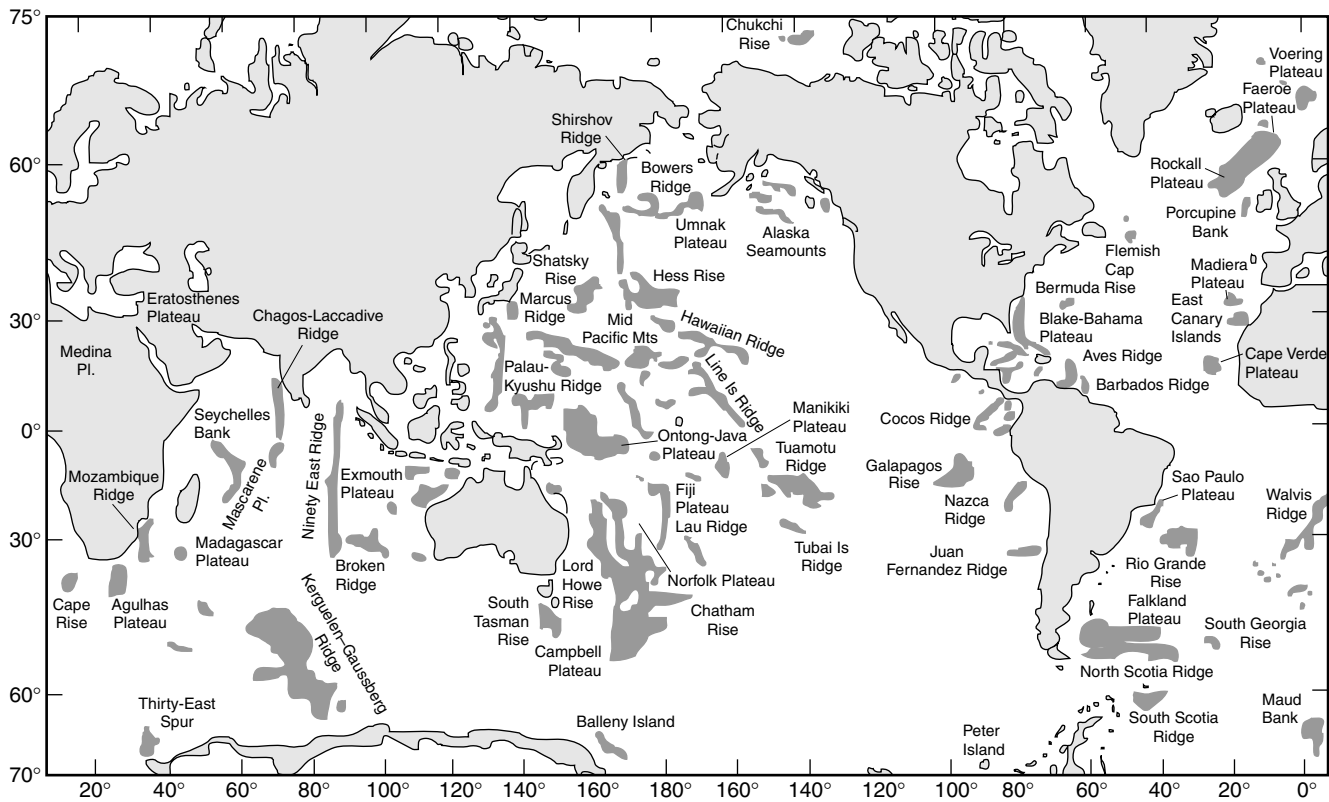
FIGURE 17.19   A map showing blocks of buoyant crust including volcanic arcs, oceanic plateaus, and continental fragments. Blocks that have darker shading rise above today's sea level. Blocks that have lighter shading are presently submerged. Convergence between any two blocks could cause collisional tectonics.

wedge. This shear stress also triggers devastating earthquakes, and development of a contractional backarc region. Perhaps because compression squeezes crustal fractures closed, magma rises slowly, and therefore has time to fractionate and/or cause partial melting of adjacent continental crust before intruding at shallow depth or erupting at the surface. Since partial melting of continental crust produces intermediate to felsic magma, and since fractionation removes mafic minerals from a melt, intermediate to felsic igneous rocks predominate at coupled convergent margins.

In an **uncoupled convergent margin,** the downgoing plate does not push tightly against the overriding plate, so compression across the margin is not great. As a consequence, shear stresses across the plate boundary are relatively small, thrust earthquakes at the boundary have smaller magnitude, and relatively little offscraping and underplating occurs. In uncoupled systems we find extensional backarcs, and cracks in the overriding plate remain somewhat open, so mantle-derived magmas rise directly to the surface before significant fractionation or crustal contamination occurs. Thus, mafic igneous rocks are more common at such convergent margins.

17.3   BASIC STAGES OF COLLISIONAL TECTONICS

As subduction consumes an oceanic plate, a piece of buoyant lithosphere attached to the downgoing plate may eventually be brought into the convergent boundary. Examples of buoyant lithosphere include large continents, small continental fragments, island arcs, oceanic plateaus (broad regions of anomalously thick oceanic crust, formed by hot-spot volcanism), and spreading ridges (Figure 17.19). Regardless of type, buoyant lithosphere generally cannot be completely subducted, and when it merges with the overriding slab, the boundary becomes a **collision zone.**[5] When the forces driving collision cease, the relative motion between the colliding blocks ceases, and when this happens, the once separate blocks of lithosphere have merged to become one. The shear surface that marks the boundary between these once-separate plates is a

[5]Here, we've described collision tectonics that follows convergent tectonics. Note that collision can also result from the closure of a rift, even if no oceanic lithosphere existed between the colliding blocks.

| TABLE 17.2	TERMINOLOGY OF COLLISION
Accreted terrane	A piece of exotic crust that has been attached to the margin of a larger continent. (Note the spelling of *terrane,* which differs from that of the geographic term for a tract of land, spelled *terrain.*)
Basin inversion	The process whereby a region that had undergone crustal extension during basin formation subsequently undergoes crustal shortening during collisional tectonics; in the process, faults that began as normal faults are reactivated as reverse faults, and the strata of the basin thrusts up and over the former basin margin.
Collision	An event during which two pieces of buoyant lithosphere move toward each other and squash together, after the intervening oceanic lithosphere has been subducted.
Delamination	In the context of collisional tectonics, this refers to the separation of the basal portion of the thickened lithospheric mantle beneath a collisional orogen; this delaminated lithospheric mantle then sinks downward through the asthenosphere.
Exotic terrane	An independent block of buoyant crust that has been brought into a convergent margin during subduction, where it collides and docks against the continent (also called **accreted terrane**). The adjective "exotic" in this context is used simply to emphasize that the block in question did not originate as part of the continent to which it is now attached, but rather came from somewhere else.
Lateral escape	The process, accompanying collision, during which crustal blocks of the overriding plate slide along strike-slip faults in a direction roughly perpendicular to the regional convergence direction; effectively, the "escape" from the collision zone resembles the movement of a watermelon seed that you squeeze between your fingers.
Orogenic collapse	The process that occurs when thickened crust in a collisional orogen weakens and starts to sink under its own weight. Effectively, gravitational loads cause horizontal extensional strain to develop. In some cases, extension is coeval with thrusting (synorogenic collapse), in other cases it occurs after thrusting has ceased (postorogenic collapse). (Also called *extensional collapse.*)
Suspect terrane	A crustal block in an orogen whose tectonic origin is unclear. The block does not appear to correlate with adjacent crust in the orogen where it now resides and thus may be exotic. A block remains "suspect" only until its origin (i.e., whether it's exotic or not) has been determined.
Suture	The shear surface within an orogen that marks the boundary between once-separate continents; commonly, slivers of ophiolites occur along a suture.
Tectonic collage	A region of crust that consists of numerous exotic terranes that have been sutured together; in other words, a tectonic collage consists of accreted terranes that docked during a protracted period of convergent-margin tectonism.

suture; slivers of ophiolites (ocean crust thrust over continental crust) crop out locally along sutures. We define these terms, as well as others used in the discussion of collisional tectonics in Table 17.2.

The types of rocks and structures formed during a particular collisional orogeny depend on numerous variables, including:

• *Relative motion between the colliding blocks.* **Frontal collisions** yield thrust faults whose movement is perpendicular to the edge of the colliding blocks, while strain in **oblique collisions** may be partitioned between thrusting and strike-slip faulting (Figure 17.20a). In addition, where blocks col-

lide obliquely, the timing of collision may be diachronous along strike (Figure 17.20b), and the blocks merge together like the two sides of a zipper.
• *Shape of the colliding pieces.* The collision of broad, smooth continental margins yields fairly straight orogens, while the collision of irregular continental margins (with **promontories,** which are seaward protrusions of the continental margin, and **recesses,** which are indentations along the continental margin) yields sinuous orogens, in map view. Promontories act as indenters that push into the opposing margin, creating a localized region of high strain. In some cases, slices of crust may move sideways (relative to the frontal collision direction) along strike-slip faults
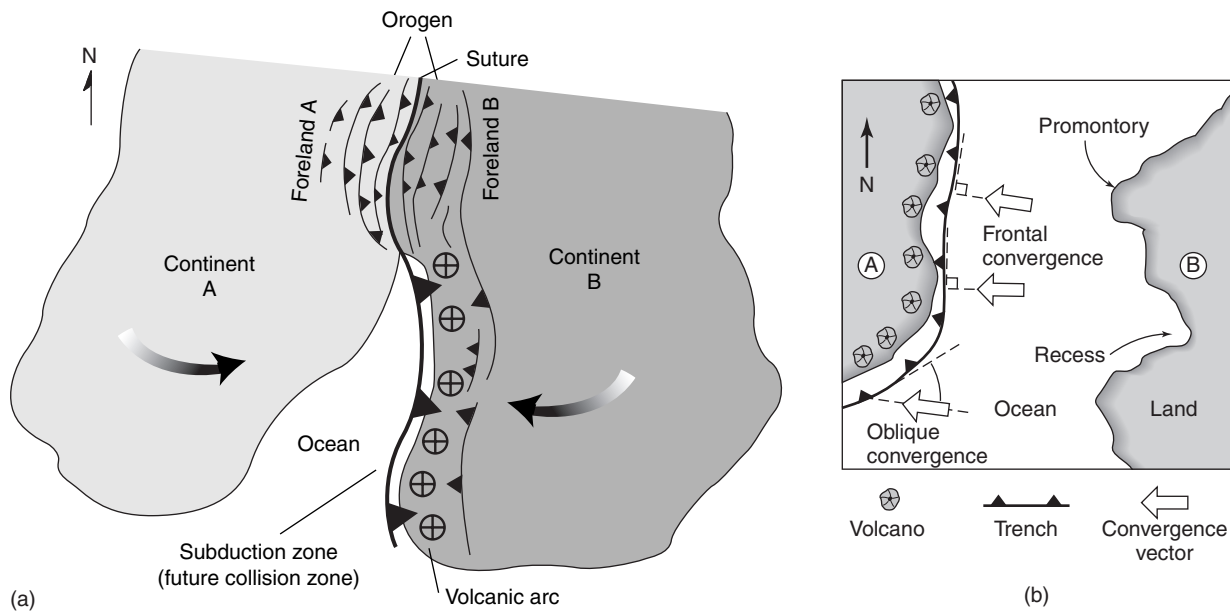
FIGURE 17.20   (a) A map showing a zipper-like collision between two continents. Here, the ocean between the two continents is closing progressively from north to south. In the collision zone, the boundary between what had originally been two separate continents. (b) A map showing the convergence of two continents. Promontories and recesses make the west coast of Continent B irregular. Because of the change in trend of the subduction zone along Continent A, frontal convergence (and, eventually, frontal collision) will occur to the north, while oblique convergence (and, eventually, oblique collision) will occur to the south.

to get out of the way of the colliding blocks. This movement is a type of **lateral escape,** and will be discussed further in Section 17.4.1.

- *Physical characteristics of the colliding pieces.* Physical characteristics, such as temperature, thickness, and composition, influence the way in which crustal blocks deform during collision. For example, warmer and, therefore, softer crust of a younger orogenic belt will develop greater strains during collision than will old, cold cratonic crust. During collision, a craton acts as a rigid indenter that pushes into the relatively soft, younger orogenic belt. The collision between India, an old craton, and southern Asia, a weak Phanerozoic orogen, illustrates such behavior. During this collision, much more deformation has happened in the weak southern margin of Asia than in strong India. In fact, a map of the collision zone (see section 17.4.1) shows that India has actually pushed into Asia, so that a transform fault now bounds each side of the Indian subcontinent.

Because so many variables govern the nature of a collisional orogeny, no two collisional orogenies are exactly the same. Nevertheless, we can provide a basic image of the collision process by outlining, in the following section, the various stages in an idealized collision between two continents (A and B; Figure 17.21a, b, c). For reference, we call the portion of the orogen that is on the craton side of the collision the **foreland** and the internal part of the orogen the **hinterland.**

17.3.1   Stage 1: Precollision and Initial Interaction

Let's begin by setting the stage for the collision between two continents. In this scenario, Continent A moves toward Continent B as the oceanic lithosphere connected to Continent A subducts beneath the margin of Continent B (Figure 17.21a). Note that the margin of Continent A is a passive margin, along which a passive-margin sedimentary basin has developed; in contrast, the margin of Continent B is an Andean-type convergent margin along which a volcanic arc has developed.

Continent A remains oblivious to the impending collision until the edge of the continent begins to bend, prior to being pulled into the subduction system by the downgoing plate (Figure 17.21b). When this happens, flexure causes the surface of the continental margin to rise, so that the continental shelf rises above sealevel and undergoes erosion. The margin of Continent A also undergoes stretching, and as a result, normal faults trending parallel to the edge of the margin start to slip.
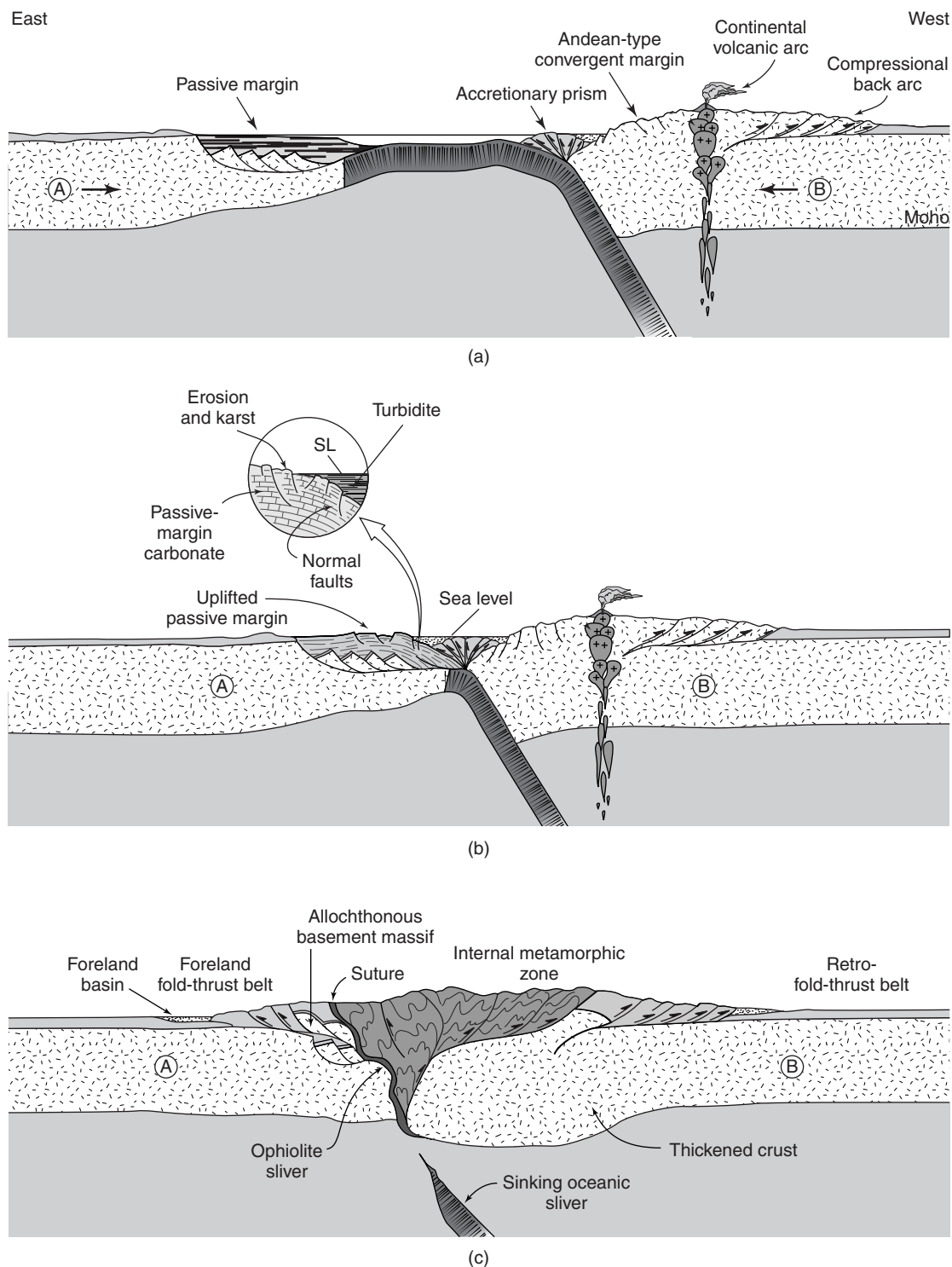
FIGURE 17.21 Stages in an idealized continent–continent collision. (a) Precollision configuration. Continent A has a passive-margin basin on its east coast, while Continent B has a convergent margin on its west coast. (b) During the initial stage of collision, the passive margin is uplifted, and an unconformity (locally, with karst) develops. Turbidites derived from Continent B soon bury this unconformity (see inset). Normal faults break up the strata of the passive-margin basin, due to stretching. But soon, thrusts begin to develop, transporting the deeper parts of the basin over the shallower parts. (c) In a mature collision orogen, the subducting slab has broken off, a suture has formed, and metamorphic rocks are uplifted and exhumed in the interior of the orogen.
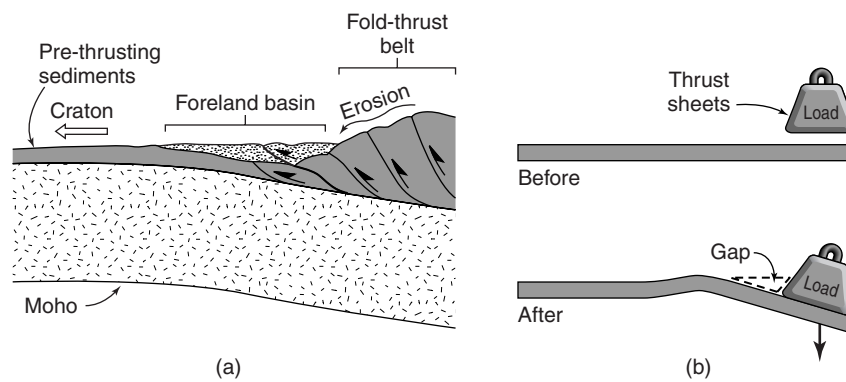
FIGURE 17.22   Origin of a foreland basin. (a) A schematic cross section showing how a stack of thrust slices, when emplaced on the edge of a continent, loads the continent. As a result, a depression develops on the continent. This depression fills with sediment eroded from the orogenic highlands and becomes the foreland basin. (b) You can envision the process of foreland-basin formation by imagining a load placed on the edge of a sheet. If the sheet has flexural rigidity, as does lithosphere, the load pushes the edge of the sheet down and creates a gap.

17.3.2 Stage 2: Abortive Subduction and Suturing

With continued convergence, the surface of Continent A's continental shelf becomes the floor of the trench (Figure 17.21b). When this happens, turbidites derived from the margin of Continent B and its volcanic arc bury the now-eroded surface of the shelf.[6] Thus, a major unconformity defines the contact between the passive-margin basin sedimentary sequence and the turbidites.

Prior to collision, formation of the accretionary prism progressed along the margin of Continent B as new thrusts cut seaward into the strata on the down-going oceanic plate. But during collision, the strata on the downgoing plate consist of thick, well-stratified sedimentary beds of the former passive-margin basin. Thus, thrusts propagate into these beds, producing a fold-thrust belt that, with time, grows toward the fore-land (i.e., in the direction of the continental interior) of Continent A. The stack of thrust slices acts as a load that depresses the surface of Continent A, yielding a **foreland sedimentary basin** that spreads out over the edge of Continent A's craton (Figure 17.22a and b).
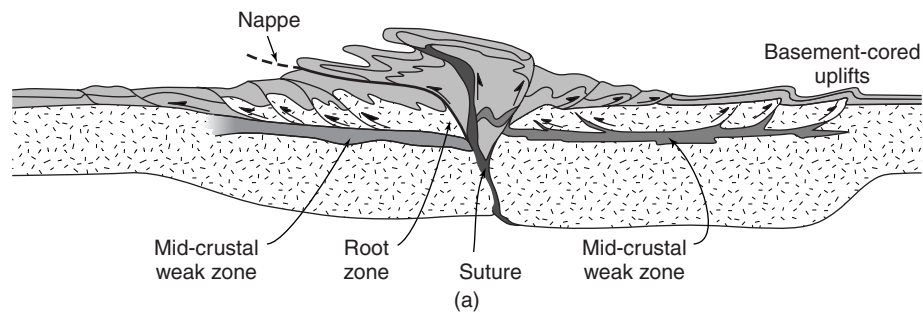
Such basins are asymmetric—they are thickest along the margin of the orogen and become thinner toward the interior of the continent. Meanwhile the backarc fold-thrust belt on Continent B continues to be active.

Shortening during the collision also reactivates the normal faults that bound basement slices at the base of the passive margin; because reactivation occurs in response to compression, these faults now move as reverse faults. This new movement emplaces slices of basement closest to the hinterland part of the fold-thrust belt over strata of the former passive margin (Figure 17.21c). The overall process of transforming the passive-margin basin into a thrust belt is called **basin inversion.** We use the term to emphasize that, during this process, a region that had previously undergone extension and subsidence during basin formation, now telescopes back together by a reversal of shear sense on preexisting faults, and undergoes uplift.

Eventually, a slice of the oceanic lithosphere that had once separated Continent A from Continent B may thrust over the inverted passive-margin of Continent A. This slice, which appears in the orogen as a band of highly sheared mafic and ultramafic rock, defines the **suture;** rock on one side of the suture was once part of Continent A, while rock on the other side was once part of Continent B.

Meanwhile, in the internal part of the orogen, or its "hinterland," the crust thickens considerably, and ductile folding (creating large, tight to isoclinal folds), shearing (creating mylonites), and regional meta-morphism (creating schists and gneisses) occur at

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[6]In older literature, this sequence of turbidites is called **flysch,** which was defined as "synorogenic" strata. Deposition of flysch was thought to signal the onset of orogenic activity (see also Chapter 20). However, turbidites form in other settings as well, so "flysch" as a tectonic term is confusing and we discourage its use in that way.
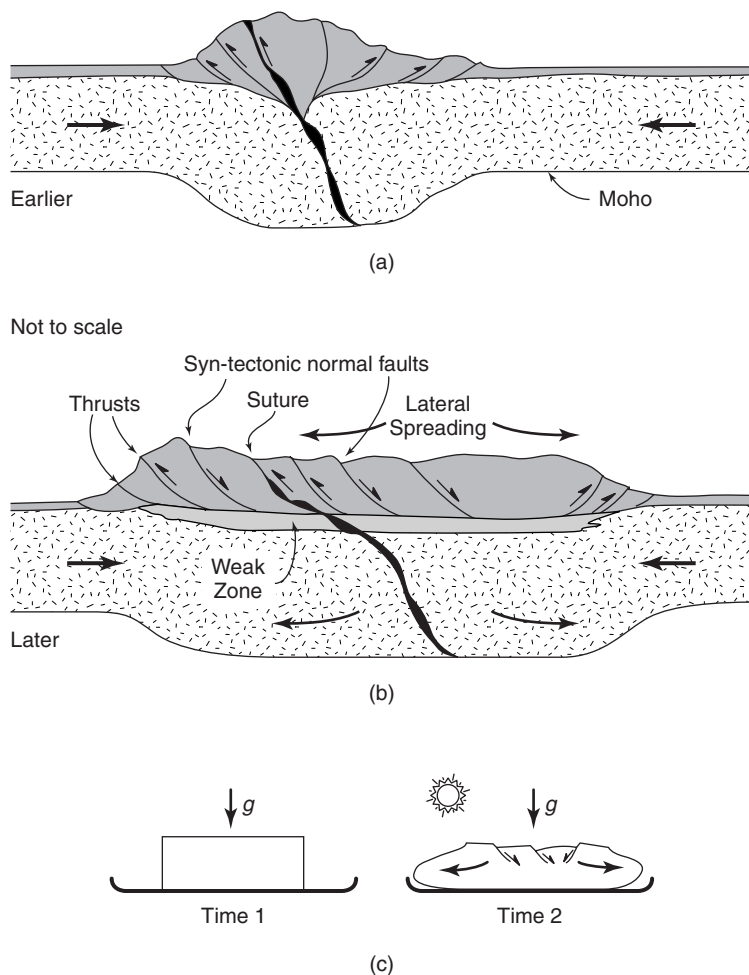
(a)

(b)

FIGURE 17.23   (a) An alternative cross section of a collisional orogen. Not all orogens are the same, and there are different ways to depict their geometry. (This version is modeled on the Alps.) Here, we do not show the erosional land surface, and we add features not shown in Figure 17.21. A mid-crustal weak zone serves as a basal detachment for faulting in the upper crust. Large recumbent nappes develop in the metamorphic interior of the orogen. (b) Photograph of a large nappe, transported on a subhorizontal thrust fault. This example is the Glarner Thrust of the Swiss Alps, placing Permian clastics over Tertiary flysch.

depth. With progressive deformation, the plastically deformed metamorphic rocks move upwards and toward the foreland. In some cases large recumbent folds develop. In the European literature, large sheets of such transported rock, locally containing recumbent folds, are called **nappes** (Figure 17.23a, b; see also Figure 18.15). Metamorphic rock of the hinterland eventually becomes exposed in the peaks of the mountain range due to **exhumation,** the combination of processes that strips off rock at the surface of the Earth to expose rock that had been deeper.

Eventually, the downgoing oceanic lithosphere breaks off the edge of Continent A and sinks slowly into the depths of the mantle. Without a source of new magma, the convergent-margin volcanic arc of Continent B shuts off. On Continent B, deformation styles are the same, but the vergence of structures is opposite to those that form on the edge of Continent A; rocks on the Continent B side of the orogen thrust toward the interior of Continent B. Thus, taken as a whole, the orogen is **bivergent,** meaning that opposite sides of the orogen, overall, verge in opposite directions.

Earlier   Moho

(a)

Not to scale

Syn-tectonic normal faults

Thrusts   Suture   Lateral
                             Spreading

Weak
Zone

Later

(b)

↓g   ☀ ↓g

Time 1   Time 2

(c)

FIGURE 17.24 The concept of orogenic collapse. (a) A schematic cross section shows that during an early stage in a collision, the crust thickens by thrusting. (b) Later, as collapse occurs, extensional faults develop in the upper crust, while plastic flow occurs at depth. This process may contribute to development of a broad plateau. (c) The soft-cheese analogy for extensional collapse. A block of cold cheese can maintain its thickness. If the cheese warms up in the sun, it loses strength and spreads laterally. The rind of the cheese ruptures, and small faults develop.

### 17.3.3 Stage 3: Crustal Thickening and Extensional Collapse

So far, we've focused on the horizontal shortening that takes place in the crust during collisional tectonics. But keep in mind that as crust shortens horizontally, it also thickens (Figure 17.24a). In fact, the crust beneath collisional orogens may attain a thickness of 60–70 km, almost twice the thickness of normal crust. Shortening during collision also causes the lithospheric mantle to thicken substantially as well.

Thickening of the crust cannot continue indefinitely because, as the crust thickens, rock at depth becomes warm and, therefore, weaker. As a consequence, the differential stress developed in the orogen due to the weight of overlying rock (the "overburden") exceeds the yield strength of the rock at depth, and the rock begins to flow and develop horizontal extensional strain (Figure 17.24b). In other words, because of the force of gravity, very thick orogens collapse under their own weight. As we pointed out in Chapter 16, you can picture this process by imagining a block of cheese

that is heated in the sun (Figure 17.24c). Eventually, the cheese gets so soft that it spreads out, and the thickness of the block diminishes. This process is called **extensional collapse** (or, **orogenic collapse**).

Ductile extensional collapse at depth in an orogen causes stretching of the upper crust, where rock is cooler and still brittle. Therefore, during collapse, rock of the upper crust ruptures and normal faults form. Because collapse decreases the thickness of the uppermost crust, it causes decompression of the lower crust. This decompression may trigger partial melting of the deep crust, or even the underlying asthenosphere, producing magmas. Melting may also be caused by **lithosphere delamination.** This means that the deep keel of lithosphere that develops during thickening drops off (Figure 17.25). Warm asthenosphere rises to take its place and heats the remaining lithosphere. Some researchers argue that a broad portion of the mid-crust may actually become partially molten at this stage. What is certain is that magma does form and intrude the upper crust in many collisional orogens after

(a)

(b)
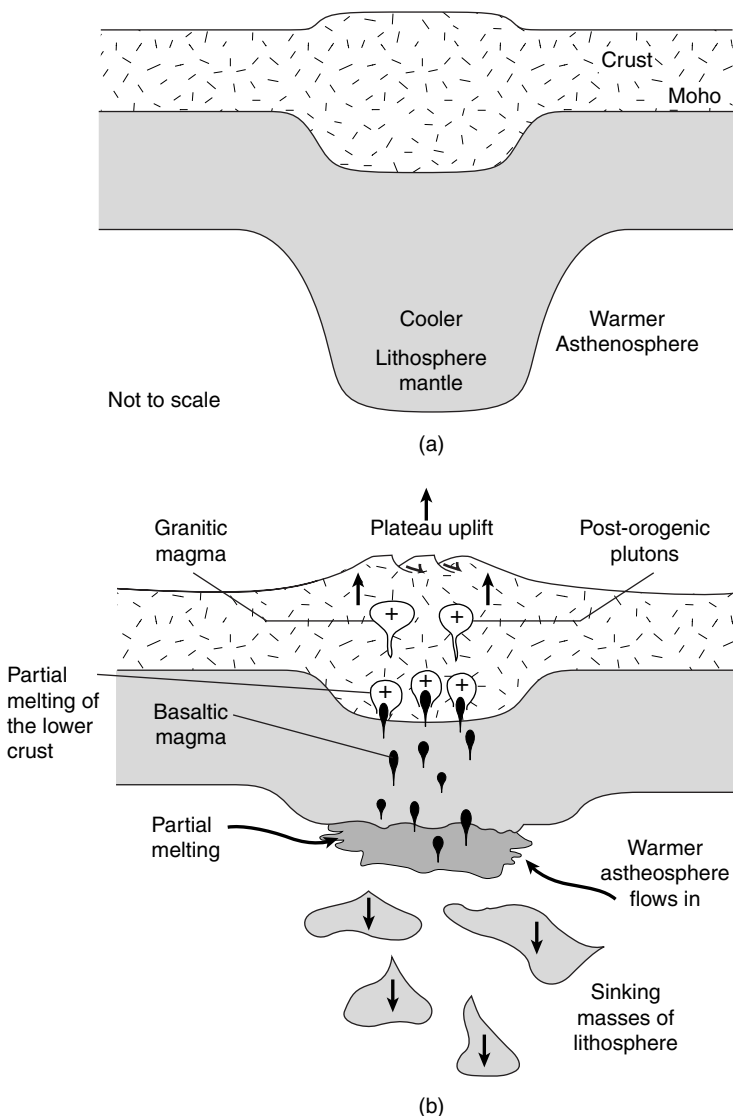
FIGURE 17.25 Postorogenic plutons and delamination.
(a) Thickening of lithosphere forms a keel-shaped mass of cool
lithosphere to protrude down into the asthenosphere. (b) The keel drops
off and is replaced by warm asthenosphere, causing partial melting and
formation of anorogenic (postorogenic) plutons. The surface of the crust
may rise as a consequence.

deformation has ceased. Because it intrudes after deformation, the granite has no tectonite fabric, so geologists refer to the granite as **postorogenic granite.**

The process of extensional collapse can occur while shortening and thrusting continue along the margins of collisional orogen, and it may continue after shortening has ceased. Extensional collapse, together with erosion, keeps mountain ranges from exceeding elevations of about 8 to 9 km, and contributes to the development of broad, high plateaus like the Tibetan Plateau of Asia (see Chapter 21).

17.4 OTHER CONSEQUENCES OF COLLISIONAL TECTONICS

In our description of the stages in an idealized continent–continent collision, we focused only on tectonic phenomena occurring adjacent to the colliding margins, and only on movements that can be illustrated by a two-dimensional cross-sectional plane. Here, we describe additional tectonic phenomena that may occur during collision, in specific situations, and consider movements in the third dimension.

17.4.1 Regional Strike-Slip Faulting and Lateral Escape

Collisional orogeny along a continental margin may lead to the development of regional-scale strike-slip faults that propagate far into the interior of the overriding plate. As an example, several large strike-slip faults start at the Himalayas and cut eastward, following curved trajectories across Asia to the Pacific margin (Figure 17.26a). Researchers suggest that some of the movement on these faults accommodates translation of large wedges of Asia relatively eastward, in response to the collision of India with Asia.

To simulate the development of these faults, researchers constructed models in which they push a wooden block into a cake of clay (Figure 17.26b). The wooden block represents the rigid craton of India, while the clay represents the relatively soft crust of Asia. (The crust of Asia is relatively soft because it formed during a protracted Phanerozoic orogeny; see Essay 21.2). In this model, a rigid wall, representing the mass of western Asia, constrains the left side of the clay cake. The right side, representing the Pacific margin, remains unconstrained. As the wooden block moves into the clay, wedges of clay, bounded by strike-slip faults, move laterally to the right. The pattern of faults resembles the trajectories of maximum shear stress predicted by the theory of elasticity; engineers refer to this pattern as a **slip-line field** (Figure 17.26c). So, based on the clay-cake experiments, geologists have suggested that the pattern of strike-slip faults in Asia represents a slip-line field resulting from stresses transmitted into the interior of Asia by the India–Asia

(a)

(b)
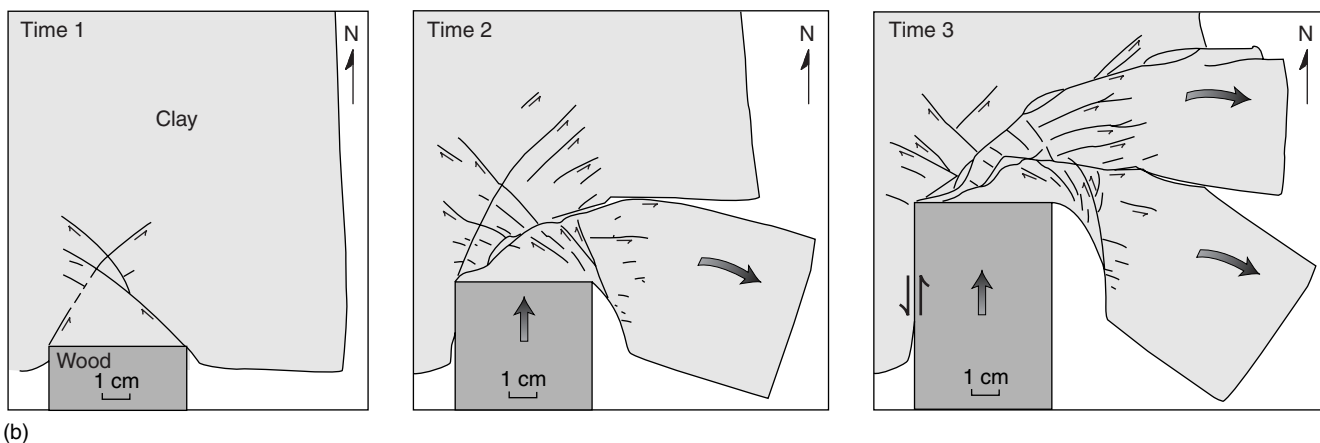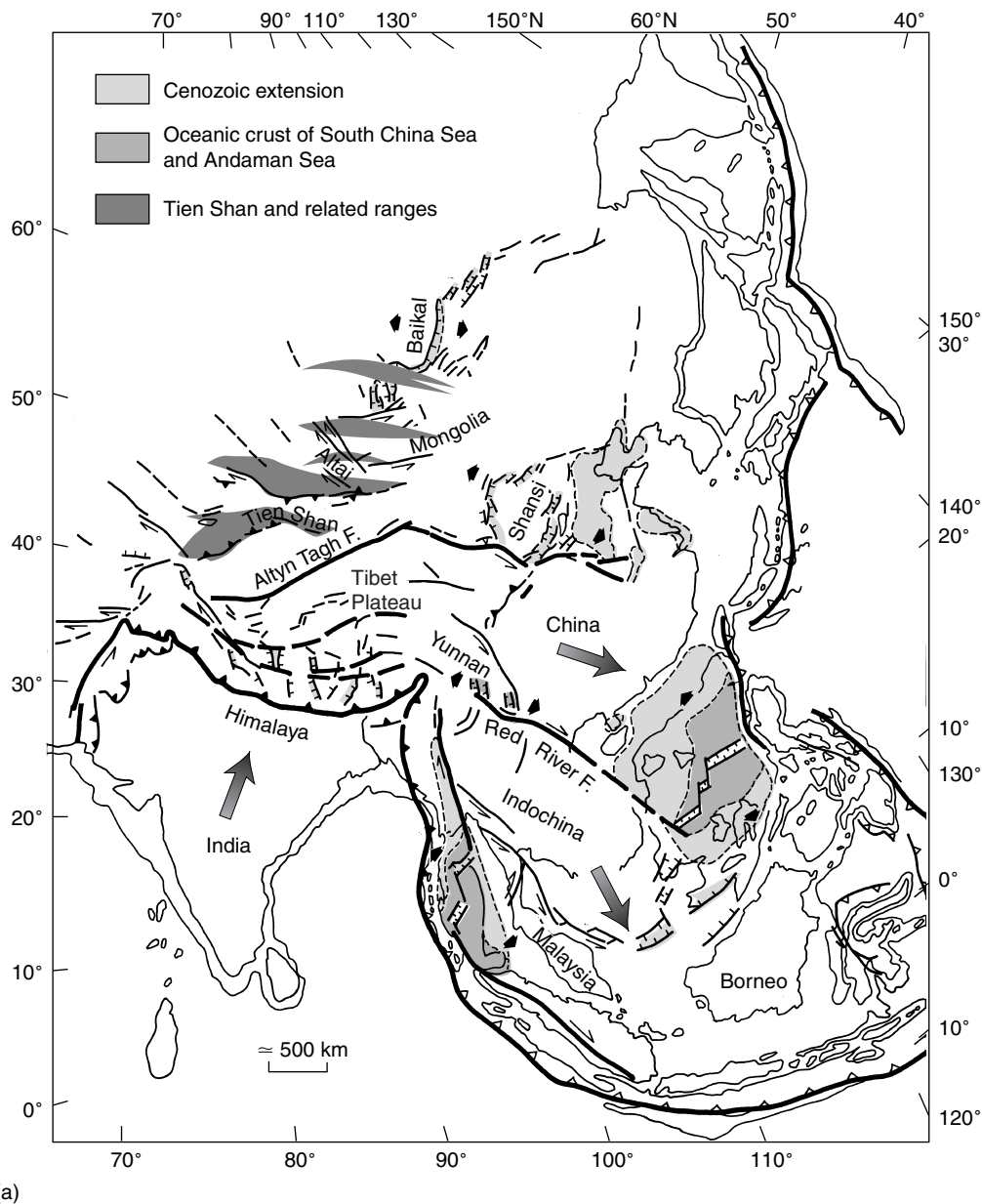
FIGURE 17.26 Lateral escape. (a) A sketch map of major structures in southeastern Asia. Note the major faults that slice across China. The large arrows indicate the motion of large crustal blocks. (b) Map-view sketch of a laboratory experiment to simulate lateral-escape tectonics. A wooden block (representing the Indian craton) is pushed northwards into a clay cake. The cake is restrained on the west, but not on the east. As the block indents, strike-slip faults develop in the clay, and large slices are squeezed eastwards. (c) Map view showing a theoretical model of a slip-line field caused by indentation of a rigid block into an elastic sheet. The lines represent the trajectories of maximum shear stress.
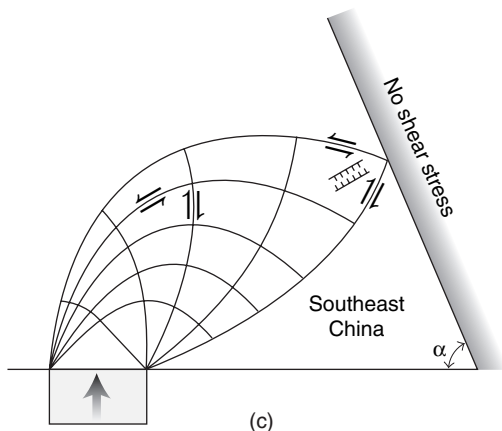
No shear stress

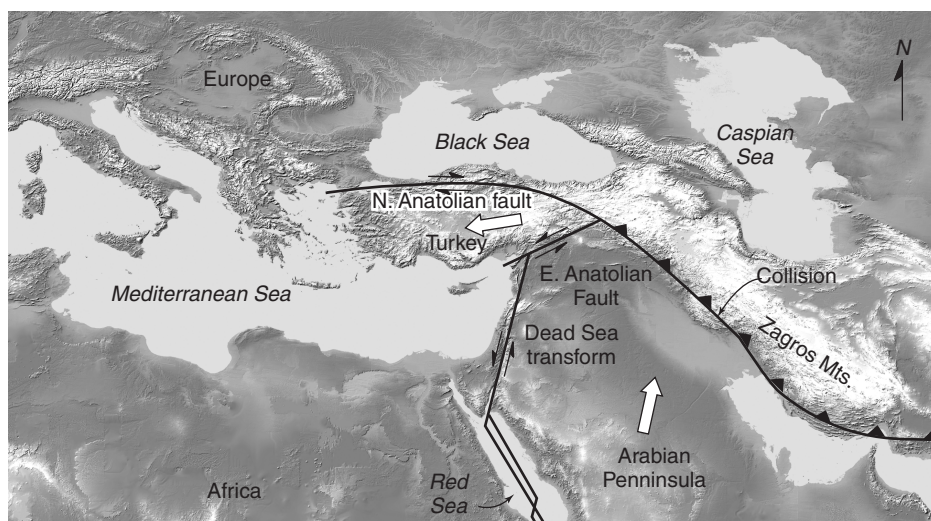Southeast China

(c)

FIGURE 17.26 (Continued)

FIGURE 17.27 Sketch map of the eastern Mediterranean region, showing how Turkey is being squeezed to the west as the Arabian Peninsula moves northwards. This lateral escape of Turkey is accommodated by slip on the North Anatolian and East Anatolian Faults.

sideways when you squeeze it between your fingers. This tectonic process is called **lateral escape.** Note that where lateral escape occurs, all strain resulting from collision cannot be depicted in a cross section, because of movement into or out of the plane of the section.

17.4.2 Plateau Uplift

Continental collisions cause the uplift of a linear belt of mountains, a "collisional range." But in some examples, collision also leads to the development of a broad region of uplifted land known as a **plateau.** As a case in point, consider the development of the Tibetan Plateau in southern Asia (Figure 17.26). Determining the time of uplift of this plateau has proven to be a challenging problem, but available evidence suggests that it rose up in concert with India's collision with Asia.

Researchers continue to debate the reason for the uplift of the Tibetan Plateau. Some consider it to be a consequence of thickening of the crust, either by plastic flow in the lower crust, or because the crust of India has been emplaced under the crust of Asia. Others suggest that it reflects heating that occurred when the lower part of the lithospheric mantle detached (delaminated) from the base of the lithosphere and sank down into the mantle (Figure 17.25). Such **lithosphere delamination** would juxtapose hot asthenosphere at the base of the remaining lithosphere and cause the lithosphere to heat up. Thus, to maintain isostatic equilibrium, the surface of the lithosphere would have to rise.

17.4.3 Continental Interior Fault-and-Fold Zones

Look again at a Figure 17.26a. An *en echelon* band of short, but high thrust-bounded mountain belts, starting with the Tien Shan, tracks across the center of the continent. This band appears to be a zone of fault reactivation caused by stresses imparted to Asia by the collision of India. Stresses that occur in a continental

collision. Substantial debate continues about the significance and age of faulting in southern Asia, and thus the applicability of slip-line field theory to these faults remains uncertain. Possibly, the strike-slip faults are reactivated preexisting faults; the specific faults that reactivated were those that had the correct orientation to be part of the slip-line field.

A map of Turkey provides another example of strike-slip faulting resulting from a collision. Here, the northward movement of the Arabian plate has caught Turkey in a vise. The resulting stresses squeeze Turkey westward into the Mediterranean as the Saudi Arabian Plate moves north. This motion is accommodated by strike-slip displacement on along the North and East Anatolian Faults (Figure 17.27). This example illustrates how blocks of the overriding plate in a collisional orogen can be squeezed sideways out of the path of an indenter, much like a watermelon seed squirts

interior as a result of tectonism along the continental margin are called **far-field stresses.** As India moves into Asia, it displaces the crust of southern Asia with respect to that of northern Asia. This displacement is accommodated by transpressional motion on faults in central Asia that are thousands of kilometers from the collision zone.
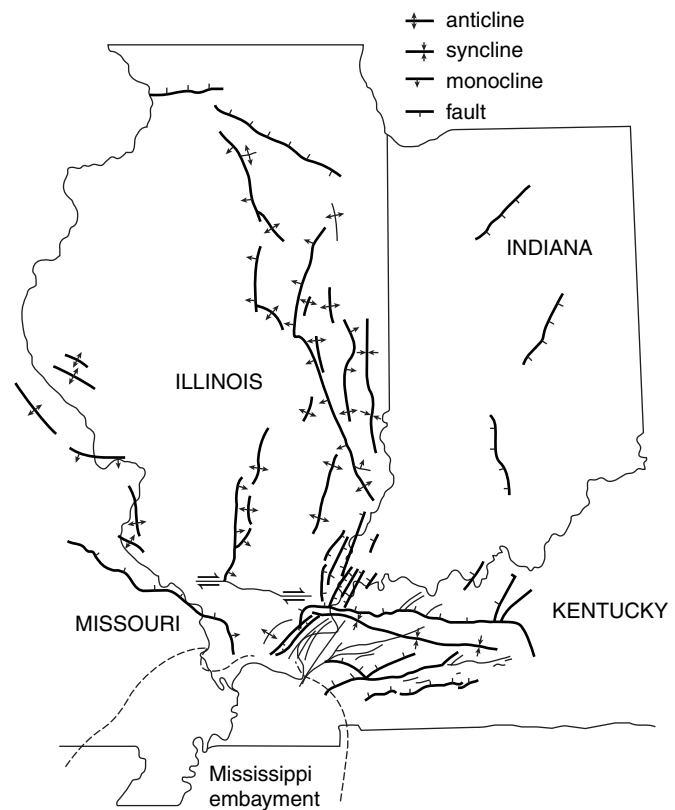
Similarly, the collision of Africa with North America during the late Paleozoic Alleghanian orogeny generated far-field stresses in the continental interior of North America. These stresses were sufficient to cause transpressional or transtensional movements on basement-penetrating faults of the region (Figure 17.28a). The reverse-sense component of motion on these faults led to the formation of monoclinal folds that drape platform strata over the uplifted blocks (see Essay 22.7). Sediments shed from the uplifts filled narrow, but deep, sedimentary basins. Exposures of the faults and associated structures (e.g., monoclinal folds, narrow sedimentary basins) can be found throughout the present-day Rocky Mountain region of the North American Cordillera in the western United States.

Geologists refer to the region in the western United States in which these Late Paleozoic faults, folds, uplifts, and basins formed as the **Ancestral Rockies,** because it occurs in the region now occupied by the present-day Rocky Mountains (a Cenozoic feature) but existed long before the present mountains formed (Figure 17.28a). Notably, structures very similar to those of the traditional Ancestral Rockies lie beneath the corn and wheat fields of the Midcontinent Region—these structures are called **Midcontinent fault-and-fold zones** (Figure 17.28b). Though they are not as well exposed as those of the west, and typically do not display as much displacement, they probably formed in much the same way, in that their movement probably occurred in response to far-field stresses generated during Paleozoic orogeny along the eastern and southern margins of North America. Movement on these faults displaced stratigraphic markers at depth and led to the development of monoclinal folds at shallower crustal levels.[7]

(a)

(b)

FIGURE 17.28 Continental-interior deformation of the United States. (a) Map of the Ancestral Rockies, and equivalent Midcontinent fault-and-fold zones. (b) Midcontinent fault-and-fold zones within the interior of the Illinois basin. These were active during the Alleghanian Orogeny (i.e., during the collision of Africa with North America).

[7]In addition to macroscopic folds and faults, the Alleghanian Orogeny also caused calcite twinning, a form of intragranular strain, to develop in the Midcontinent Region (see Figure 5.29). The strain due to this calcite twinning is only on the order of 1–6%.

The change in differential stress in continental interiors, and/or flexural loading (i.e., the emplacement of a load such as a stack of thrust slices) on a continental margin, due to collision may also trigger **epeirogeny,** meaning the gentle vertical displacement of broad regions, in continental interiors. These movements include the uplift of regional domes and arches as recorded by the development of unconformities, and the subsidence of intracratonic basins as recorded by sudden increases in the rate of basin subsidence. We discuss epeirogeny, and the factors that may cause it, in Essay 22.7.

17.4.4 Crustal Accretion (Accretionary Tectonics)

If you look at the present-day Pacific Ocean region, you will find many pieces of crust that differ in thickness and/or composition from the typical oceanic crust produced at a mid-ocean ridge (Figure 17.19). These pieces include:

- Small fragments of continental crust, such as Japan or Borneo, that rifted off larger continents in the past.
- Volcanic island arcs, such as the Mariana Arc, which formed along convergent plate boundaries.
- Seamount chains and oceanic island chains formed above hot spots.
- Oceanic plateaus, broad regions of anomalously thick crust, probably composed of basalts extruded at particularly productive hot spots.

All of these pieces are buoyant, relative to the asthenosphere, and thus cannot be subducted. Therefore, if subduction continues along the eastern margin of Asia for many more millions of years, the pieces would eventually collide with and suture to Asia. After each such **docking event,** a new convergent margin may form on the outboard (oceanic) side of each sutured piece. As a consequence, the continent grows. This overall process is called **crustal accretion** or **accretionary tectonics.**[8] The small crustal pieces that have been attached to a larger continental block by accretion are called **accreted terranes.**[9] In some cases,

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[8]An orogenic event during which a continent grows significantly larger by the addition of broad regions of accreted crust can be called an **accretionary orogeny.**

[9]They may also be called **exotic terranes,** to emphasize that they came from elsewhere (an "exotic" place) before attaching to the larger continent. A block of crust that might be accreted (or exotic) can be referred to as a **suspect terrane,** until its origin has been confirmed. Once geologists prove that a crustal block in an orogen is exotic, then the name "suspect" should no longer be used for it. Note the spelling of "terrane." This differs from the spelling of "terrain," which refers to a topographic surface.

the area of a continent grows by the addition of a broad belt of accreted terranes called a **tectonic collage** (Figure 17.29a and b).

How do we identify accreted terranes in an orogen? The first hint that a block of crust may be an accreted terrane comes from studying the geologic history preserved in a block. Do the rocks and structures of the block correlate with those of adjacent crust? If not, then the block is probably accreted. Geologists can test this proposal further by studying paleomagnetic data, for these may demonstrate that the accreted block and the continent to which it is now attached did not have the same apparent **polar-wander paths** prior to the time at which accretion occurred. Paleontological data may also indicate that the block originated at a different latitude from the host continent.

Based on mapping, paleomagnetic study, and paleontologic study, geologists have demonstrated that, during the Mesozoic, the North American Cordillera grew westward by crustal accretion. In fact, much of the vast tract of land that now comprises most of California, Oregon, Washington, and Alaska in the United States, and British Columbia and the Northwest Territories in Canada originated as crustal fragments outboard of North America that were accreted to the continental margin (Figure 17.29b). Much of this accretion occurred during oblique convergence, so strike-slip faults developed in the orogen and movement on these transported whole accreted terranes, or slivers of them, along strike to the north. In western North America, for example, bits and pieces of an accreted terrane known as Wrangelia crop out from Washington (maybe even Idaho) to Alaska ("W" in Figure 17.29b; see also Section 19.4). The Appalachian Mountains preserve a similar story. Thus, the eastern edge of what had been North America in the Precambrian lies well inboard of the continent's present coastline.

When two major continents collide along a margin that previously was the locus of terrane accretion, you can imagine that the resulting collisional orogen will be very complex. It will contain several sutures separating different blocks, and each block has its own unique geologic history. Most major collisional orogens involve accretion of exotic terranes prior to collision of the larger continents and final closure of the intervening ocean, so such complexity is the norm rather than the exception. Therefore, you should not assume that the history in one particular region represents that of the whole orogen, nor should you be surprised to find radically different geologic histories preserved in adjacent areas of crust in the orogen.

(a)
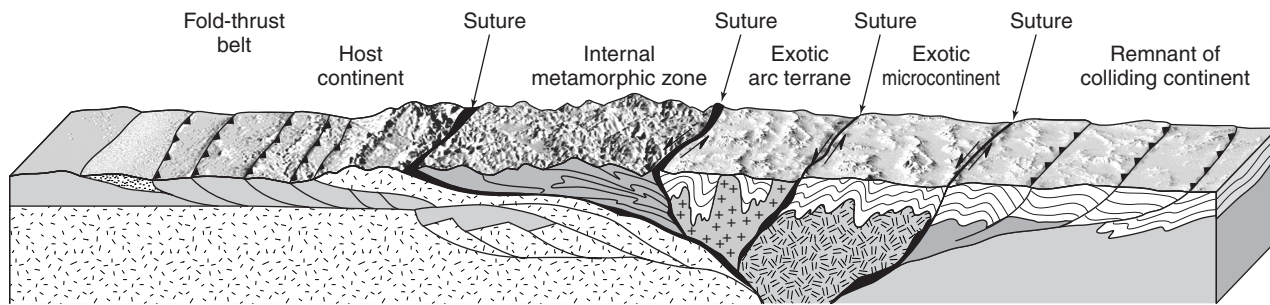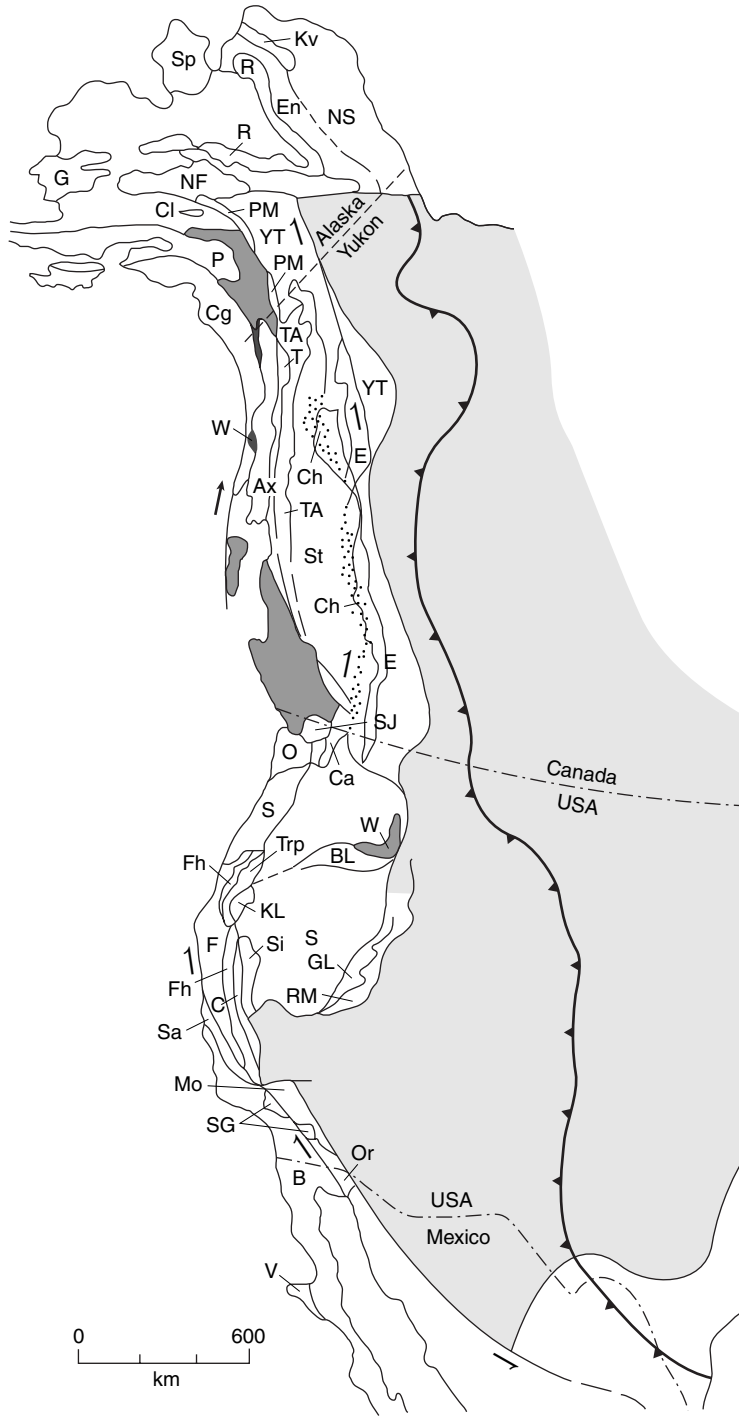
Alaska

NS	North Slope
Kv	Kagvik
En	Endicott
R	Ruby
Sp	Seaward Peninsula
I	Innoko
NF	Nixon Fork
PM	Pingston and McKinley
YT	Yukon-Tanana
Cl	Chuiitna
P	Peninsular
W	Wrangellia
Cg	Chugach and Prince William
TA	Tracy Arm
T	Taku
Ax	Alexander
G	Goodnews

Canada

Ch	Cache Creek
St	Stikine
BR	Bridge River
E	Eastern assemblages

Washington, Oregon, and California

Ca	Northern Cascades
SJ	San Juan
O	Olympic
S	Siletzia
BL	Blue Mountains
Trp	Western Triassic and Paleozoic of Klamath Mountains
KL	Klamath Mountains
Fh	Foothills belt
F	Franciscan and Great Valley
C	Calaveras
Si	Northern Sierra
SG	San Gabriel
Mo	Mohave
Sa	Salinia
Or	Orocopia

Nevada

S	Sonomia
RM	Roberts Mountains
GL	Golconda

Mexico

B	Baja
V	Vizcaino

(b)

FIGURE 17.29 Concept of exotic terranes. (a) Cross section of an orogen, based loosely on the southern Appalachians, showing how the orogen includes an accreted volcanic arc, an accreted microcontinent, and crust that had been part of a large colliding continent. (b) Map of the North American Cordillera, showing the regions that consist of accreted crust. This crust can be divided into numerous exotic terranes.

17.4.5 Deep Structure of Collisional Orogens

The depth in the crust to which faults penetrate during collision is not clear. Some researchers suggest that much of the deformation in a collisional orogen is confined to crust above a **mid-crustal detachment** or weak zone, at a depth of around 20 km (Figure 17.24b). This detachment may form at the lower boundary of discrete faulting—below the detachment, pressure, temperature, and stress conditions allow the crust to behave more like a viscous fluid that deforms independently of the crust above. In some cases, partial melting may occur in the vicinity of the mid-crustal detachment, creating magmas that seep along the detachment and perhaps follow faults to the surface, eventually crystallizing as granites during a late stage in the orogeny.

So far, we have mainly focused our attention on the crust, but it is important to keep in mind that the lithospheric mantle also thickens during collisional orogeny. What happens to this part of the lithosphere remains a subject of debate. Some geologists suggest that the lithospheric mantle separates from the crust during collision and sinks into the deeper mantle. Others suggest that the lithospheric mantle itself shortens and thickens during collision, leading to interesting consequences. For example, if the lithospheric mantle thickens by the same proportion as the crust during collision (i.e., doubles in thickness), then the base of the lithosphere will reach a depth of 200–300 km. Calculations show that the relatively cool rock composing the lithospheric mantle is denser than the rock composing the surrounding asthenosphere and, therefore, is negatively buoyant. Under such conditions, the lower part of the lithospheric mantle beneath a collisional orogen may peel off and sink into the mantle, a process we referred to earlier as **lithospheric delamination** (Figure 17.25). Lithospheric delamination may lead to postorogenic plutonism and to isostatic uplift of the overlying continent.

17.5 INSIGHT FROM MODELING STUDIES

In recent years, researchers have explored the process of convergent and collisional orogeny by means of sandbox models[10] and by computer models. This work allows

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[10]Sand is a good laboratory-scale analog for continental crust. This means that, if continental crust were scaled down to the size of a typical laboratory sandbox (1 m × 1 m), the strength of crustal rock would be comparable to the strength of unconsolidated sand.

(a)

(b)
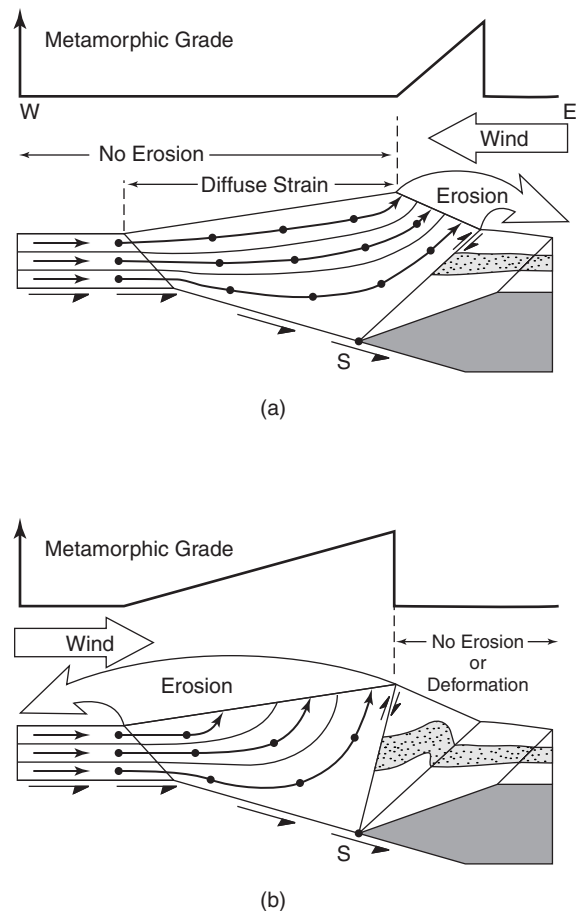
FIGURE 17.30 Results of a computer model study showing the flow path of points in an orogen under different weather conditions. (a) Wind comes from east, so exhumed metamorphic rocks occur in a narrow belt on the east side. (b) If the wind comes from the west, a broad wide metamorphic belt occurs on the west.

researchers to simulate the evolution of orogens in front of their eyes and to determine how a variety of variables affect this evolution. Variables that can be tested include detachment strength, subduction angle, crustal strength, and surface erosion rates. The effect of these variables on the geometry of an orogen can be understood in the context of critical taper theory (see Chapter 18) as applied to doubly vergent orogens.

With critical taper theory in mind, let's consider the effect of changing variables on an orogen. If there is an increase in the erosion rate (due to a change in climate), material transfers from the hinterland to the foreland, thereby decreasing the taper. This causes the orogen to grow wider and causes **exhumation** (erosion and removal of crust at the Earth's surface, thereby allowing the uplift of crust that had been at depth) in the interior of the orogen. The geometry of erosion depends on the prevailing wind direction, for this determines whether erosion attacks the retrowedge or

the forewedge (Figure 17.30). The location of erosion, in turn, determines where exhumation takes place and thus where metamorphic rocks formed at depth eventually become exposed. As another example, if the foreland region contains a particularly weak detachment horizon, then the critical taper angle decreases, and orogen grows wider without becoming as thick. Again, this will affect metamorphic patterns in the interior of the orogen. If the geothermal gradient increases, crust at depth becomes weaker, which in turn decreases the taper angle. This decrease in crustal strength allows extensional collapse to take place so that the wedge achieves the lower taper angle.

17.6   CLOSING REMARKS

In the descriptions provided in Chapters 16 and 17, we've illustrated the start and finish of one loop in the cycle of opening and closing of ocean basins. Geologic mapping around the world emphasizes that patterns of structures created during continental breakup influence patterns of structures developed during collision, and vice versa. Rifts are superimposed on ancient collisional orogens, and in turn, rifts are the likely sites of later collision (so rift basins ultimately get inverted). Because rifts generally control the location of collisional orogens, and vice versa, rocks in a single orogen tend to record the effects of multiple phases of contraction and extension. For example, the eastern United States records a history that involves two principal phases of rifting (Late Precambrian and Middle Mesozoic) and two principal phases of continental collision (Late Precambrian and Paleozoic). Multiple reactivations have kept Phanerozoic collisional orogens weak relative to continental interiors (cratons). In a way, they are like weak scars that never heal, and thus protect cratons from major deformation in the Phanerozoic.

Our discussion of convergent and collisional tectonics is not yet complete. In the next chapter (Chapter 18), we narrow our focus and look into the world of fold-thrust belts, and in the final chapter of this book we will explore the geology of several collisional orogens in detail.

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