21 Convergent Plate Boundaries

The convergence of two tectonic plates develops some of the most remarkable structural and topographic features on our planet. At convergent plate margins, great slabs of oceanic lithosphere slide ponderously into Earth's internal abyss—the deep mantle. As they slowly disappear from the surface, spectacularly deep trenches form graceful arcs on the seafloor. The subducted plates strive to reach mechanical and chemical equilibrium with the mantle, and, in the process, many of Earth's most dramatic landscapes and structures are created. Earthquakes, volcanic arcs, deep-sea trenches, and the continents themselves are the result of converging plates. But perhaps the most fascinating phenomena resulting from plate collision are the great mountain ranges of the world: the Alps, Andes, Rockies, and Himalayas. The internal structure of mountains shows intense folding, thrust faulting, and other features of intense horizontal compression resulting from plate collision. Young orogenic belts are elevated to great

heights and subjected to vigorous erosion by streams and glaciers.

An example is the great Alpine fold and thrust belt of southern Europe. The Tyrolean Alps of northern Italy illustrate the results of deformation when plates collide, in this case the African and Eurasian plates. Thick sequences of sedimentary strata, originally deposited on the seafloor, were deformed into anticlines and synclines and uplifted as much as four kilometers above sea level. These rocks now form the highest peaks. Erosion has removed a large volume of rock so that what we see in the high peaks is only a small fraction of the total volume of rock.

Convergent plate margins are where continental crust is born, just as divergent plate margins are the birthplaces of oceanic crust. This is perhaps the most important fact to remember as you study these important plate boundaries. This new granitic crust is so buoyant that it can never sink into the denser mantle below. Consequently, the rocks of the continents are much older than those in the ocean basins. They preserve a record of much of Earth's ancient history—a record in the form of faults, folds, mountain belts, batholiths, and sediments.

In this chapter, we examine the basic types of convergent boundaries, how they develop, and the rocks they form. We review their intimate relationship to Earth's most destructive earthquakes and volcanoes and to deformation at all scales. In addition, we will see that the deformed rocks, high-grade metamorphic rocks, and igneous intrusions developed at convergent plate margins are the building blocks of the continents.

Courtesy of Dallas and John Heaton/CORBIS

MAJOR CONCEPTS

1. Convergent plate boundaries are zones where lithospheric plates collide. The three major types of convergent plate interactions are (a) convergence of two oceanic plates, (b) convergence of an oceanic and a continental plate, and (c) collision of two continental plates. The first two involve subduction of oceanic lithosphere into the mantle.
2. Plate temperatures, convergence rates, and convergence directions play important roles in determining the final character of a convergent plate boundary.
3. Most subduction zones have an outer swell, a trench and forearc, a magmatic arc, and a backarc basin. In contrast, continental collision produces a wide belt of folded and faulted mountains in the middle of a new continent.
4. Subduction of oceanic lithosphere produces a narrow, inclined zone of earthquakes that extends to more than 600 km depth, but broad belts of shallow earthquakes form where two continents collide.
5. Crustal deformation at subduction zones produces melange in the forearc and extension or compression in the volcanic arc and backarc areas. Continental collision is always marked by strong horizontal compression that causes folding and thrust faulting.
6. Magma is generated at subduction zones because dehydration of oceanic crust causes partial melting of the overlying mantle. Andesite and other silicic magmas that commonly erupt explosively are distinctive products of convergent plate boundaries. At depth, plutons form, composed of rock ranging from diorite to granite. In continental collision zones, magma is less voluminous, dominantly granitic, and probably derived by melting of preexisting continental crust.
7. Metamorphism at subduction zones produces low-temperature–high-pressure facies near the trench and higher-temperature facies near the magmatic arc. Broad belts of highly deformed metamorphic rocks mark the sites of past continental collision.
8. Continents grow larger as low-density silica-rich rock is added to the crust at convergent plate boundaries and by terrane accretion.

TYPES OF CONVERGENT PLATE BOUNDARIES

Three distinctive types of convergence are recognized: (1) the convergence of two oceanic plates, (2) the convergence of a continental plate and an oceanic plate, and (3) the convergence of two continental plates.

We have learned a great deal about convergent boundaries from geophysical studies that include gravity surveys, measurements of heat flow, and seismic-reflection profiles. In addition, geochemical studies of the igneous rocks erupted at these boundaries tell the tale of their generation, rise, and differentiation. These results, combined with field studies of ancient and modern mountain belts and arcs, give us an integrated picture of the geologic features of convergent plate margins and the processes that have shaped them.

If you study Figure 21.1, you will observe that **convergent boundaries** involve either the convergence of two oceanic plates (numerous trenches in the Pacific), the convergence of an oceanic and a continental plate (the western margin of South America), or two continental plates (India and Asia).

Convergence of Two Oceanic Plates

The simplest type of convergent plate margin—**ocean-ocean convergence**—consists of two oceanic plates. As the plates collide, one is thrust under the other,

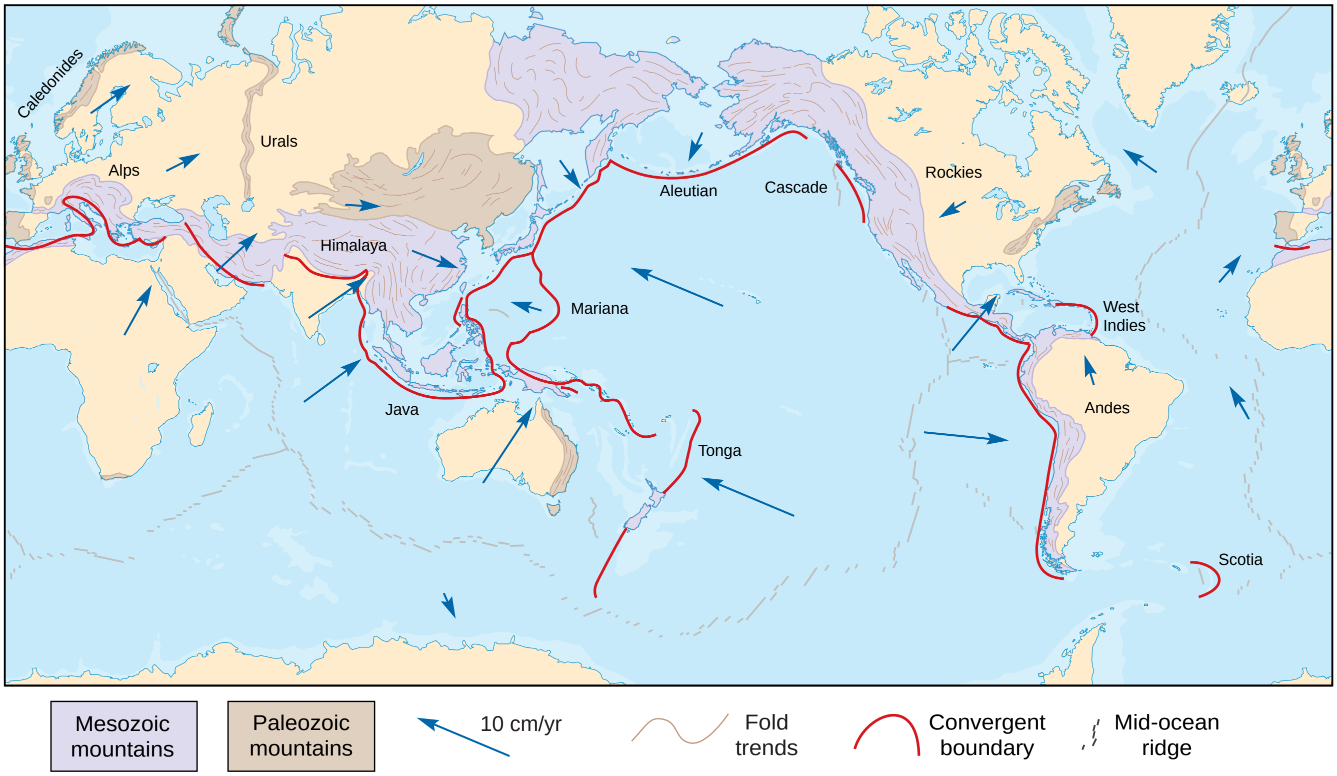


FIGURE 21.1 **Convergent plate margins** are marked in two ways: either by deep trenches, where plates of oceanic lithosphere converge and one descends to be recycled into the mantle, or by high folded mountain belts. In both cases, earthquakes and magma are generated. Absolute plate motions are shown with arrows. Trenches and mountain belts are labeled.

forming a **subduction zone.** The subducting plate descends into the mantle, where it is heated, triggering the generation of magma. The magma, being less dense than the surrounding rock, rises and erupts on the seafloor, ultimately building an arc of volcanic islands (Figure 21.2). Andesite is the volcanic rock that characteristically forms at such sites.

Several important structural and topographic features form at many subduction zones (Figure 21.2). A broad rise or bulge in the downgoing plate, known as an **outer swell,** commonly develops where the plate bends to dive down into the mantle. Closer to the island arc, a deep **trench** and a **forearc ridge** form. The forearc commonly traps sedimentary deposits and is underlain by faulted and highly deformed sedimentary and metamorphic rock. Behind the **volcanic arc,** the **backarc** is a broad region of variable character that may be compressed or extended.

What are the typical features produced by ocean-ocean convergence?

The Tonga Islands in the western Pacific show the structure and topography of a simple island arc (Figure 21.1). The volcanoes are dominated by the eruption of andesite, and the backarc region is extending to form a basin.

Convergence of Oceanic and Continental Plates

A subduction zone also develops where oceanic and continental plates converge (Figure 21.3). The less-dense continental crust always resists subduction into the dense mantle and overrides the oceanic plate. Consequently, the volcanic arc forms on the continent, and compression may deform the continental margin into a **folded mountain belt.** Moreover, the deep mountain roots are intruded by granitic plutons and metamorphosed. In this setting, a trench, a deformed forearc, and a backarc region of deformation are all important.

Ocean-continent convergence has created the Andes Mountains of western South America. The Cascade Range of western North America is another example, forming above an east-dipping subduction zone. An older example is the Rocky Mountain chain of western North America, which was deformed during late

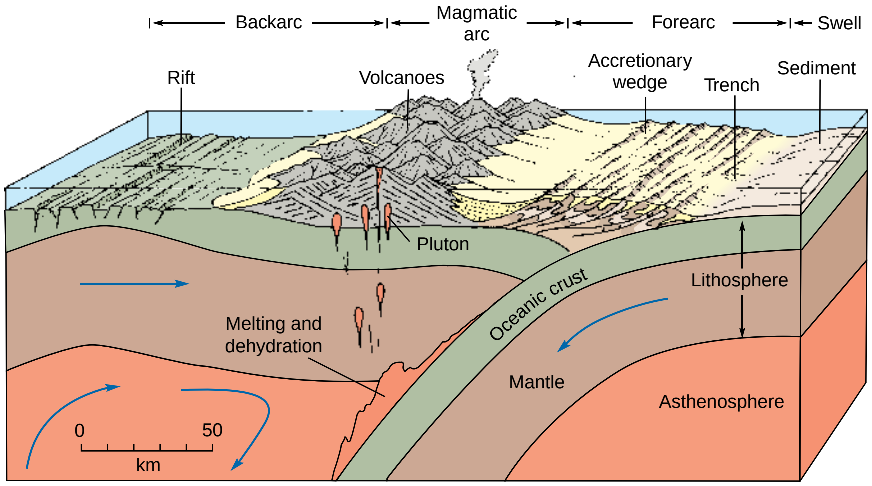


FIGURE 21.2 Ocean-ocean convergence is dominated by volcanic activity and construction of an island arc. Features developed include an outer swell, a forearc, a volcanic arc, and a backarc basin. The forearc is underlain by an accretionary wedge—sediment scraped off the downgoing slab. Widespread metamorphism and large granitic intrusions are rare or absent.

Mesozoic and early Tertiary time (about 150 to 50 million years ago). The Appalachian Mountains in the eastern United States were deformed several times in the Paleozoic Era (about 500 to 300 million years ago). During some part of their histories, all of these mountain chains experienced subduction of oceanic lithosphere beneath a continental margin.

Convergence of Two Continental Plates

If both converging plates contain continental crust, neither is subducted into the mantle, because continental crust is too buoyant (Figure 21.4). In **continent-continent convergence,** one plate overrides the other for a short distance. In sharp contrast to the other two types of convergence, there is no outer swell, deep subduction zone, trench, or forearc wedge. Instead, both continental masses become compressed, and the continents ultimately are "fused" into a single block, with a folded mountain belt marking the line of the suture. **Orogenic metamorphism** and granite generation mark this kind of convergence.

What structures and rock sequences are produced by continent-continent convergence?

Collisions between two continents are occurring in several places. The most dramatic is the one that produced the Himalaya Mountains and Tibetan Plateau of southern Asia. The Himalayas, Earth's highest mountain belt, is a wide and highly deformed zone of mountains that rose as India collided with the Eurasian continent. Russia's Ural Mountains also formed during late Paleozoic time when the Siberian continental mass collided with Europe. The Urals are not tectonically active and are deeply eroded, so they are much lower than the Himalayas (Figure 21.5).

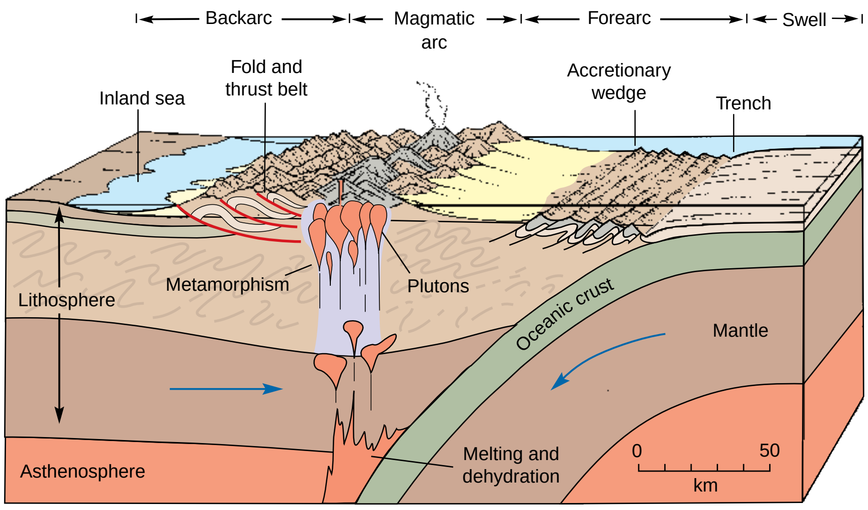


FIGURE 21.3 At ocean-continent convergent plate boundaries, major geologic processes include formation of an accretionary wedge, deformation of the continental margin into a folded mountain belt, metamorphism due to high pressures and high temperatures in the mountain roots, and partial melting of the mantle overlying the descending plate. The resulting magmas commonly differentiate to form andesite and even more silicic magmas, which cool to form plutons. Explosive volcanism is also common. Granitic batholiths and metamorphosed sedimentary rocks develop in the deeper zones of the orogenic belt.

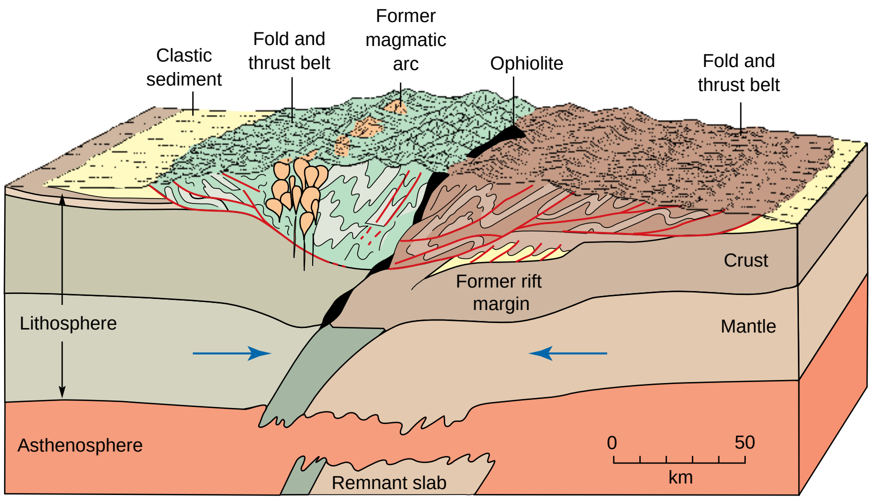


FIGURE 21.4 Continent-continent collision is marked by complete subduction of the oceanic crust. A high mountain belt forms by folding, thrust-faulting, and doubling of the crustal layers as one continent is thrust beneath the other. Ophiolites are thrust into the suture zone. Granite magma and high-grade metamorphic rocks form deep in the mountain belt.

FACTORS INFLUENCING THE NATURE OF CONVERGENT PLATE BOUNDARIES

Plate buoyancy, convergence rates and directions, and the thermal structure of a subduction zone are all important to the development of convergent plate boundaries.

Plate Buoyancy

Many geologic processes at convergent plate boundaries are influenced by density differences that make one plate more buoyant than the other. The most obvious expression of this fact is that subduction occurs because oceanic plates cool and become denser than the underlying mantle. Another important example is the sharp contrast in density between oceanic plates and continental plates. Oceanic crust is composed mostly of basalt (about 3.0 g/cm^3) and is much denser than continental crust (about 2.8 g/cm^3). Thus, at ocean-continent plate boundaries, the oceanic plate descends beneath the continental plate.

(A) The young Himalaya mountain chain formed as a result of the ongoing collision of India and Eurasia. It is Earth's highest range, with some peaks more than 7000 m above sea level, but deep valleys have been cut by river and glacier erosion. *(Courtesy of Paolo Koch/Photo Researchers, Inc.)*

(B) The Ural Mountains formed in the late Paleozoic (about 350 million years ago) when Europe collided with Asia. The mountains have been deeply eroded so that no peaks are higher than 2000 m, but the internal structure reveals its origin. *(Courtesy of Wolfgang Kaehler)*

FIGURE 21.5 High folded mountain belts are formed during the collision of two continents but are gradually eroded away, as seen in this comparison of the Himalayas and Urals.

Less obvious, but nevertheless important, are density differences caused by differences in the thickness of the crust. For example, seamounts and thick plateaus of basaltic lava that erupted on the seafloor can make the lithosphere slightly more buoyant than sections of oceanic lithosphere that lack this thick crust and are made mostly of dense mantle instead. Consequently, oceanic lithosphere with thick crust resists subduction and may bend into the mantle at a relatively low angle. Moreover, as a seamount chain approaches a subduction zone, it may clog up the subduction process or become scraped off the oceanic lithosphere and accreted to an island arc or continental margin.

Temperature also affects the buoyancy of the lithosphere. We have already seen how oceanic lithosphere gets denser as it moves away from the midoceanic ridge and cools. Imagine what would happen if a continental subduction zone developed immediately next to an oceanic ridge. The hot young lithosphere would be only slightly less buoyant than the old cold continental lithosphere. Consequently, it would not subduct readily and the hot slab might dip into the mantle at a low angle.

The Thermal Structure of Subduction Zones

The physical and chemical behavior of most materials are profoundly affected by temperature. There are many familiar examples. Warm honey flows much more readily than refrigerated honey. Snow a few degrees below freezing is crisp and brittle, but near its melting point, it is slushy and flows. A similar situation occurs in lithospheric plates. The temperature variations at convergent plate margins exert strong controls on rock dynamics.

The Cold Slab. The most obvious feature of the thermal structure of a subduction zone (Figure 21.6) is the deep penetration of the cold subducting plate into the hot asthenosphere. Rocks are very good insulators, and heat diffuses very slowly through them. Consequently, subducted lithosphere heats very slowly as it moves down through the hot mantle. As a result, temperatures as low as 400°C may be found in the plate at a depth of 150 km. This is a strikingly anomalous situation; in an area with a normal temperature gradient, the temperature would be as high as 1200°C at this depth. Even at a depth of more than 600 km, the central zone of

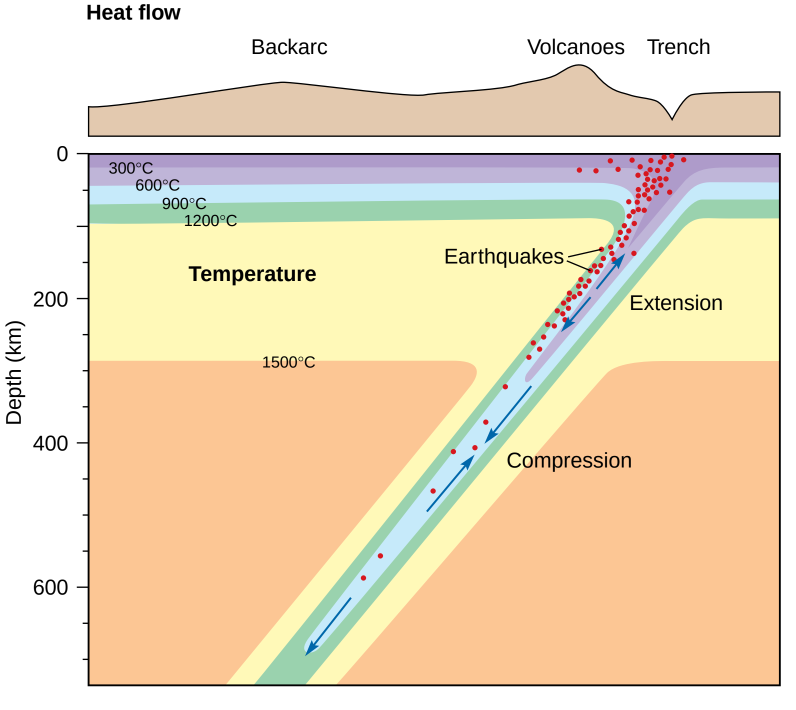


FIGURE 21.6 The thermal structure of a subduction zone is dominated by the underthrusting of a thick, cold slab of oceanic lithosphere into the hot mantle. The descending lithosphere remains cold compared with the surrounding rock to considerable depth. Gradually, the slab heats as it dives deeper and deeper into the hot mantle. The heat flow from the volcanic arc is higher than adjacent regions because heat is carried upward by magma and perhaps by convection in the wedge-shaped area above the slab. Note that earthquakes (dots) occur only in the cold, brittle parts of the slab. *(After R. Stein and C. A. Stein)*

the subducted plate is still as much as 600°C cooler than the surrounding mantle. Because rocks are such poor conductors of heat, it would take an estimated 12 million years for the plate to reach even this relatively low temperature.

Why are a cold slab and a hot arc produced at convergent margins?

The fact that the subducting lithosphere is so much colder than the hot asthenosphere through which it moves explains a great deal about the slab's behavior. The cold slab is much more brittle, stronger, and resistant to ductile flow. It resists mixing with the rest of the mantle and continues to move downward as a discrete plate for hundreds of kilometers. Recent seismic investigations show that the cool temperature of a subducting slab persists all the way to the bottom of the mantle—a distance of about 2700 km and a time of more than 100 million years—before it completely warms up (Figure 21.7).

The Hot Arc. A second important feature of the thermal structure of a convergent plate boundary is the elevation of the heat flow in the volcanic arc (Figure 21.6). The high heat flow in this region reflects the large amount of heat transported from the mantle by magmas generated in the subduction zone.

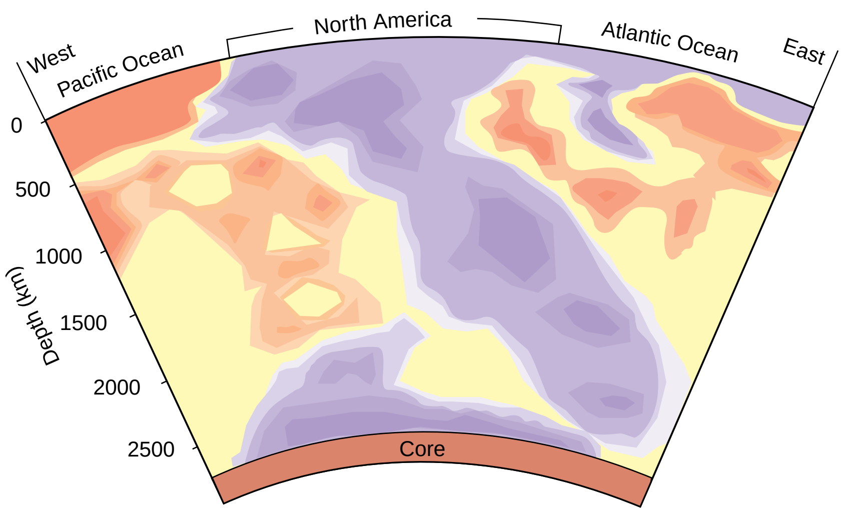
In addition to this magmatic heat transport, flow in the asthenosphere above the subducting plate may enhance heat flow. Because the subducting plate drags the overlying asthenosphere downward, hotter asthenosphere from a greater depth must flow upward to replace it, as shown by the arrows. This convective movement delivers extra heat to the region immediately above the descending plate.

Plate Motions: Directions and Velocities

As you might expect, the velocity and direction of plate motion play important roles in the dynamics of convergent plate margins (Figure 21.1). Plates moving directly toward each other (such as the Pacific and Eurasian plates) converge with high energy and develop long, continuous subduction zones, intense compressive deformation, and vigorous igneous activity.

Plates converging at oblique angles (such as the North American and the Pacific plates) tend to slide past each other and have a strong shearing component. They develop short discontinuous convergent boundaries interspersed with long transform faults. In fact, many convergence zones are like this and have components of extension or transform movement along the plate boundary.

Extremely rapid plate motion develops extensive geologic activity, such as the "Ring of Fire" around the fast-moving Pacific plates. Likewise, the Australian-Indian plate is moving at a high rate northward into the nearly fixed Eurasian



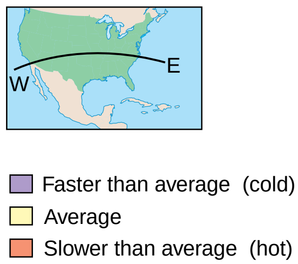


FIGURE 21.7 The trace of a cold subducted slab appears to extend from the continental margin all the way to the core-mantle boundary in this seismic tomograph of Earth's interior beneath North America. Blue represents the cold parts of the mantle, which have high seismic wave velocities; red represents the warmest parts of the mantle, which have lower seismic velocities. *(Modified from S. P. Grand and R. D. van der Hilst)*

continent, explaining the high Himalaya range. Plate velocities are also important for the angle of subduction. Rapidly moving plates generally subduct at lower angles than slower plates.

SEISMICITY AT CONVERGENT PLATE BOUNDARIES

In subduction zones, earthquakes occur in a zone inclined downward beneath the adjacent island arc or continent. But in continental collision zones, where subduction is minimal, earthquakes are shallow and widely distributed. Many of Earth's most devastating earthquakes occur at convergent plate boundaries.

The most widespread and intense earthquake activity occurs at convergent plate boundaries. Almost 95% of the total energy released by all earthquakes comes from these margins. These belts of seismic activity are obvious on the world seismicity map (Figure 21.8).

Earthquakes and Subduction Zones

A strong concentration of shallow, intermediate, and deep earthquakes coincides with the descending plate in a subduction zone. Earthquakes beneath the Tonga arc in the South Pacific illustrate this point nicely (Figure 21.9). The **inclined seismic zone** plunges into the mantle to a depth of more than 600 km. The angle of the seismic zone is usually between 40° and 60°.

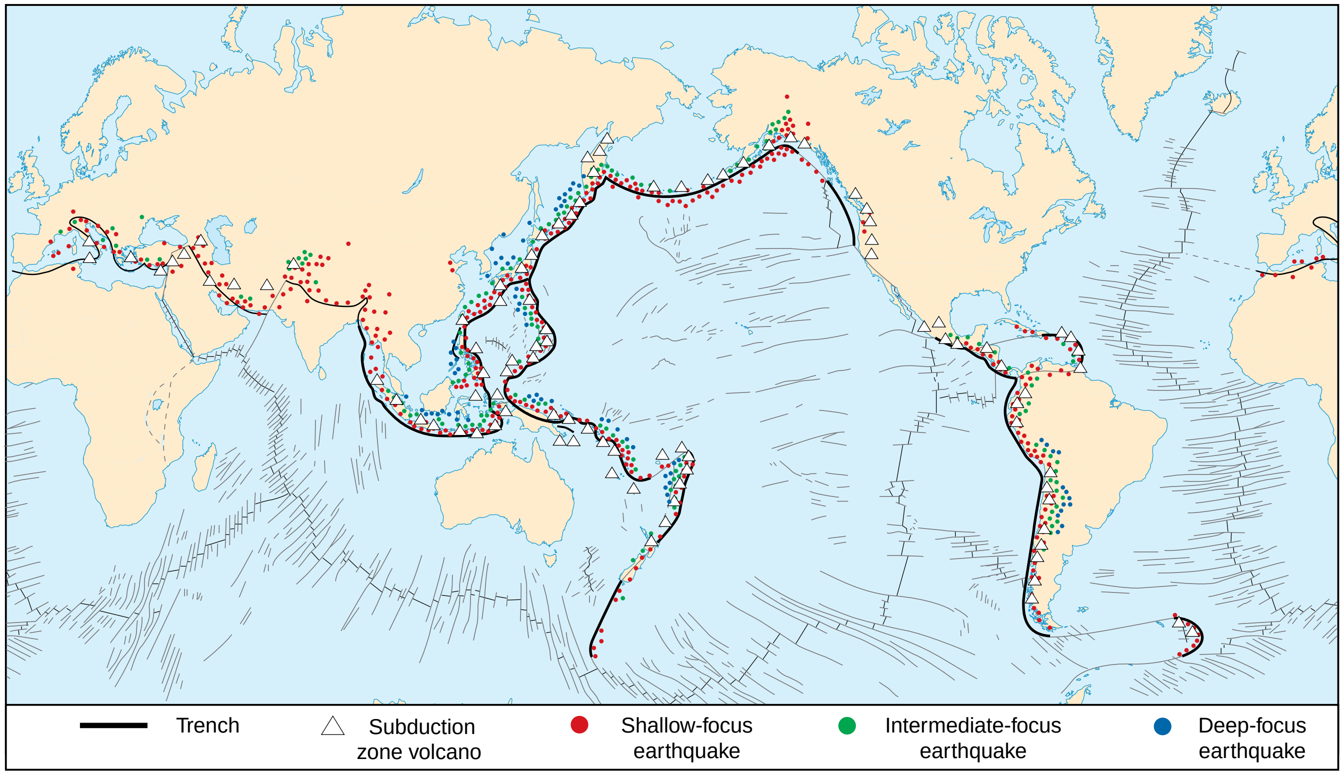


FIGURE 21.8 Earthquakes and volcanoes at convergent plate boundaries are common. Earthquakes occurring here are the most devastating. This map shows the locations of some of the tens of thousands of earthquakes that occurred during a 5-year period. Shallow-focus, intermediate-focus, and deep-focus earthquakes form the inclined zones of earthquakes characteristic of subduction zones. Volcanoes at subduction zones are also the most destructive kind. Subduction zone volcanoes form the "Ring of Fire" around the Pacific Ocean and the arcs of the Mediterranean and Indonesia. Shorter volcanic arcs are found in the Caribbean and South Atlantic. Zones of continental collision, such as the Himalayan region, are quite different from subduction zones. They have abundant, but shallow, earthquakes, and they lack prominent active volcanoes.

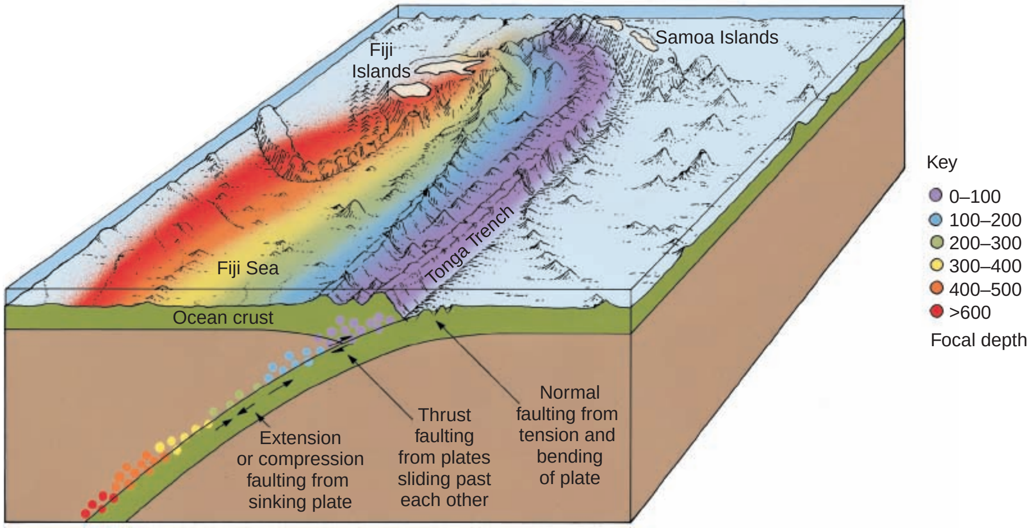


FIGURE 21.9 **Earthquake foci in the Tongan region** in the South Pacific occur in a zone inclined from the Tonga Trench toward the Fiji Islands. The top of the diagram shows the distribution of earthquake epicenters, with focal depths represented by different-colored bands. The cross section on the front of the diagram shows how the seismic zone is inclined from the trench. The colored dots represent different focal depths. This seismic zone accurately marks the boundary of the descending plate in the subduction zone. *(Modified from L. R. Sykes)*

Near the top of the subducting slab, a zone of shallow earthquakes forms where the downgoing slab shears against the overriding plate. Deeper in the subduction zone, earthquakes originate within the slab and not in the surrounding asthenosphere (Figure 21.9). These earthquake zones must correspond to the regions of most intense shearing in the cold, brittle part of the plate.

Why do earthquakes at subduction zones occur so deeply within the mantle? In no other tectonic setting do we detect earthquakes much deeper than 25 km. The answer seems to lie in the thermal structure of a subduction zone (Figure 21.6). Cold slabs of lithosphere plunge down into the mantle in these zones. These "cold" rocks break by brittle fracture when stress exceeds their elastic limits, generating earthquakes. The surrounding warmer mantle also deforms, but it does so by slow ductile flow. In addition, some deep earthquakes may be caused by abrupt metamorphic mineral changes in the subducting slab.

How does an earthquake zone at a convergent margin differ from that of a divergent margin?

Not all earthquakes in subduction zones are generated by simple compression. Studies of seismic waves indicate that the type of faulting varies with depth. For example, near the walls of the trench, normal faulting is typical, resulting from tensional stresses generated by the bending of the plate as it enters the subduction zone (Figure 21.9). In the zone of shallow earthquakes, thrust faulting dominates as the descending lithosphere slides beneath the upper plate. At intermediate depths, extension and normal faulting result when a descending plate that is denser than the surrounding mantle sinks under its own weight. Compression results when the mantle resists the downward motion of the descending plate. In many subduction zones, the deepest earthquakes result from compression, indicating that the mantle material at that depth resists the movement of the descending plate.

Earthquakes and Continental Collision

Earthquakes in continental collision zones are spread out over much broader areas than those involving subduction of oceanic crust and do not form an inclined zone of seismicity. For example, the Himalaya and adjacent Tibetan Plateau are marked by a wide belt of shallow earthquakes that reveals the northward movement of the Indian plate, currently about 0.5 cm/yr. This east-west zone of earthquakes is about 2500 km long and between 1200 and 2000 km wide (Figure 21.8). Moreover, because there is no longer a cold, brittle slab subducting beneath this region, almost all of the earthquakes are relatively shallow. Similarly, the convergent boundary between the Africa-Arabia and the Eurasian plate coincides with a wide seismic belt in the Mediterranean region and across Turkey, Iraq, and Iran (Figure 21.8).

DEFORMATION AT CONVERGENT BOUNDARIES

Intense deformation occurs along convergent plate margins. At subduction zones, melange is produced in the forearc accretionary wedge. In some arc and backarc regions, compression creates folds and thrust faults, but in others extension causes rifting. Collision of two continents is marked by strong horizontal compression that causes folding and thrust faulting.

The single most distinctive feature of the rocks at a convergent plate margin is their structural deformation by folding and faulting. The scale of deformation ranges from small wrinkles in mineral grains or fossils to huge folds and faults tens of kilometers wide that combine to form mountain belts hundreds of kilometers wide and thousands of kilometers long. Understanding this deformation is key to understanding the structure of most continental crust, because convergent plate boundaries are where most continental crust forms.

Compression at Subduction Zones

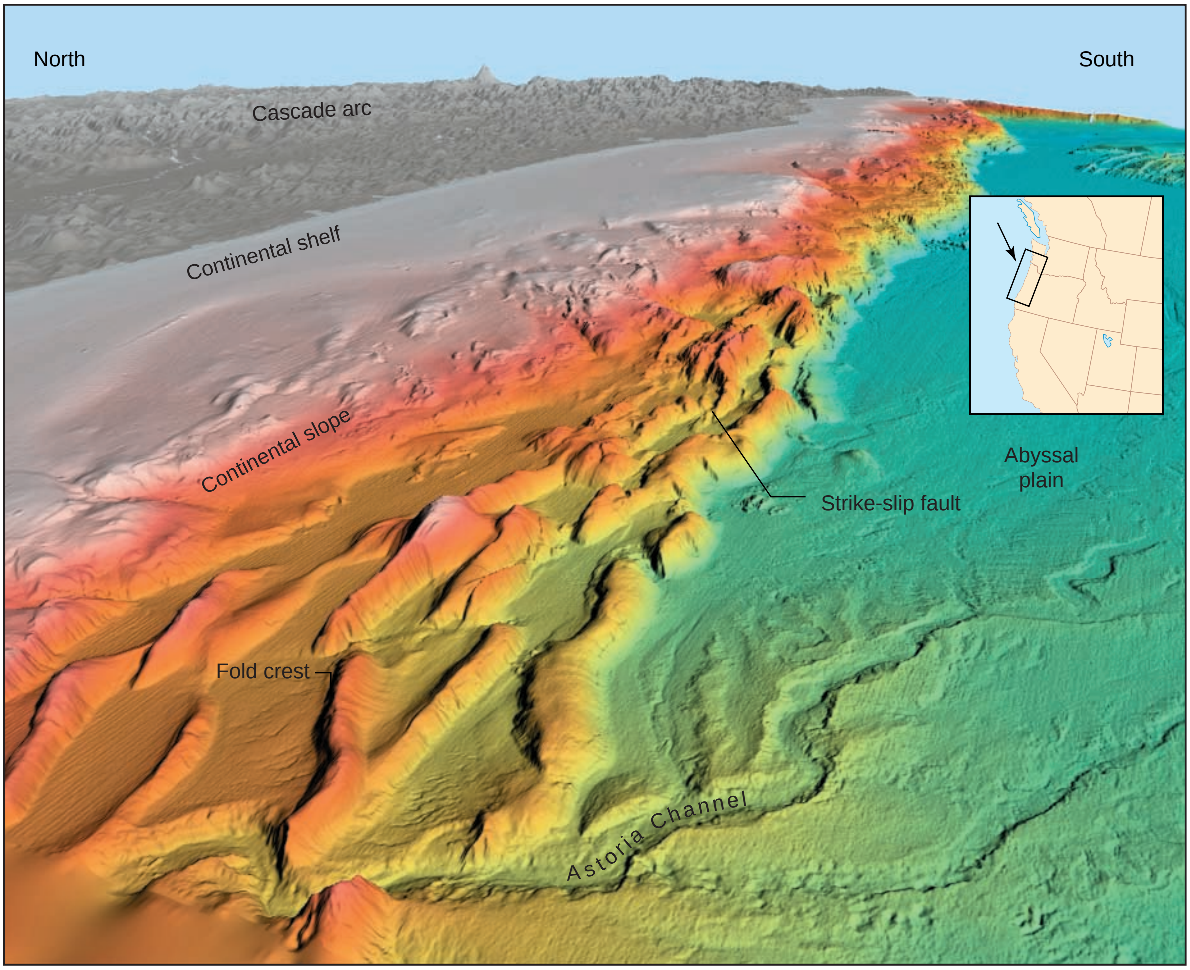
Strong contraction caused by horizontal compression occurs at many convergent plate boundaries, including ocean-continent and continent-continent boundaries. Especially striking is the deformation in the forearc region of a subduction zone. Unconsolidated sediment is scraped off the descending plate and piles up into a long wedge—called the **accretionary wedge**—in front of the overriding plate (Figure 21.10). The structures in an accretionary wedge are similar to those in the deformed snow that accumulates in front of a moving snow plow. The weak layer of snow is sliced off the solid road surface and stacks up in a thick, internally deformed pile immediately in front of the rigid blade. In this analogy, the overriding plate is the snowplow.

Folds of all sizes form in the accretionary wedge (Figure 21.10). As expected from the orientation of the applied stresses, the hinge planes of these folds are parallel to the trench and dip in the same direction as the subduction zone. Temperature and pressure changes during compression cause metamorphism, and slaty cleavage develops during folding. Thrust faults also cut through the soft sediment (Figure 21.10). These generally dip in the same direction as the subduction zone. As subduction continues and more sediment is scraped off, the deformed mass grows toward the trench, simultaneously shortening and thickening. Consequently, the front of the accretionary wedge grows steeper until it becomes unstable. It then collapses along faults that allow extension and thinning of the deformed mass. This cycle of tectonic shortening followed by gravitational collapse occurs repeatedly. The net result is uplift of deeper rocks to the surface.

How is an accretionary wedge formed?

This complicated deformation produces an accretionary wedge that is a chaotic mixture of rock types known as **melange,** one of the most structurally complex rock bodies in the crust. The deformed rocks are mostly sediments, but in some subduction zones, volcanic seamounts or other fragments of igneous oceanic crust also are scraped off the downgoing plate and incorporated into the wedge.

Not all of the sediment on the oceanic crust is scraped off to form an accretionary wedge. Apparently some is subducted deep into the mantle. As much as 20% to 60% of the sediment on the oceanic crust is subducted. Some sediment may be carried deep into the mantle, possibly to the core-mantle boundary.



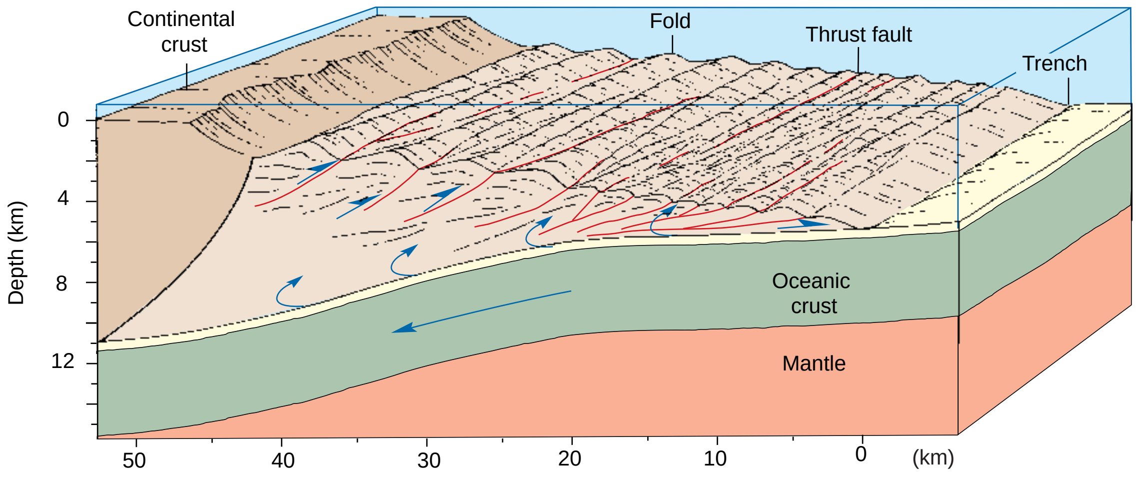


FIGURE 21.10 **Accretionary wedges** form at convergent plate margins as sediment and some igneous rock are scraped off the downgoing slab of oceanic rock. A southward-looking sonar image of the accretionary wedge along the coast of Oregon reveals that its surface is like a folded carpet. The ridges mark anticlines and areas where thrust sheets are stacked on top of one another. Transverse faults are strike-slip faults. Here the Juan de Fuca plate is being subducted beneath the North American plate. The block diagram shows the internal structure of an accretionary wedge derived from a seismic reflection profile. Faults and folds deform the rocks in the wedge. As sediment is removed from the downgoing plate, it is added to the base of the accretionary wedge. Stacks of thrust faults form above the downgoing oceanic crust. Folds of all sizes form between the thrust faults. *(Courtesy of L. Praxton and R. Haxby)*

FIGURE 21.11 **The Andes Mountains of South America** are forming by subduction of oceanic lithosphere beneath continental crust. Here, in the Atacama Desert of northern Chile, you can see a row of andesitic stratovolcanoes towering over an intensely deformed series of layered sedimentary rocks Deep in the mountain belt, metamorphic rocks are probably forming today. *(Hubert Stadler/CORBIS)*

What is distinctive about compression along a subduction zone at a continental margin?

If a subduction zone lies along a continental margin, compression produces a long **orogenic belt** (folded mountain belt) parallel to the continental margin. The Andes, parts of the Rockies, and the European Alps are classic examples.

In the central Andes, horizontal compression and crustal elevation are active today in the arc and backarc (Figure 21.11). Here, compression is causing pronounced folding, thrust faulting, and thickening of the crust. The fold hinges and thrust faults trend parallel to the arc and the linear belts of granitic batholiths. Intrusion of granitic plutons has added to the deformation. Not surprisingly, much of the lower and middle crust has been metamorphosed from the tremendous heat and pressure during compression and intrusion. The crust beneath the Andes is 75 km thick, probably thickened by intrusion of mantle-derived magmas as well as by folding and contraction (Figure 21.12). The deformed continental margin and its volcanic crest rise up to 6 km above sea level. Rapid erosion of such mountain belts occurs simultaneously with volcanism, intrusion, and deformation.

Another good example of compression at an ocean-continent boundary is provided by the Mesozoic history of western North America (Figure 21.13). During the Early Mesozoic, as oceanic lithosphere was subducted beneath the continent, a long magmatic arc developed along the western margin of North America. Behind the arc, thrusting and folding created a mountain belt. This extra load depressed the crust in much the same way that thick glacial ice sheets make the crust subside. Because of this subsidence, a shallow sea expanded onto the continent behind the thrust belt. Clastic sediments were shed from the thrust belt, forming coarse alluvial fans, floodplains, deltas, and shallow-marine shales that interfinger with carbonates deposited in deeper water. Thin beds of volcanic ash, erupted from volcanoes in the arc, also fell into the shallow sea. Today, these sedimentary basins are rich in natural resources, including coal, oil, and natural gas.

The European Alps have a long and complicated history that started with convergence of the oceanic part of the African plate beneath Eurasia (see page 479). Convergence has not yet completely consumed the oceanic crust of the Mediterranean Sea. Remnant volcanic arcs and scraps of unsubducted oceanic crust mark places where suturing of the two continents is still incomplete. In the Alps, great overturned folds called **nappes** (French, "tablecloths") show enormous amounts

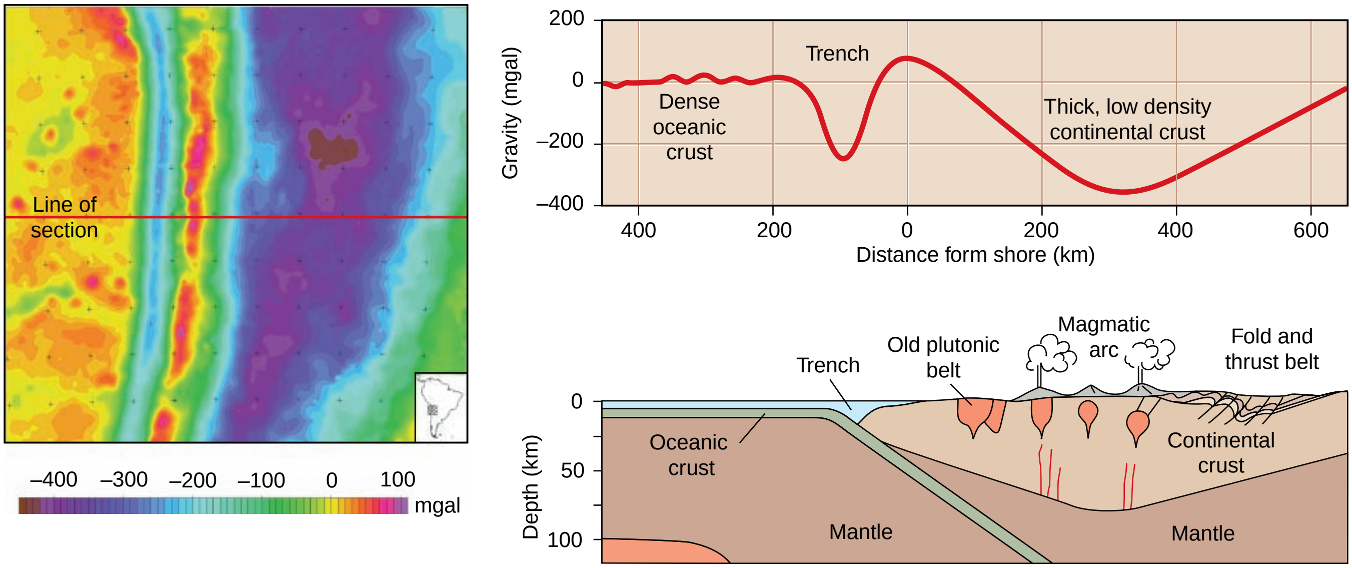


FIGURE 21.12 **The thick crust beneath the Andes** is revealed by the gravity field. The gravity profile and geologic cross section show that the outer bulge on the downgoing slab is marked by a positive gravity anomaly (red tones). Gravity is lower over the deep trench (light blue), which is filled with low-density water, and over the accretionary wedge, which is made of low-density sediment. Gravity is highest in the part of forearc that is underlain by the cold, dense subducting slab. The volcanic arc and folded mountain belt are marked by the lowest gravity anomalies (deep purples) because of the great thickness (as much as 70 km) of low-density crust beneath the Andes. *(Gravity map courtesy of M. Kösters and H. J. Götze)*

of crustal shortening (Figure 21.14C). The rocks are so strongly deformed that spherical pebbles were stretched into rods as much as 30 times longer than the original pebble diameters! Deformation is most intense near the continental margin and dies out northward toward the continental interior.

Compression in Continental Collision Zones

When two lithospheric plates that carry continents converge, the orogenic belt has very different characteristics than those formed during ocean-continent convergence. An excellent example of deformation related to continental collision is the Himalaya chain of southern Asia. The course of this collision is shown in Figures 21.15 and 21.16.

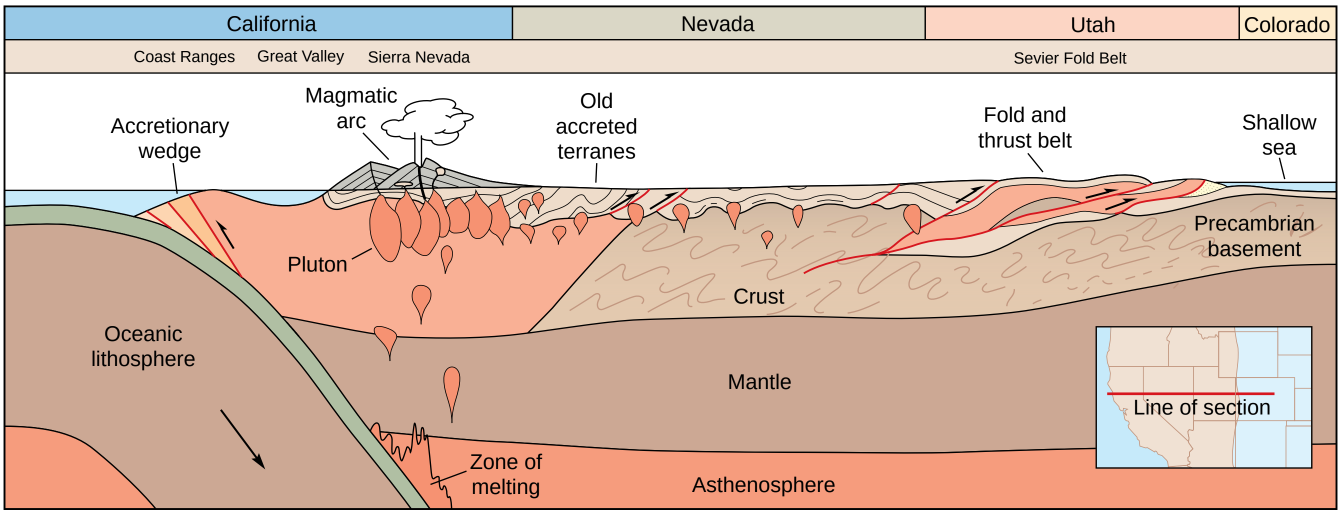
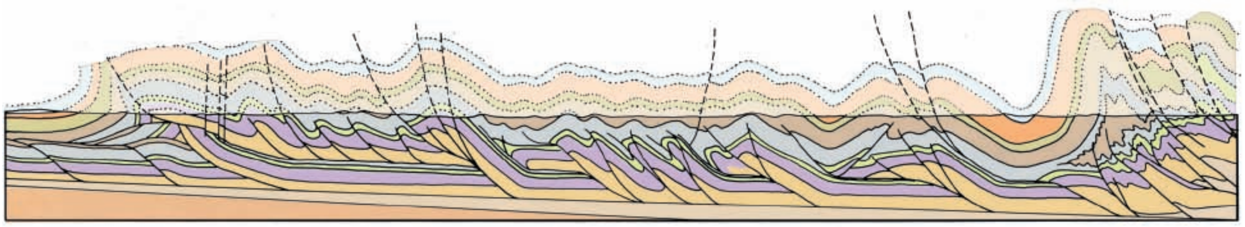
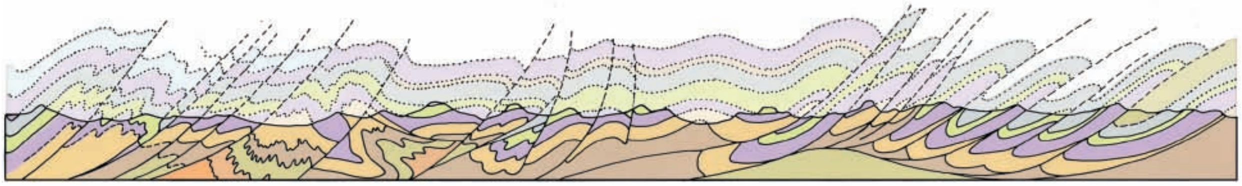


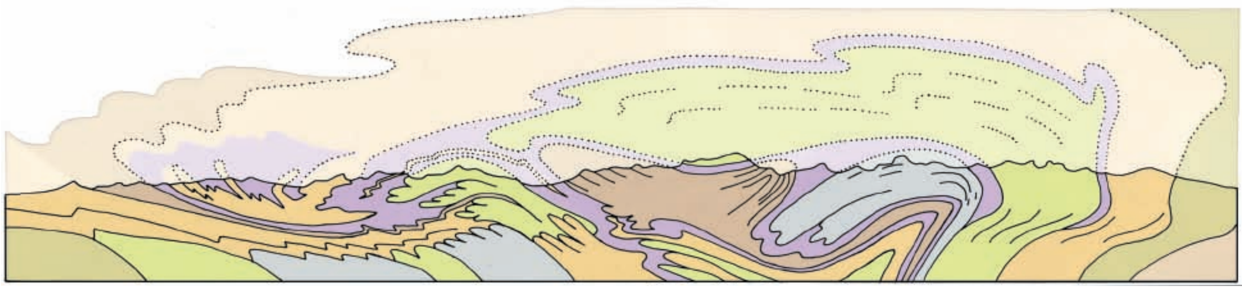
FIGURE 21.13 **Much of western North America developed at a convergent plate margin** 150 to 60 million years ago. The former locations of the accretionary wedge, magmatic arc, and folded mountain belt are shown. A sedimentary basin formed in the backarc region because of tremendous weight added by the mountain belt. The eroding mountains supplied sediment for the thick deposits that have accumulated along the margins of the mountains. With increasing distance from their sources, the sedimentary environments include alluvial fans, floodplains, deltas, and shallow-marine settings where shale interfingered with carbonates. *(After E. Miller and P. Gans)*



(A) The Canadian Rockies contain both folds and thrust faults.



(B) The Appalachian Mountains consist of tight folds and thrust faults. These geologically older mountains have been eroded to within 1000 to 3000 m of sea level. Resistant sandstones form the remaining mountain ridges.



(C) The Alps are a young range that consist of complex folds, many of which are overturned.

FIGURE 21.14 **The structure of folded mountain belts** reflects intense compression at convergent plate boundaries. Yet, each range can have its own structural style, as shown in these cross sections.

Folded Mountain Belts

The vast Himalaya orogenic belt formed during the past 100 million years, as oceanic lithosphere that was carrying India moved northward and was subducted beneath Asia. As a result, an accretionary prism developed on the southern edge of Asia. The sediments along the continental margin were also folded and faulted (Figure 21.15A). Simultaneously a magmatic arc developed as oceanic lithosphere was progressively consumed at the subduction zone (Figure 21.15B). When the two continental masses began to collide about 50 million years ago, subduction-zone volcanism ceased. The floors of the Black Sea and the Caspian Sea are remnants of oceanic lithosphere that were not subducted. Farther east, slices of oceanic crust were thrust onto the continent. These ophiolites occur north of the Himalayas along the suture (Figure 21.15C). As the two continents collided, India was thrust under Asia, effectively doubling the thickness of the crust to about 70 km. Its buoyancy prevented it from descending into the mantle more than perhaps 40 km below its normal depth. Thrust faults and folds formed a belt of deformed mountains, mainly in the overriding Asian plate. Much of the deformation was ductile and accompanied high-grade metamorphism in the deeper part of the crust. Because of compression and crustal thickening, the Himalayas and the extensive highlands of the Tibetan Plateau rose (Figure 21.15D). Mount Everest, Earth's highest peak, lies over the greatly thickened part of the crust. The continental masses were welded together to make a single large continent with an internal mountain range. At some point, the slab of descending oceanic lithosphere must have become detached and then sank, independent of the Indian continent. When it sank, volcanic activity and deep earthquakes ceased. Deformation associated with the collision drove Southeast Asia and parts of China eastward

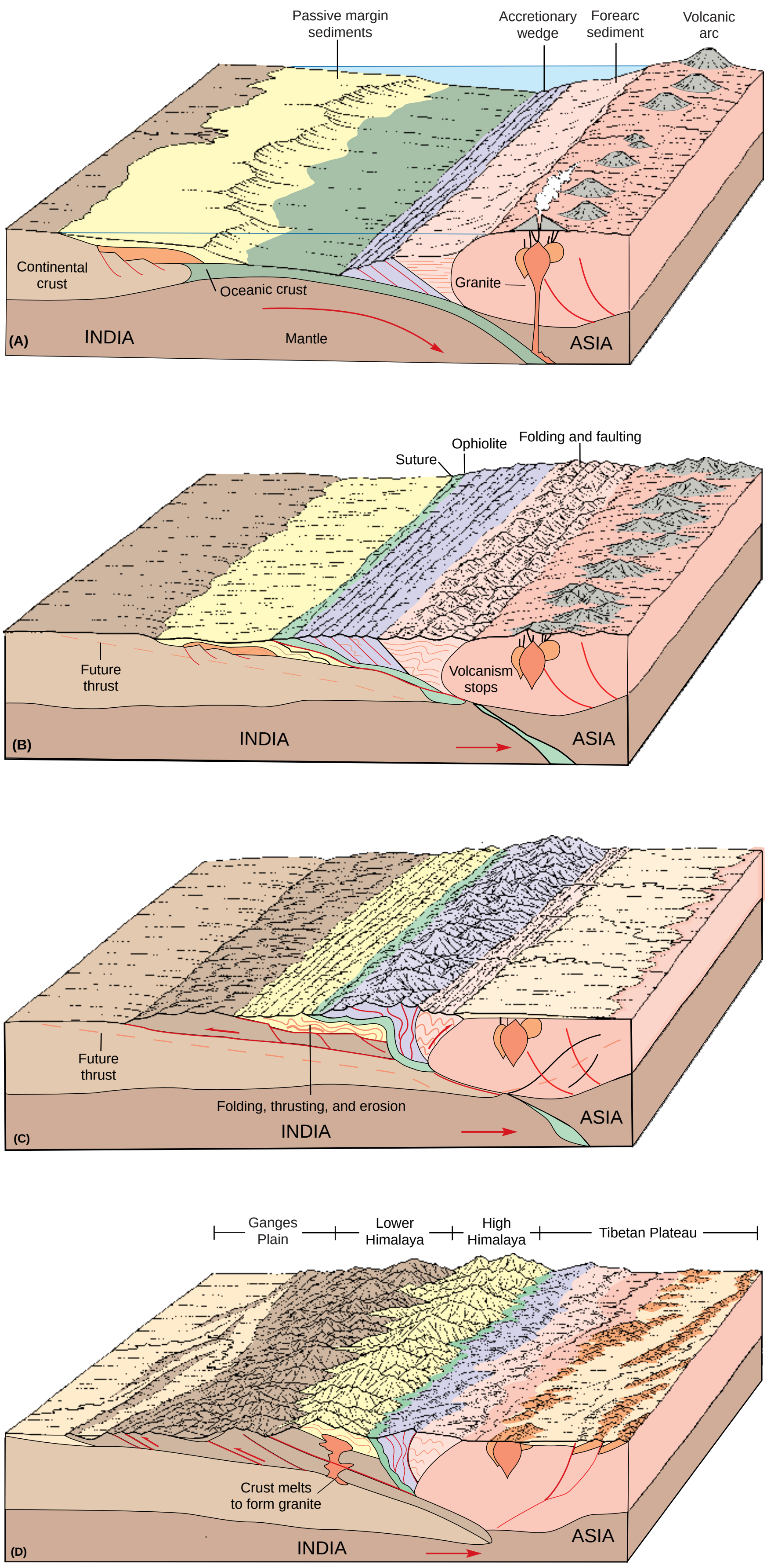


FIGURE 21.15 Continental collision formed the Himalaya Mountains and involved the deformation of oceanic and shallow marine sedimentary rocks. These were originally deposited along a passive continental margin. Collision produced a complex mountain range with large nappes and gently dipping thrust faults. As India and Asia converged, slivers of oceanic crust were thrust onto the continents as ophiolites. A double layer of continental crust formed, resulting in very high mountains. The continents were "welded" together. Eventually, the descending oceanic portion of the plate detached from the rest of the plate and sank independently. Once the slab was consumed, volcanic activity and deep earthquakes ended. However, during high-grade metamorphism in the roots of the mountain range, the continental crust itself may partially melt to form granite with distinctive compositions; these are found in no other tectonic setting.

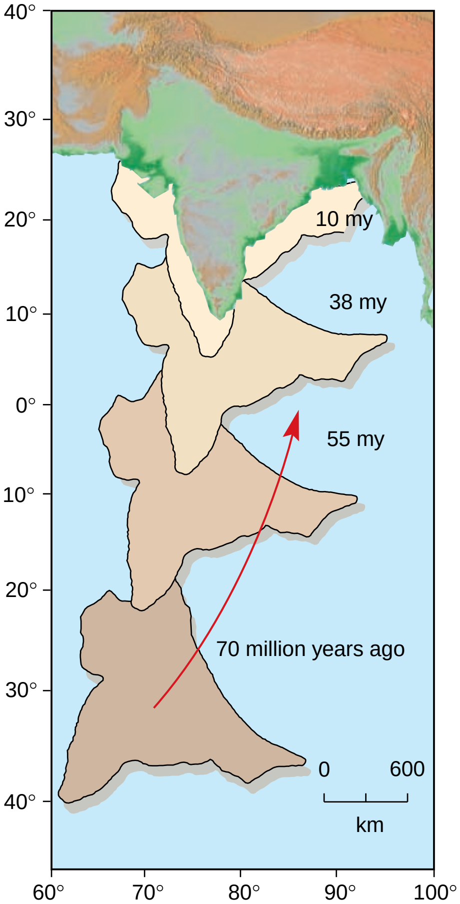


FIGURE 21.16 The Himalaya mountain belt formed by the collision of the Indian and Eurasian plates. Seventy million years ago, India rifted away from Africa and moved rapidly northward until it collided with Asia. The collision started about 25 million years ago and built the high Himalaya range and the Tibetan Plateau to the north.

along faults that fan away from India (Figure 21.17). Of course, erosion of the mountain belt continued throughout this long process.

It is interesting to contrast compression in the Himalayas with compression in the Appalachian Mountains of the eastern United States. Compression in the Appalachians was produced by several episodes of subduction of oceanic lithosphere, followed by collision between North America and Europe or Africa. These collisions occurred during the Paleozoic Era, making the Appalachian Mountains more than 300 million years old. A cross section of the Appalachians shows a different style and extent of deformation than in the Himalayas (Figure 21.14B). The major structural features are tight folds and thrust faults. Orogenic (or regional) metamorphism accompanied the collisions, but most of the plutonic rocks are subduction-related and preceded the collisions.

Extension at Convergent Boundaries

It may seem paradoxical that extension occurs in regions where two plates of lithosphere converge. However, mild extension—not compression—is the dominant deformation occurring at most oceanic volcanic arcs and at some continent arcs as well. Before we consider why, let us describe the evidence for extension and its effects.

Extension Above Subduction Zones. In island arcs and some continental arcs, grabens and normal faults typically are centered on the active volcanic region. For example, the modern-day andesitic composite volcanoes of Ecuador lie in a graben hundreds of kilometers long. The trends of these faults are parallel to the trench, indicating that the direction of extension is perpendicular to the trench. Extension like this is very common in island arcs. Compression, marked by thrust faults and folding, is found in few modern-day island arcs.

The backarc basins found behind most oceanic island arcs also reveal the effects of extension. These shallow oceanic basins are traversed by normal faults and have high heat flow and active seafloor volcanism. Extension may ultimately lead to **backarc spreading.** The Mariana and Tonga-Kermadec arcs of the western Pacific have such basins. The backarc regions of many continental arcs are also marked by extension and subsidence like that behind oceanic island arcs. For example, backarc extension is active in the Aegean Sea behind a volcanic arc rooted in continental crust (Figure 21.18). Normal faulting has produced many narrow grabens with trends that are parallel to the arc and has allowed most of the area to drop below sea level. However, no oceanic crust has yet developed in this region of extension and subsidence.

The islands of Japan, with their striking composite volcanoes and active subduction zones, were once part of the Asian mainland before backarc extension opened the sea of Japan. This sea marks more extensive development of rifting above a subduction zone and is underlain by oceanic crust.

Extension and Continental Collision. As a secondary effect of the collision of two continents, extensional tectonics may also develop. The Rhine Graben of central Europe was created by the collision that raised the Alps. Lake Baikal in southern Siberia lies in one of Earth's deepest continental rifts, but the stresses responsible for it are related to the collision that formed the Himalayas.

Why does extension occur in backarc basins?

Compression Versus Extension. What controls whether a particular convergent margin experiences horizontal compression or extension? No one yet knows for sure, but several possible causes have been identified. One important variable might be the angle of the subducting plate. For example, the Mesozoic fold and thrust belt of western North America formed during a period of rapid convergence. The subducting plate was probably inclined at a low angle, perhaps even dragging along the base of the overriding lithosphere. Such a close coupling between subducting plate and overriding plate may cause contraction in the backarc.

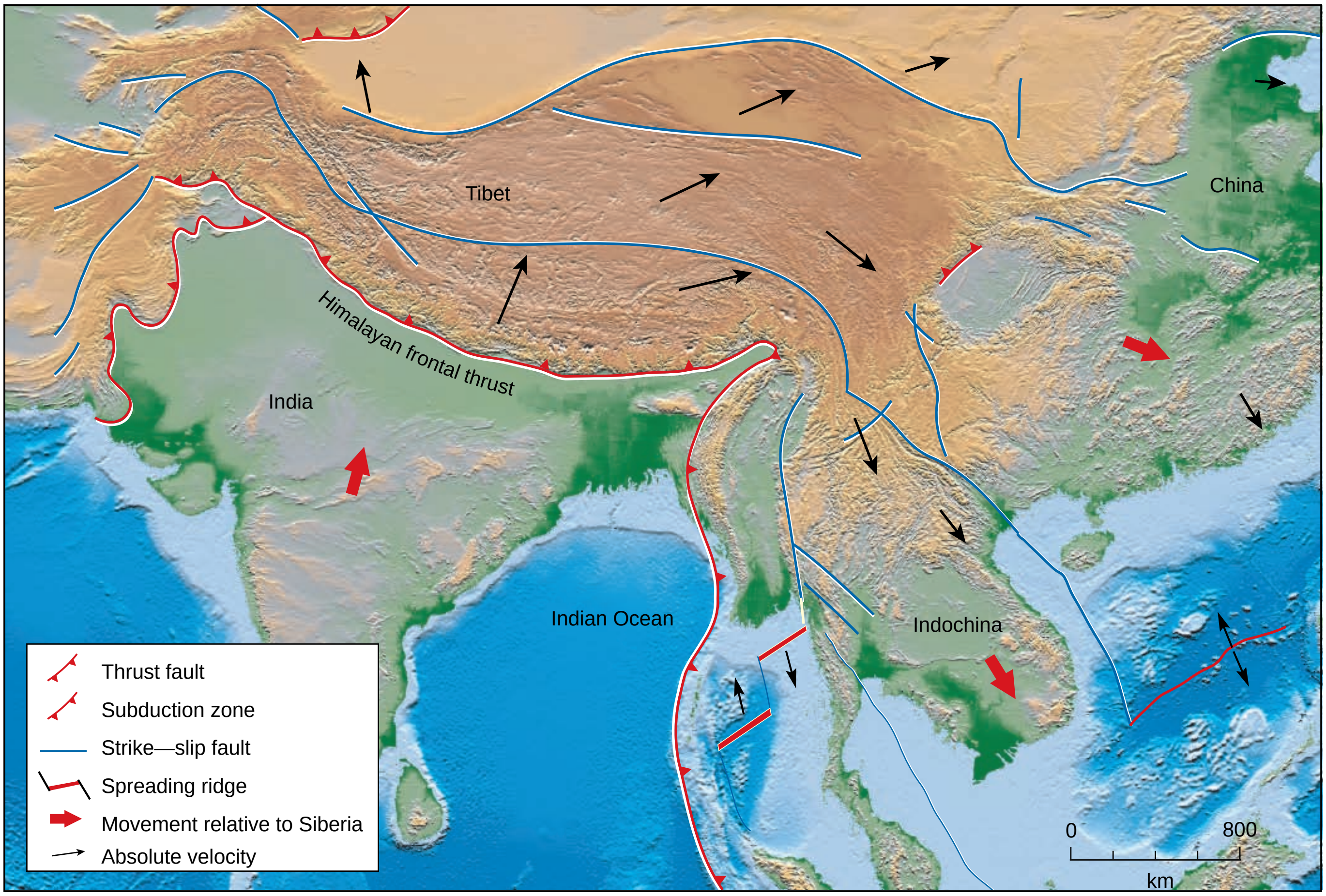


FIGURE 21.17 Complex folds, mountains, and plateaus mark the collision zone between India and Eurasia, as shown on this digital shaded relief map. The collision also drove parts of Southeast Asia and China eastward along lengthy strike-slip faults or shear zones. *(Courtesy of Ken Perry, Chalk Butte, Inc.)*

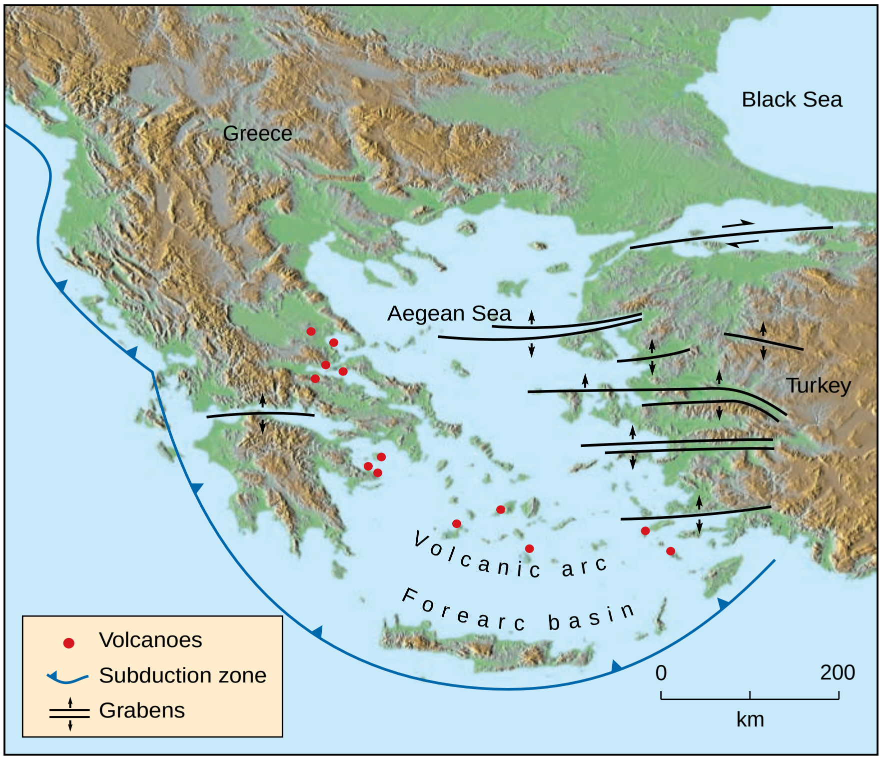


FIGURE 21.18 The Aegean backarc basin developed in continental crust above a subduction zone in the eastern Mediterranean Sea. Like backarc basins formed within ocean basins, the Aegean basin has subsided, and normal faulting has caused extension. However, no oceanic crust has yet developed. *(Courtesy of Ken Perry, Chalk Butte, Inc.)*

In addition, the subduction of young, hot, and buoyant oceanic lithosphere that drags along the base of the overriding slab may be important for the development of some fold and thrust belts. Another way to cause a compressional fold and thrust belt is by collision of a continent with minor arcs or continental fragments. On the other hand, backarc extension may be dominant where these factors are missing.

One cause of extension at convergent plate boundaries may be the convective flow of the mantle beneath the arc (Figure 21.2). As described above, the downgoing slab may drag the viscous mantle with it, causing hot asthenosphere from deeper in the mantle to flow upward and take its place. A convection pattern is thus formed in the asthenosphere above the subducting plate. It thins the lithosphere and may cause spreading behind the volcanic arc.

The absolute motion of the overriding plate (rather than the more obvious relative movement) also may play a role in developing extensional structures (Figure 21.1). For example, extension is common in arcs where the overriding plate is moving away from the trench.

MAGMATISM AT CONVERGENT BOUNDARIES

Magma in a subduction zone is probably generated when water in the descending oceanic crust is driven out and rises into the overlying mantle. The addition of water lowers the melting point of the mantle rock and causes partial melting. Differentiation of this magma produces andesite and rhyolite, which rise and intrude as plutons or extrude to make long-lived composite volcanoes or calderas.

What produces a magmatic arc at convergent margins?

Most volcanoes erupting above sea level are clearly associated with subduction zones at convergent plate boundaries (Figure 21.8). The geographic setting for volcanic activity along such zones depends on the type of plate interaction. Where two oceanic plates converge, an island arc forms, as you have seen. Where a continent is on the overriding plate, similar volcanic activity develops in a folded mountain belt. In both cases, the close association of volcanism and convergent plate boundaries is clear. In contrast, where two continents collide, volcanism is rare, although magmas form granite plutons at depth.

Island Arc Magmatism

On a map, the most obvious manifestation of convergence of two oceanic plates is an arcuate chain of volcanoes that rise from the seafloor (Figure 21.8). The volcanic arc forms on the overriding plate and is parallel to the curving trench. Typical island arcs are the Tonga Islands, the Aleutian Islands, and the West Indies (Antilles) Islands (Figure 21.1).

Igneous Crystallization

These volcanoes lie about 100 km from the trench and 100 km above the inclined seismic zone that shows the location of the subducted slab. They mark a zone of voluminous magma production and high heat flow (Figure 21.6). The volcanoes are built on igneous intrusions and deformed metamorphic rocks, including remnants of oceanic crust formed at a midoceanic ridge. Most of these volcanoes are large composite volcanoes that erupt large volumes of andesite and lesser amounts of basalt and rhyolite.

The major volcanoes rise 1 to 2 km above their surroundings and are quite regularly spaced, about every 50 to 75 km. Smaller extrusions build a nearly continuous ridge connecting the major cones. Most island arcs are several hundred kilometers wide and extend discontinuously along the length of the trench.

Continental Arc Magmatism

A continental volcanic arc is a chain of many composite volcanoes on the margin of the continent above a subduction zone. The active volcanoes and underlying plutons are about 100 to 200 km landward from the trench. Subsidiary vents, lava domes, cinder cones, and fissure vents dot the landscape between the major volcanoes. The deep part of the arc consists largely of plutonic rocks that are the roots of volcanic systems. Multiple plutons intrude one another and form long, linear batholiths. The plutons are typically diorite to granite in composition. These plutons are commonly larger and more silicic than those found in island arcs. Moreover, they intrude into preexisting continental crust that is made of folded and thrust-faulted sedimentary rocks overlying a basement of older igneous and metamorphic rocks.

Generation of Magma in Subduction Zones

Magmatism along subduction zones is quite different from the basaltic fissure eruptions of divergent plate boundaries. The magmas generated at subduction zones are characteristically andesite or even rhyolite. They are richer in silica than basaltic magma and thus are more viscous. Consequently, water dissolved in the magma cannot escape easily to form gas bubbles. Moreover, subduction zone magmas contain more water and other dissolved volatiles than do those formed at divergent plate boundaries. These two characteristics result in violent, explosive eruptions from central vents. They commonly produce ash flows, composite volcanoes, and collapse calderas, as well as viscous lava flows and domes. These important differences are probably caused by dramatically different mechanisms of magma generation.

How do andesitic and silicic magmas form in subduction zones?

What triggers the generation of magma at subduction zones? How does insertion of a *cold* slab into the mantle create *hot* molten magma? The probable answer to this paradox lies not in the temperature of the plate, but in its high water content. The water in the descending plate was incorporated into the oceanic crust by metamorphism at a midoceanic ridge. This water is eventually released from the subducting oceanic lithosphere and rises into the overlying wedge of mantle. There the water lowers the melting point of the mantle rock.

Figure 21.19 illustrates how the release of water from the subducting slab is involved in the generation of magma. Water moves along with the descending plate into a subduction zone as a pore fluid in sediments and as water included in the

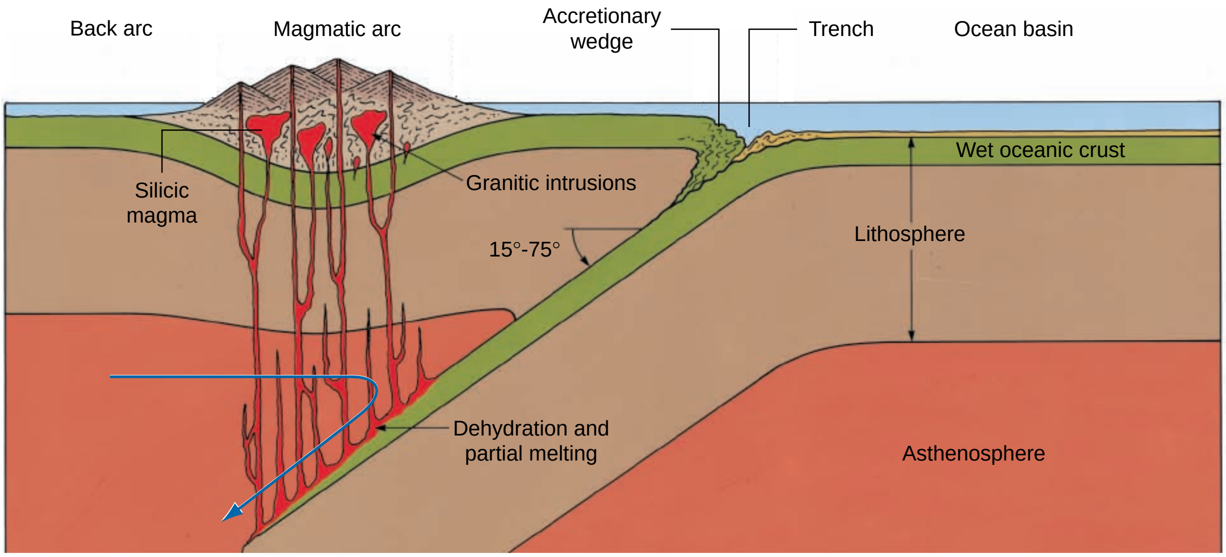


FIGURE 21.19 **Magma at convergent plate boundaries** is generated at depths of about 100 to 150 km. Subduction of oceanic crust carries sediment and basalt into the hot asthenosphere. (This sediment and basalt were altered earlier by ocean ridge metamorphism at a divergent plate boundary.) The descending slab is slowly heated; eventually, the hydrous minerals in the crust decompose and release water. At this critical depth, the water rises into the overlying mantle, causing it to melt partially. This basaltic magma rises buoyantly into the crust, where it may differentiate to form andesite and rhyolite. The magmas may crystallize to make plutons or erupt at the surface.

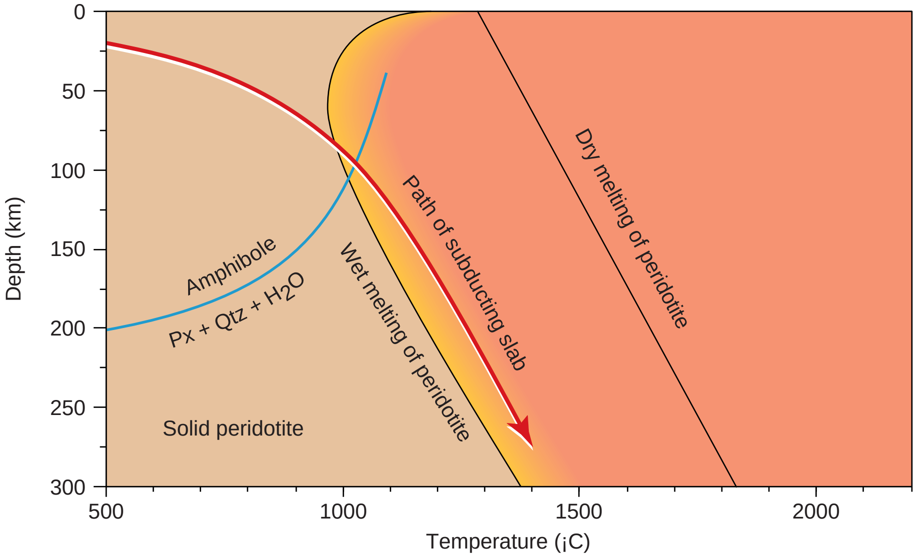


FIGURE 21.20 **The generation of magma in a subduction zone** is primarily due to the role played by water. As a descending plate slides into the mantle, it follows a path of increasing pressure and temperature (red arrow). Where the path crosses the breakdown curve for amphibole (blue line), an important mineral in metamorphosed oceanic crust, water is released. The buoyant fluid rises into the overlying mantle and there induces partial melting. Wet peridotite begins to melt at a temperature nearly 500°C lower than dry peridotite. This new mafic magma is wetter and more oxidized than magma produced at midocean ridges and may differentiate to make silicic magma such as andesite or rhyolite.

Why is magma generated at convergent margins?

structures of minerals formed by ocean ridge metamorphism. The descending slab is subjected to progressively higher temperatures and pressures (red line in Figure 21.20). Water trapped in the sediments is probably squeezed out at shallow depths beneath the accretionary wedge, but some of the water tied up in the chemical structures of minerals may remain in the oceanic crust until depths greater than 100 km are reached.

Figure 21.20 shows a typical temperature-pressure path encountered by the upper part of the cold slab as it descends into the hot mantle. Somewhere between 100 and 200 km deep, the slab has warmed enough that the hydrated minerals are no longer stable. Consequently, they break down to form new minerals that lack water, together with a separate water-rich fluid. Figure 21.20 shows that the mineral amphibole in wet oceanic crust may **dehydrate** to form pyroxene, quartz, and water.

The water released rises buoyantly into the hot mantle peridotite overlying the cold slab, causing the peridotite to partially melt (Figure 21.19). Careful examination of Figure 21.20 reveals why melting happens: Water dramatically lowers the temperature at which peridotite begins to melt. You can see this relation by comparing the melting temperature at 100 km on the dry-melting curve with the melting temperature on the wet-melting curve. Consequently, the wet mantle above the subducting slab may melt without increasing its temperature. (Here, water acts as a flux, much like fluorite added to iron ore in a steel furnace will make it melt at a lower temperature than pure iron ore.)

As this hybrid magma (derived partially from the oceanic crust and partially from the overlying mantle), rises it reacts extensively with the overlying crust (Figure 21.21). Thus, this magma may contain components derived from oceanic sediments, from metamorphosed oceanic basalt, from peridotite in the mantle wedge, and from the overlying crust. On the way to the surface, the magma also may mix with other batches of magma. Or, it may cool and experience fractional crystallization to form andesite or rhyolite magmas, which are even richer in silica.

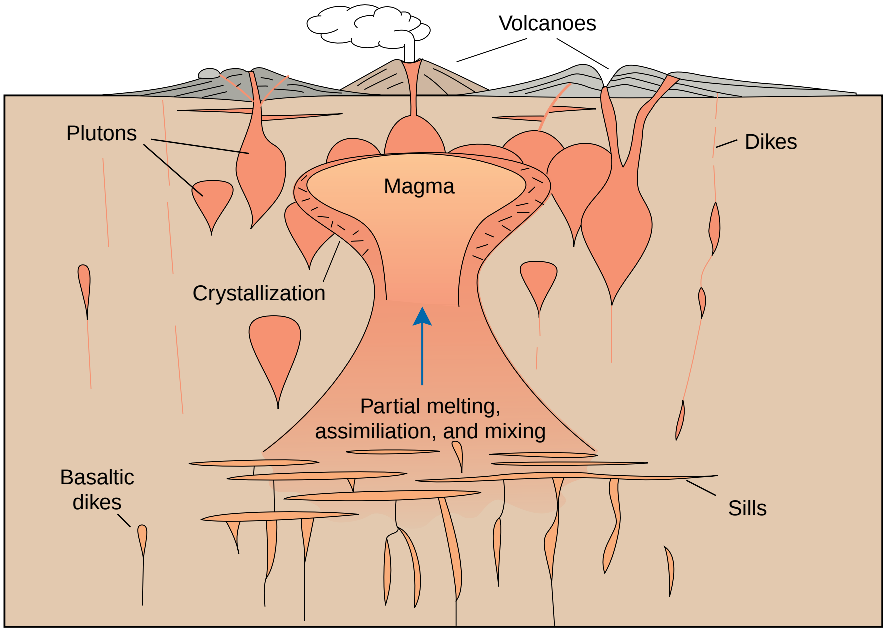


FIGURE 21.21 **Intrusions at convergent plate margins** are one of the major ways that continental crust is produced. Hot magma rising from a subduction zone may assimilate crustal rocks or mix with other magma simultaneous with fractional crystallization. Through these processes, basaltic magma differentiates to make andesite and rhyolite. Magma may rise in teardrop-shaped bodies or through fractures or dikes that merge into larger and larger masses.

Eventually, magma that started deep in the subduction zone cools to form plutons or extrudes as lava or ash flows. The fundamental point is that silica-rich continental crust is formed by the extraction of low-density material from the mantle.

Note that at subduction zones, magma is generated by *partial melting*—a process that differentiates and segregates the materials of Earth. Here, magma rich in silica is produced. It is concentrated in island arcs or in granite plutons in mountain belts of the continents. Unlike basalt, which may become dense enough to be subducted, this silica-rich material cannot sink into the mantle. It becomes concentrated to form additional continental crust.

Subduction zone magmas are distinct from those in most other tectonic settings. We have already emphasized that the typical subduction zone magma is andesitic in composition, but the full spectrum of igneous rock compositions occurs. Moreover, it is very important to remember that subduction zone magmas are characteristically enriched in the water, as well as other volatile components such as chlorine, sulfur, and oxygen. These elements were probably extracted from the subducted oceanic crust which had been altered by ocean ridge metamorphism.

Generation of Magma in Continental Collision Zones

Where two continents collide, hot mantle-derived magmas do not form after the subduction zone disappears (Figure 21.15). Nonetheless, small volumes of magma are produced. A distinctive granite is the most important igneous rock. In this setting, continental rocks, including metamorphosed shales and other clastic sedimentary rocks, can partially melt to form granitic magmas that are rich in silica and aluminum. Many of these granites contain minerals rarely found in other types of granite (such as muscovite, garnet, tourmaline, and cordierite). Such magmas do not rise far from their sources before they crystallize; only rarely do they erupt to form lava or ash flows. The heat for melting probably comes from deep burial by tectonic underthrusting of the continental crust. In the Himalayas, for example, partial melting of continental crust has produced sill-like sheets of muscovite granite that intrude into previously folded and metamorphosed rocks.

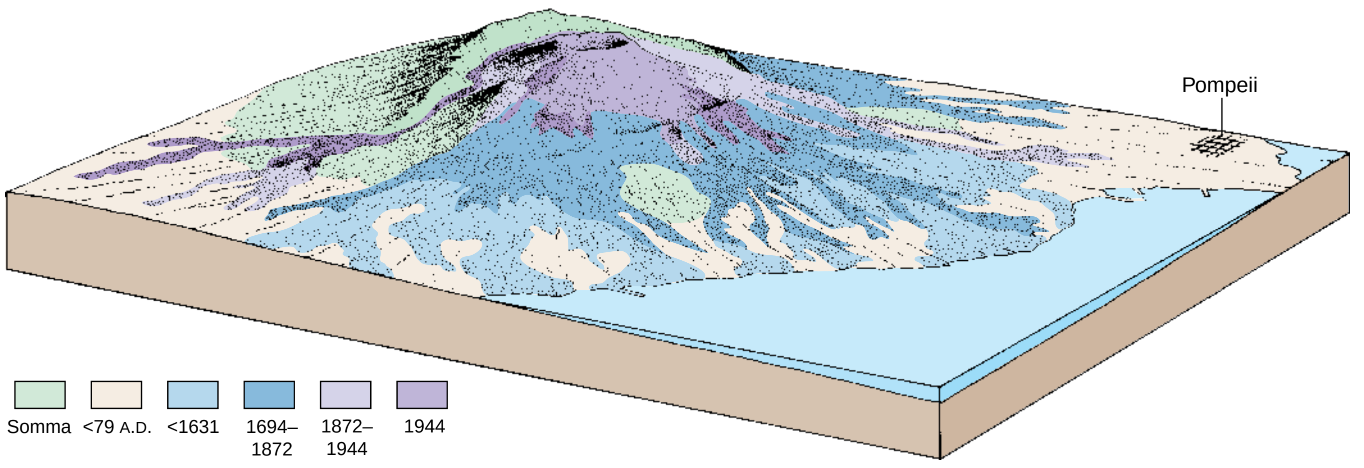


FIGURE 21.22 **Vesuvius** erupted and buried Pompeii, Italy, with ash in A.D. 79. It is one of several composite volcanoes that lie above a westward-dipping subduction zone beneath Italy. People asphyxiated by poisonous gas during the eruption were buried in the ash. Eventually, the bodies decomposed, leaving cavities in the ash. By filling these cavities with plaster, archeologists have made detailed casts. Excavations provide important insights into volcanic activity at convergent plate margins.

VOLCANIC ERUPTIONS AT CONVERGENT BOUNDARIES

Volcanoes above subduction zones commonly erupt violently to form viscous lava flows, lava domes, or ash flows or ash falls. Tsunamis, lahars, and debris avalanches are also common. Although the volcanoes erupt infrequently, some eruptions can be predicted.

There have been many volcanic eruptions at convergent plate boundaries over the past 1000 years. Many have been fatal. In the last 100 years, about 100,000 people have died as a result of volcanic eruptions. Historical accounts of a few of these eruptions help us understand the nature of volcanic activity associated with converging plates. The following sections recount four of the most spectacular, devastating eruptions in recorded history: Mount Vesuvius, Krakatau, Mont Pelée, and Mount St. Helens.

A.D. 79—Mount Vesuvius, Italy

The spectacular cone of Mount Vesuvius looms above the skyline of Naples, Italy. Over the ages, it has repeatedly erupted magma generated in a subduction zone beneath the Italian peninsula (Figure 21.22). In A.D. 79, Mount Vesuvius erupted catastrophically. An extraordinarily vivid eyewitness account of the eruption was recorded by Pliny the Younger, then a 17-year-old boy. He related details of how his famous uncle, Pliny the Elder, died while observing the volcano. Along with him, many perished in the destruction of the two cities of Pompeii and Herculaneum.

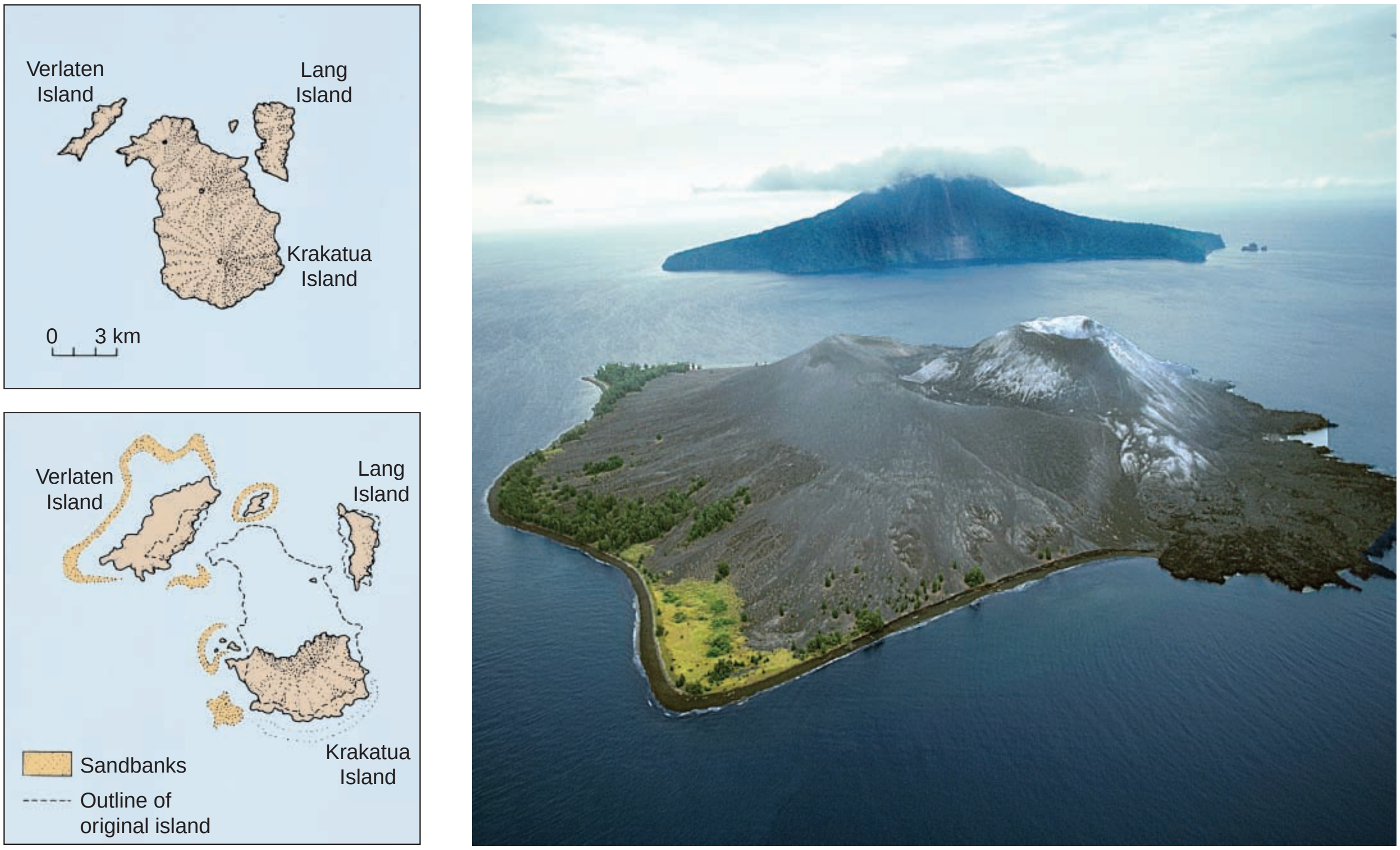


FIGURE 21.23 **Maps of Krakatau before (top) and after (bottom) its 1883 eruption** show the force of violent volcanic eruptions at convergent plate boundaries. Krakatau is a composite volcano along the Indonesian arc. All that remains of the volcano are several small islands like the one in the background. A small volcanic cone (foreground) has been rebuilt over the center of the old volcano. *(Charles O'Rear/CORBIS)*

Beginning in A.D. 63 and continuing for 16 years, earthquakes shook the western coast of Italy. Then, on the morning of August 24, A.D. 79, Mount Vesuvius exploded with a devastating eruption of white-hot ash and gas. Within two days, ash falling from the cloud buried Pompeii, which was directly downwind and near the volcano. Many people suffocated by sulfurous fumes or were burned by the searing heat of brief pyroclastic surges that swept through the town; others died in their homes when roofs collapsed from the weight of the tephra. The entire town and most of its 20,000 inhabitants were buried by ash and forgotten for more than 1000 years, until Pompeii was excavated in 1748 (Figure 21.22). By contrast, the town of Herculaneum was buried nearly instantaneously by rapidly moving ash flows that accumulated to a depth of 20 m.

The ash fall and flow that buried Pompeii and Herculaneum are first-class examples of the type of violent eruption that is common in volcanoes along convergent margins. The once-smooth and symmetrical cone of Mount Vesuvius was shattered by the explosion, which created a large caldera where a peak once existed. Subsequent eruptions have built a new cone inside the older caldera.

1883—Krakatau, Indonesia

Krakatau is a small volcanic island west of Java, part of an island arc along the subduction zone associated with the Java Trench (Figure 21.1). After remaining dormant for two centuries, Krakatau began to erupt on May 20, 1883. The eruption culminated in a series of four great explosions on August 26 and 27. One of them was heard in Australia, 4800 km away. The explosions are considered the greatest in recorded history. The whole northern part of the island, which stood about 600 m high, was blown off, forming a huge caldera, 300 m below sea level (Figure 21.23).

Tremendous quantities of ash were thrown high into the atmosphere, and some circled the globe for 2 years. Krakatau was uninhabited, but more than 36,000 people were killed in Java and Sumatra by the huge tsunami produced by the eruption.

1902—Mont Pelée, West Indies

Ash flows are an important part of volcanism along convergent plate margins, and the great eruption of Mont Pelée, on the island of Martinique in the West Indies, helped initiate an understanding of this type of eruption. Pelée's eruption was preceded by nearly a month of warnings, in the form of extrusions of steam and fine ash from the volcanic vent, accompanied by numerous small earthquakes.

Then, on May 8, 1902, a gigantic explosion blew ash and steam thousands of meters into the air. The dense, hot ash fell back out of the atmosphere and moved as an ash flow, sweeping down the mountain's slopes like a high-speed avalanche. In fewer than 2 minutes, the hot, incandescent ash flow moved 10 km from the side vent on Pelée and swept over the city of St. Pierre. It annihilated the entire population of more than 30,000 people, except for one man, a prisoner who was being held underground in the city jail. Every flammable object was instantly set aflame. The rushing cloud of ash moved over the waterfront and across the sea surface, capsizing all of the ships.

What characteristics do volcanoes at convergent plate margins have in common?

The ash flow was a mixture of hot glass shards and pumice that flowed at the base of a billowing cloud of gas. The fundamental force that causes ash to flow rapidly is simply the pull of gravity, just like an avalanche of snow. But the key to the speed of the cloud is that such a mixture of hot gas and fragments of ash is highly mobile.

Intermittent ash-flow eruptions continued on Mont Pelée for several months. By October, a bulbous dome of lava, too viscous to flow very far, had formed in the crater. A spire of solidified lava was then slowly pushed up from a vent in the dome, like toothpaste from a tube. The spire repeatedly crumbled and grew again from the lava dome. Each collapse produced a hot avalanche that rushed down the slopes of the volcano.

1980—Mount St. Helens, Washington State

Mount St. Helens

The best-documented eruption of a composite volcano related to a subduction zone is Mount St. Helens in Washington State. On May 18, 1980, it erupted with a force estimated to be 500 times greater than the atomic bomb that destroyed Hiroshima, Japan, at the end of World War II (Figure 21.24).

Mount St. Helens is part of the Cascade Range, which extends about 1500 km from British Columbia to northern California. Mount St. Helens is the youngest of the 15 major volcanoes in the Cascade Range. It consists of coalesced dacite domes, lava, and interlayered ash deposits.

Mount St. Helens had been dormant for 123 years, but on March 20, 1980, it began to stir with a series of small earthquakes. After a week of increasing local seismicity, it began to eject steam and ash. A series of moderate eruptions continued intermittently for the next six weeks. Within a few days after these first eruptions started, the U.S. Geological Survey issued warnings. During the weeks to come, the U.S. Forest Service and Washington State officials closed all areas near the mountain, undoubtedly saving thousands of lives.

By the second week of activity, more than 30 geologists had gathered to conduct a wide variety of studies. In particular, they monitored the development of a large bulge on the north flank of the mountain. By the end of April, the bulge was 2 km long and 1 km wide and was expanding horizontally at a steady rate of 1.5 m/day. Clearly, the mountain was being inflated by magmatic intrusion.

By monitoring the bulge, seismicity, and the gas emissions, geologists believed they would detect some significant change to warn of an imminent large eruption. However, no anomalous activity occurred. In fact, seismic activity decreased. Thirty-nine earthquakes were recorded on May 15 and only 18 on May 17.

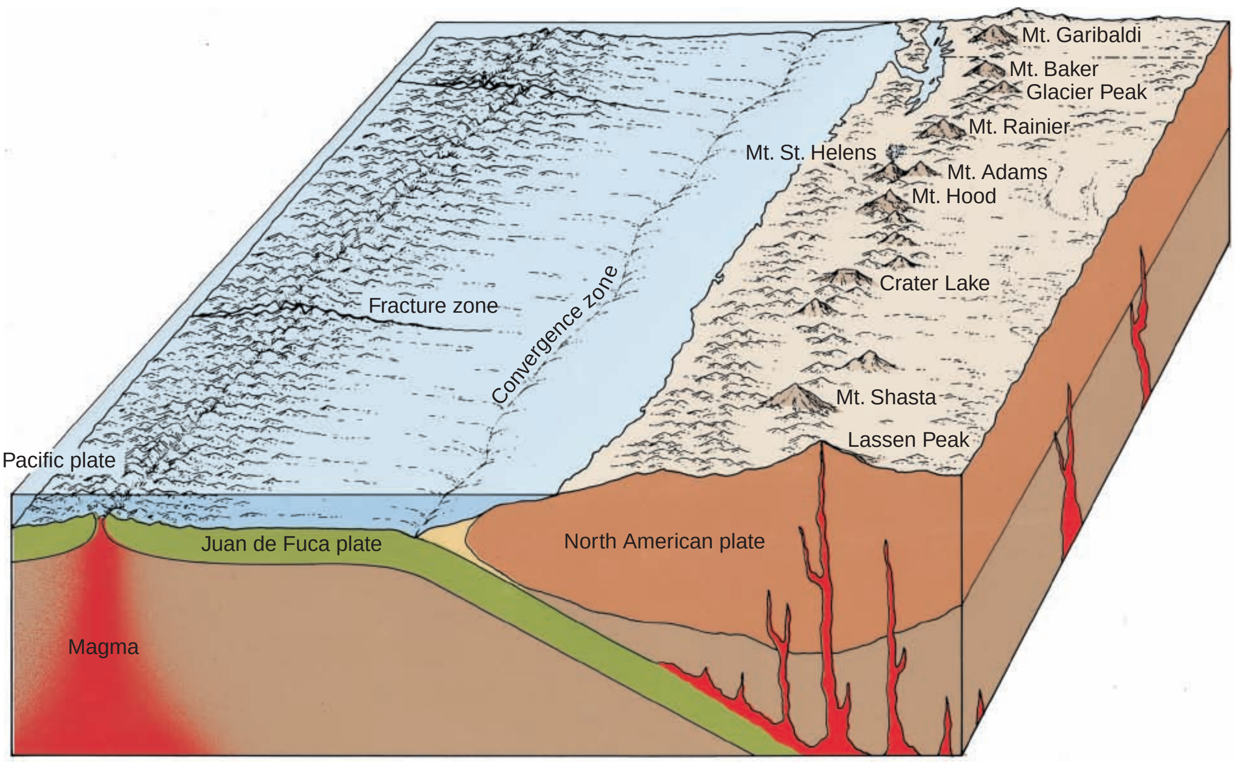


FIGURE 21.24 **The 1980 eruption of Mount St. Helens** in Washington State was one of the largest and most scientifically important to occur in the United States. The eruption and devastation by explosive blasts and ash are typical of composite volcanoes built above subduction zones. The Cascade Range contains 15 large composite volcanoes, extending in a line from British Columbia to northern California. These volcanoes are formed by subduction of the Juan de Fuca plate beneath the North American plate. *(Photography courtesy of U.S. Geological Survey)*

(A) Mount St. Helens had been dormant for 123 years until it erupted on May 18, 1980.

(B) The May 18 eruption was triggered when a landslide removed the side of the volcano and caused a lateral blast of incandescent gas and ash toward the north.

(C) The eruption removed a large part of the northern flank, leaving a breached crater. The landscape was ravaged by the blast and by later pyroclastic flows and lahars. A small lava dome has subsequently developed in the horseshoe-shaped crater.

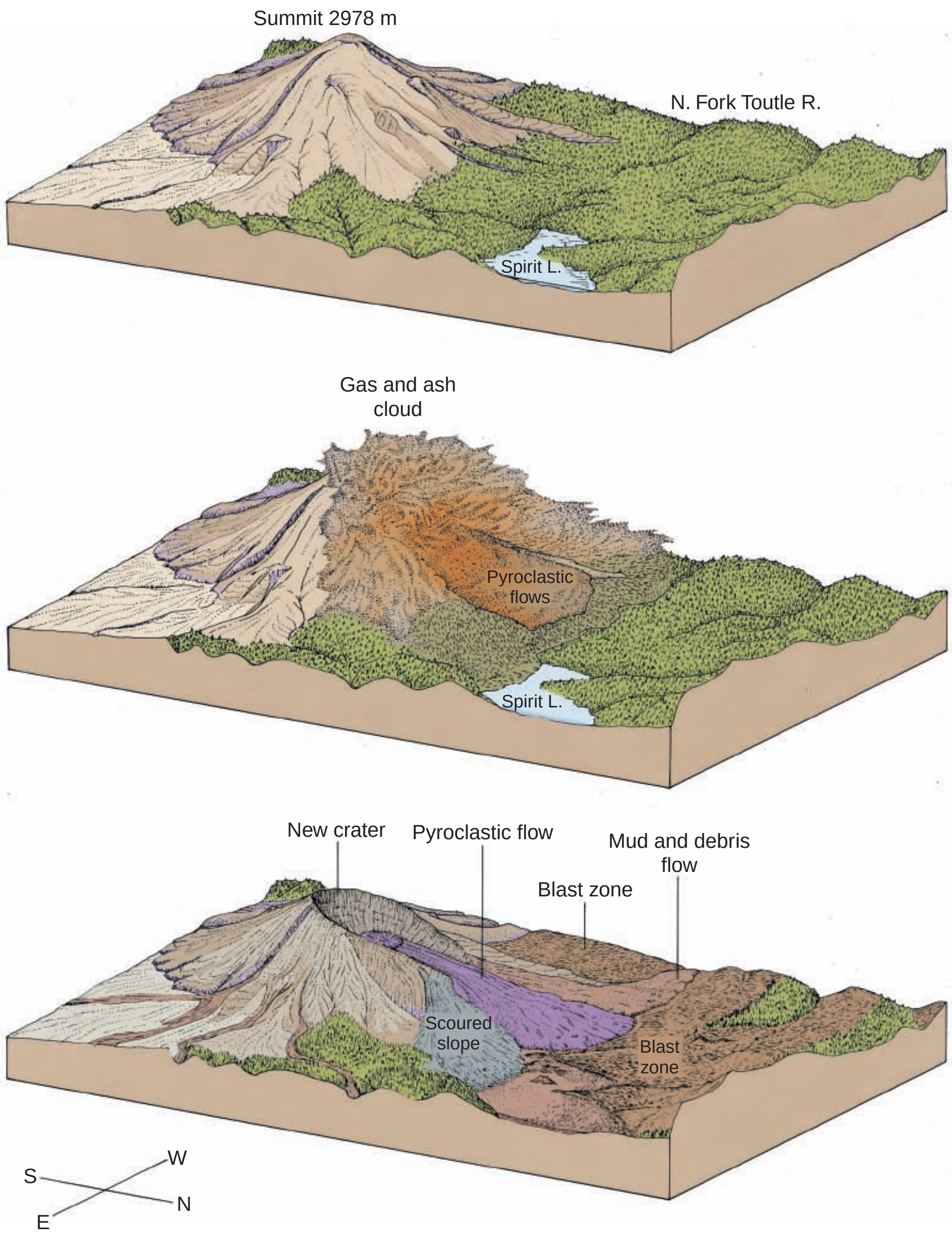


FIGURE 21.25 **The sequence of events in the eruption of Mount St. Helens**. *(Modified from Geo-Graphics, Portland, Oregon)*

On Sunday morning, May 18, the mountain was silent. Only minor plumes of steam rose from two vents. David Johnston, a 30-year-old geologist, was monitoring gas emissions and observing 8 km northwest of the volcano's crater. Abruptly, he cried over a two-way radio: "Vancouver! Vancouver! This is it!" Moments later, Johnston vanished in the blast of hot ash and gas as more than 4 km^3 of material was thrown from the blast on the north side of the mountain.

The best way to understand the nature of this eruption is to study the sequence in Figure 21.25. At 8:32 A.M., the mountain was shaken by an earthquake with a magnitude of approximately 5. The bulge on the north slope destabilized and then moved downslope as a great landslide. This uncapped the bottled-up magma and gas bubbles formed and explosively expanded in the low atmospheric pressure. Consequently, the eruption blasted horizontally across the collapsing slope. This lateral blast of rock, ash, and gas caused most of the destruction and loss of life. The blast wave leveled the forest in an area 35 km wide and 23 km outward on the mountain's north flank (Figure 21.25).

The eruption caused three separate, but interrelated, processes: (1) lahars, (2) ash flows, and (3) ash falls. The lahars originated largely from water-saturated ash on the mountain's upper slopes. Most of the lahars were hot. At the height of the flow, the lower Toutle River was heated to 90°C. The debris flows swept up 123 homes, as well as cars, logging trucks, and timber, and carried them downstream, destroying bridges and other constructions. So much sediment from the Toutle was carried into the Columbia River that the downstream depth of the Columbia was reduced from 12 m to 4 m within a day, and ships upstream were trapped.

An important part of the eruption was the extrusion of numerous ash flows. Traveling as fast as 130 km/hr, the incandescent ash and debris (at a temperature of about 500°C) extended northward a distance of 9 km. This was the same kind of devastating ash flow that came from Mont Pelée. Some ash flows reached Spirit Lake, where, together with debris flows that had been deposited earlier, they blocked the lake's outlet. Consequently, the level of water in the lake rose 60 m.

Immediately after the lateral blast, a vertical ash cloud rose to 18 km altitude. The ash cloud then fanned out downwind (eastward), and ash began to settle like a soft, gray snow. The cloud then moved in a broad arc across the United States and had completely circled Earth by June 5.

Other smaller ash-flow eruptions and lahars continued throughout the summer of 1980. The closing of the eruption episode was marked by the slow rise of magma through the central conduit to form a new lava dome in the summit crater.

A Summary of Volcanic Eruptions

The violent eruptions that characterize volcanoes at convergent plate boundaries result from the magma's high silica content, which makes it strong and viscous. Moreover, subduction zone magmas are rich in dissolved water. The dissolved gases cannot escape easily from the viscous melt. As a result, tremendous pressure builds up in the magma, and when eruptions occur, they are highly explosive. The eruptions produce huge quantities of ash, often in hot ash flows. Eruption of thick, viscous lava commonly precedes or follows the explosive activity. Tsunamis, lahars, and debris avalanches (caused by collapse of the sides of the volcanoes) are other significant hazards associated with steep stratovolcanoes like those formed at convergent plate margins.

METAMORPHISM AT CONVERGENT MARGINS

In the forearc of a subduction zone, metamorphism occurs at high-pressure–low-temperature conditions. In a magmatic arc, or in a zone of continental collision, metamorphism occurs at higher temperatures and lower pressures. Most metamorphic rocks in the continental crust were formed at convergent plate boundaries.

Why are metamorphic processes associated with plate convergence? Recall that metamorphism is driven by *changes* in a rock's environment, mainly changes in temperature and pressure. Because systems always seek equilibrium, these changes cause new minerals to form that are in equilibrium with the new conditions. At convergent plate boundaries, the unique tectonic and magmatic processes create dramatic temperature and pressure changes. Hence, metamorphism is a major process at convergent boundaries. Most metamorphic rocks in the continents were created at convergent plate boundaries.

At both modern and ancient subduction zones, two distinctive types of metamorphic rocks are juxtaposed to create **paired metamorphic belts** (Figure 21.26). The pair consists of an outer and an inner belt. The outermost metamorphic belt

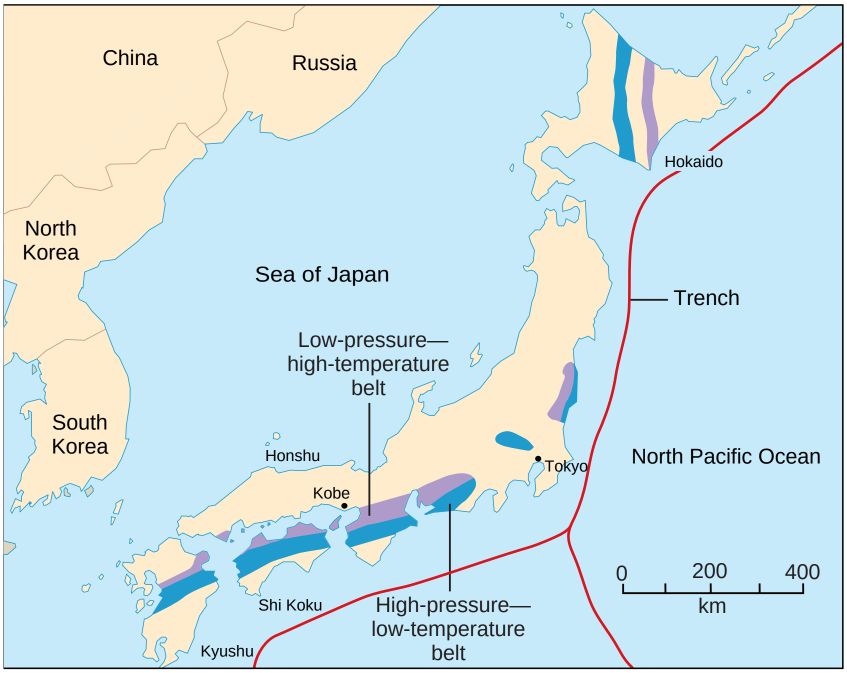


FIGURE 21.26 **Metamorphism at convergent plate margins** is an important process. Paired metamorphic belts formed in Japan during Mesozoic subduction. A high-pressure–low-temperature belt formed near the trench in the accretionary wedge and a belt of low- to intermediate-pressure metamorphism formed in the region of the magmatic arc.

forms in the accretionary wedge. Here fine-grained schists and slates contain the distinctive blue amphibole called glaucophane and other minerals indicative of the **blueschist facies** (Figures 21.27 and 6.16). These minerals form here because they are stable under the conditions unique to the forearc region: high pressure and relatively low temperature (less than 300°C). These unique high-pressure–low-temperature conditions are explained by the thermal structure of a subduction zone (Figure 21.6). Low temperatures result directly from the cold slab of subducting oceanic lithosphere. High pressure is attained because the slab drags cold oceanic rocks as deep as 30 to 50 km in the mantle, where the pressure is 10 to 15 kilobars. Once formed, blueschist facies rocks are brought rapidly back to the surface of the accretionary wedge by faulting, as described earlier. Belts of blueschist are found in Japan, California, New Zealand, and in the Alps, all sites of present or past plate convergence (Figure 21.1). Most of the blueschist metamorphic rocks are fragments of oceanic lithosphere, including chunks of pillow lava and gabbro. However, blueschists represent only a small fraction of the metamorphic rocks formed at convergent plate boundaries because of the very distinctive conditions required to form them. Serpentine from mantle peridotite is also present.

The innermost part of a paired metamorphic belt consists of rocks near the magmatic arc that recrystallized at higher temperatures and over a wide range of pressures or depths. In this zone of high heat flow (Figure 21.6), metamorphism is driven by heat released from cooling magmas that intruded to form plutonic belts. Where plutons intrude into sedimentary or other rocks near the surface, narrow contact metamorphic aureoles form, and hornfels dominates. Wider zones of orogenic metamorphism develop at moderate depths. In these belts, thrusting and folding also bury rocks to great depths where they become hotter. The metamorphic rocks are typified by mineral assemblages of the greenschist and amphibolite facies (see Figure 6.16). These are the most common kind of metamorphic rocks found on the continents.

Because of the higher temperatures, intensive plastic deformation accompanies orogenic metamorphism at convergent plate margins. The original sedimentary and volcanic rocks become strongly foliated schists and gneisses. The horizontal

FIGURE 21.27 Blueschist belts form by high-pressure–low-temperature metamorphism in accretionary wedges near subducting plates of oceanic lithosphere. The irregular blocks of blueschist shown here are part of a melange formed in a Paleozoic subduction zone along the east coast of New Zealand. The blocks are metamorphosed fragments of oceanic crust.

stresses generated by the converging plates cause foliation that is perpendicular to the direction of stress. Hence, slaty cleavage, schistosity, and gneissic layering in the deeper parts of a mountain range are characteristically vertical or dip at high angles. In a few collision zones, sedimentary rocks containing carbon were thrust so deeply (more than 100 km) that diamond formed from the carbon. Diamonds like this have been found in China, the Alps, Kazakhstan, and Norway. However, it is the depth of the rock, not the horizontal pressure from collision, that produces the pressure that drives recrystallization.

In the deeper parts of a mountain belt, metamorphism can become intense enough to produce migmatite—a complex mixture of thin layers of once-molten granitic material sandwiches between sheets of schist or gneiss. It develops largely from the partial melting of preexisting rocks. In these zones, high temperatures and pressures soften the entire rock body, which then behaves like a highly viscous liquid if it is subjected to stress. As a result, metamorphic rocks in deeper parts of orogenic belts exhibit complex flow structures (see Figure 6.11). Migmatites are probably the sources of some granitic magma.

FORMATION OF CONTINENTAL CRUST

Continents grow by accretion at convergent plate boundaries. New continental crust is created when silicic magma is added to deformed and metamorphosed rock in a mountain belt.

We have already seen how orogenic belts form at convergent boundaries. Strong deformation of rocks on a continental margin occurs during subduction and collision. These rocks are metamorphosed and intruded by silicic magmas that have low densities. Portions of this magma are extracted from the mantle or from subducted oceanic crust and form new additions to the continental crust. By virtue of their low densities, these rocks cannot be subducted and must remain in the continental crust. The gradual growth of the continents by addition of magma and deformation of preexisting crust is known as **continental accretion.** Consequently, both the origin and the evolution of Earth's continental crust are intimately tied to the processes that occur at convergent plate margins.

STATE OF THE ART

Magnetic Maps Show Accreted Terranes

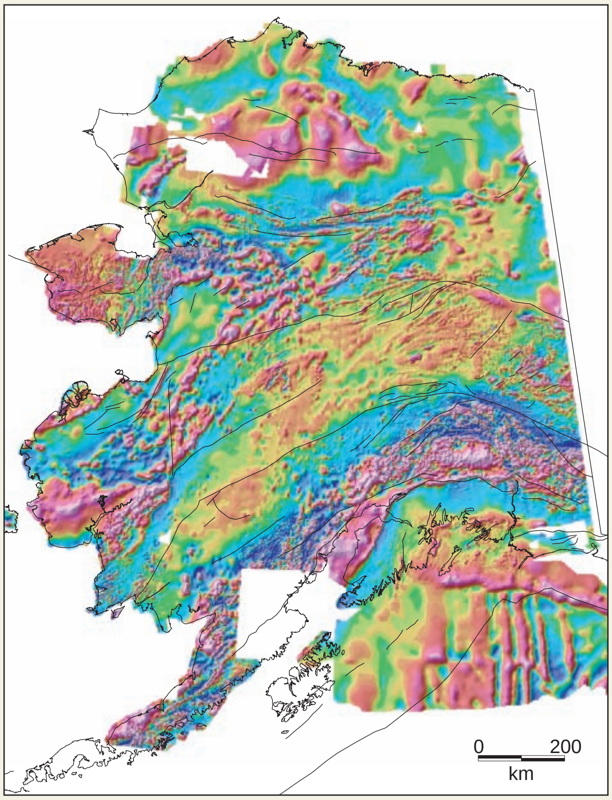
Geologists are always trying to find ways to see below the obscuring skin of soil and vegetation and discover the nature of the rocks hidden below. Geophysical techniques provide tools that can sense important characteristics of rocks beneath such a cover and for some depth into the interior. The strength of Earth's magnetic field is not uniform. Its strength varies because of the changing pattern of flow inside Earth's core, because of changes in the solar wind, and, of importance here, due to differences in the magnetic properties of rocks in the crust.

Magnetic minerals (such as magnetite) in rocks cause distortions in the magnetic field. Magnetic minerals induce local magnetic fields that either add or subtract from Earth's field. *Magnetic susceptibility* is a measure of how much magnetism can be induced in a rock. In general, sedimentary and most metamorphic rocks have relatively low magnetic susceptibilities. Igneous rocks, on the other hand, tend to have more magnetite and to be more magnetic.

(Courtesy of K. Nyman, Geological Survey of Finland)

By measuring the strength of the magnetic field at a multitude of individual spots, a map of the distribution of magnetic and nonmagnetic materials in the crust forms a powerful interpretive tool for geologists. For land-based magnetic surveys, the most commonly used *magnetometer* is the proton precession magnetometer that measures only the strength of Earth's magnetic field, not its polarity. This kind of magnetometer contains a cylinder filled with water that is surrounded by a coil of conductive wire. The hydrogen nuclei (protons) in the water behave like tiny spinning dipole magnets. Because Earth's magnetic field applies a force to these protons, they begin to precess or wobble as they spin. This precession induces a small but measurable current in the coil. The frequency of this current correlates with the strength of the local magnetic field. Airborne magnetometers are usually towed behind aircraft or mounted on wing tips. By repeatedly flying back and forth across a region, a magnetic map is constructed. Line spacings may be as little as 200 m, but more typically they are a kilometer or so apart.

An aeromagnetic map of Alaska shows the power of this technique. Much of Alaska's landscape is difficult to traverse. High mountains, glaciers, surging rivers, short field seasons, and hordes of mosquitoes make normal field geologic investigations difficult. A series of airborne magnetic surveys, however, can be conducted quickly and stitched together with a computer to see the magnetic fabric of the state and nearby offshore areas. The magnetic variations reveal the distribution of various rock types, ages, and the folded structures of the mountain belts. They define the boundaries of arcuate accreted terranes that were added to Alaska during millions of years of plate convergence. The aeromagnetic map also shows the striped fabric of the seafloor before it subducts down the oceanic trench.



(Courtesy of R. Saltus, U.S. Geological Survey)

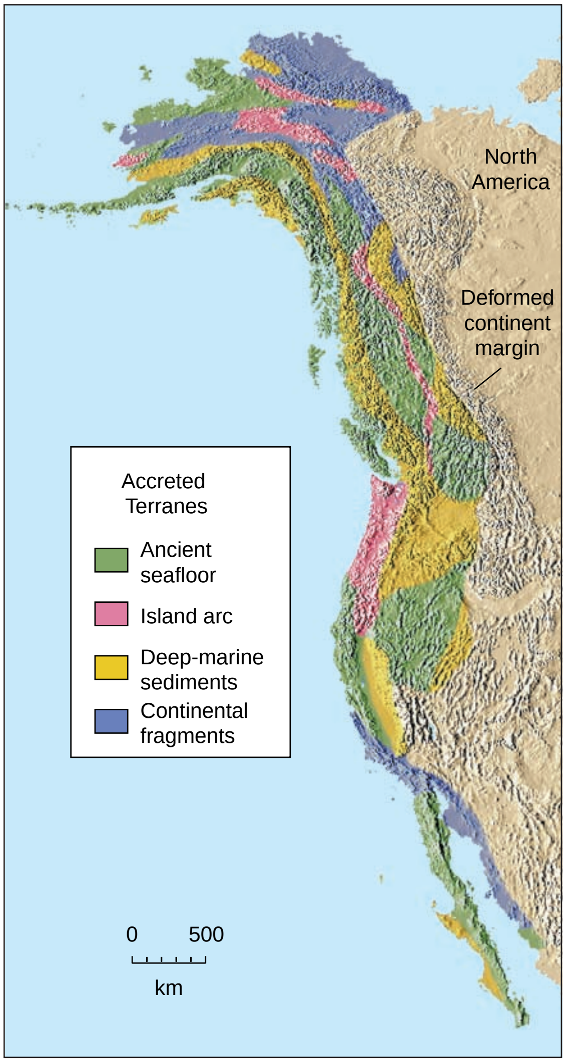


FIGURE 21.28 Accreted terranes along convergent plate margins are an important component of most continents. Western North America is composed of rocks that moved and became attached during episodes of convergence in the Mesozoic and Cenozoic eras. Before reaching their present positions, these rocks were island arcs, fragments of rifted continents, or oceanic plateaus. After accretion, they were shuffled along the margin by strike-slip faults.

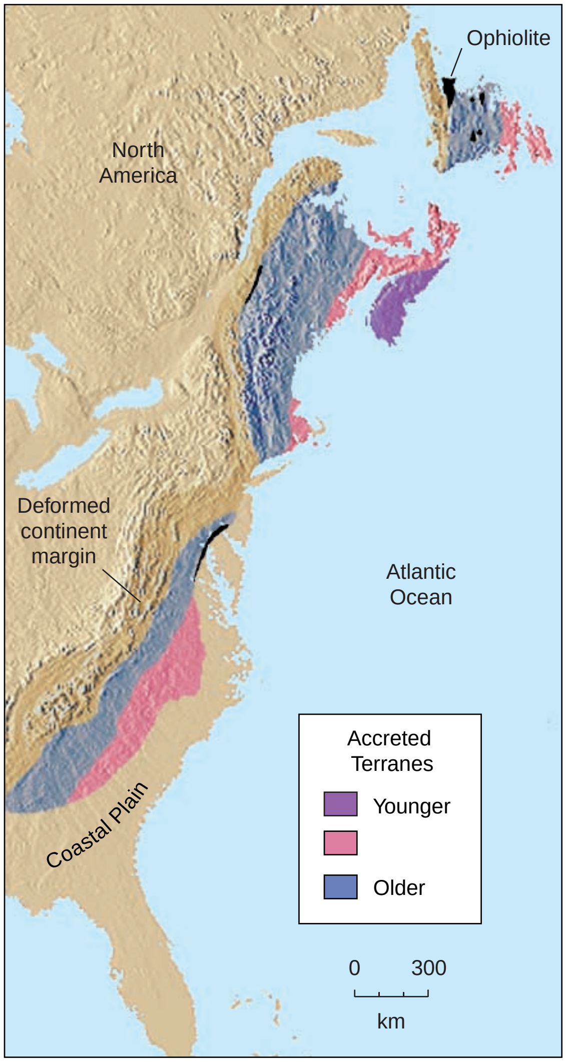


FIGURE 21.29 Accreted terranes form much of eastern North America. The Appalachian Mountains, which extend from Newfoundland to Alabama, contain terranes that were once parts of ancient Europe, Africa, island arcs, and even oceanic islands. These terranes were accreted to the continent during plate convergence and continental collision in the Paleozoic Era, millions of years before the western North American accretions.

Accreted Terranes

Studies worldwide reveal that many continental margins consist of a multitude of separate crustal blocks, each with its own distinctive origin and history. These blocks have been juxtaposed against one another by major faults. (Figure 21.28) Each block is a distinctive terrane, a term that refers to a region or group of rocks sharing a common age, structure, stratigraphy, and origin. These exotic segments of the orogenic belt are called **accreted terranes.** The terranes vary in size, and their rocks, fossils, histories, and magnetic properties contrast sharply. Fossils indicate that each terrane formed at different times and in very different environments than any other; paleomagnetic data show that the various terranes originated at different latitudes thousands of kilometers away.

Why are accreted terranes a common feature in many mountain belts?

Accreted terranes in the orogenic belt of western North America are a prime example (Figure 21.28). In this region, many independent terranes are squeezed together, each with its own internal structure, rock types, and fossils. Each terrane

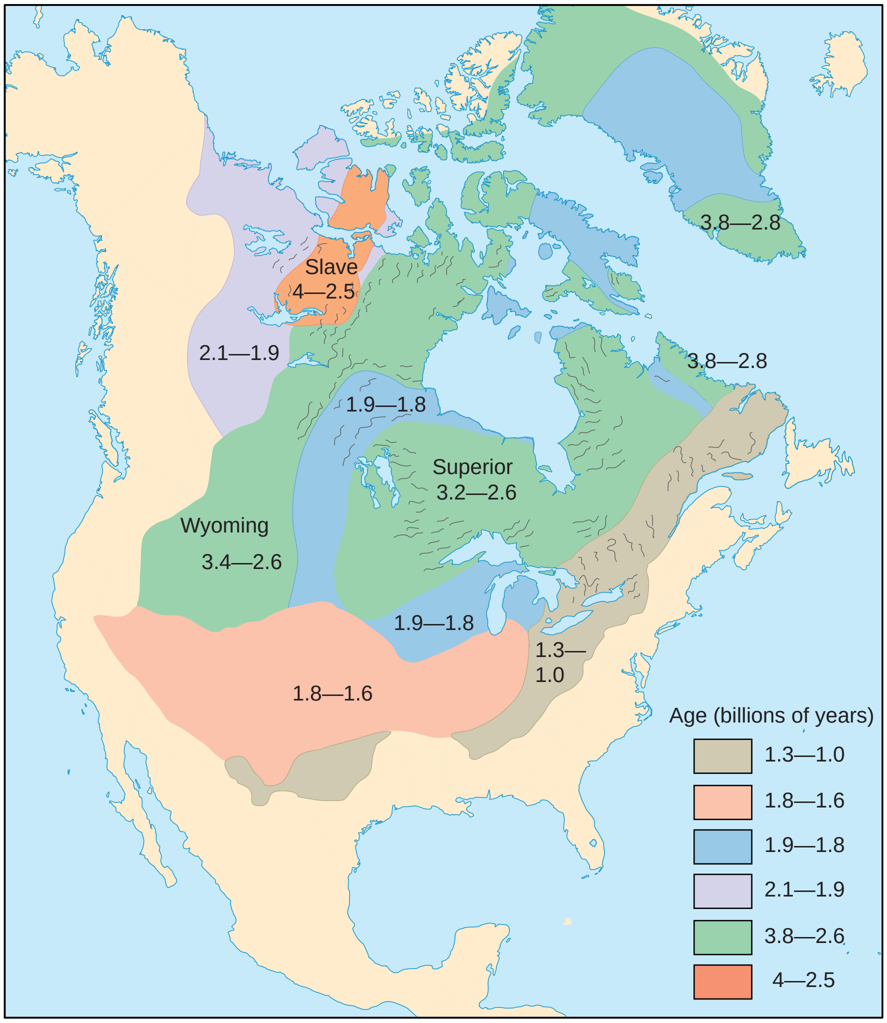


FIGURE 21.30 Radiometric ages of basement terranes in North America show several geologic provinces, each representing a mountain-building event. The ages of the major granitic intrusions are in billions of years, and the lines represent the trends of the folds and structural trends in the metamorphic rocks. The continent apparently grew by accretion as new mountain belts formed along its margins.

contrasts sharply with adjacent segments. One contains remnants of ancient seamounts related to mantle plumes; another, segments of shallow-marine limestones like those forming in the Bahamas; and still others are pieces of old volcanic arcs formed by subduction. Some are crustal fragments of metamorphosed basement complexes. Each terrane is separated from the next by long major fault systems, most of which show evidence of strike-slip movement. For example, paleomagnetic properties and fossils suggest that many of the terranes in British Columbia originated far to the south and traveled hundreds of kilometers northward, along the continental margin during the Mesozoic Era

The Appalachian Mountains, which stretch from Newfoundland to Alabama, are another example of continental accretion. They contain slices of ancient Europe, Africa, and oceanic islands accreted to the continent during the Paleozoic Era (Figure 21.29).

Accretion of North America

Evidence for even older orogenic and accretion events can be seen on the map of North America in Figure 21.30. It summarizes many radiometric ages for the metamorphic and granitic basement, as well as structural details compiled from years of field mapping and geophysical studies. The ages of the basement rocks of North America form a regular pattern. The oldest rocks found thus far are in northern Canada and are about 4 billion years old in the Slave Province (p. 204). Other ancient scraps of continental crust are in southern Greenland, the

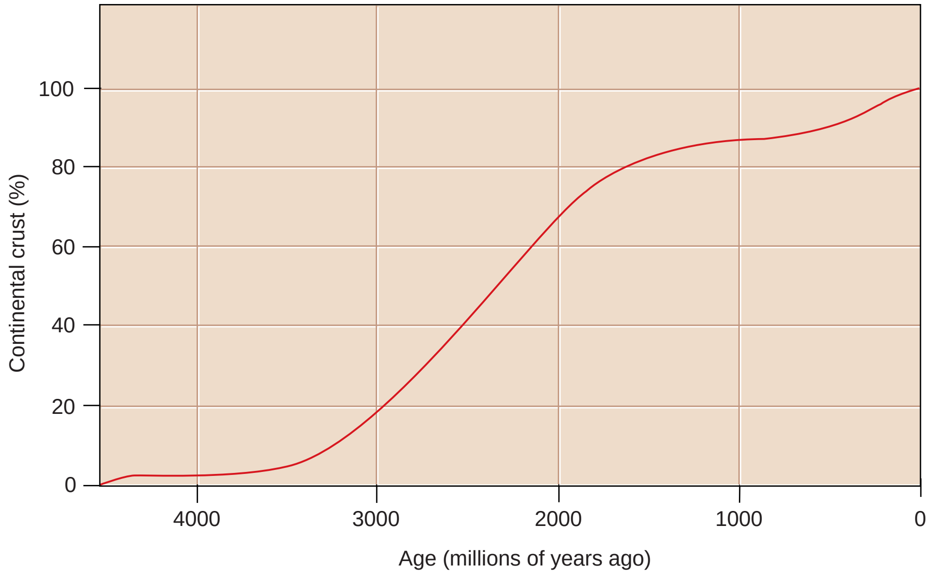


FIGURE 21.31 **The amount of continental crust has grown** over the last 4 billion years of Earth's history. The curve shows our best estimate of the rate of growth. During the last few billion years, the rate of growth was not as high as during the earlier history. Today, most continental crust forms at subduction zones.

Superior, and Wyoming Provinces, where granites and metamorphic rocks have ages between 2.5 and about 3.8 billion years.

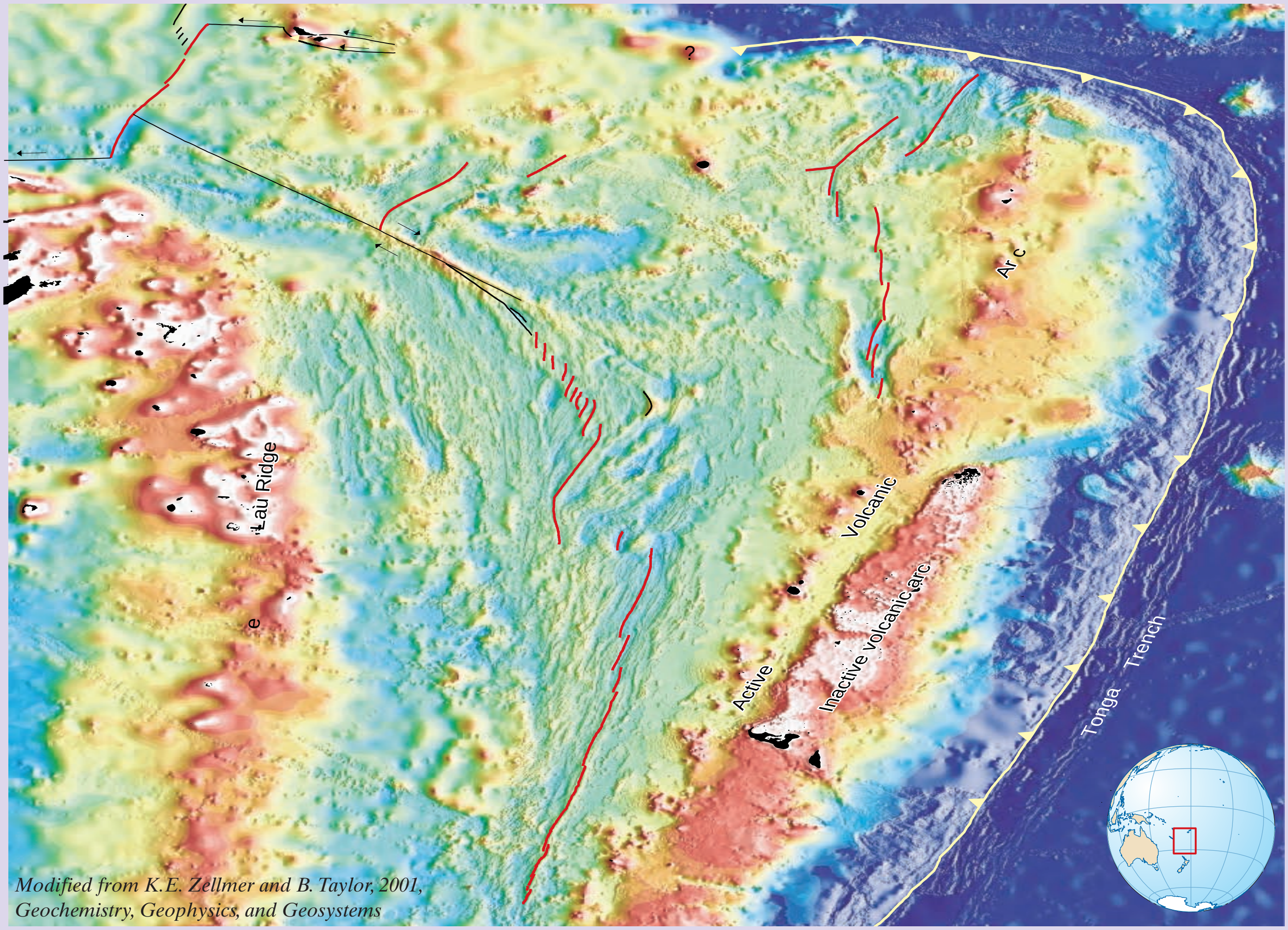
Surrounding the Superior Province to the south, west, and north is a vast area of gneiss and granite from 1.8 to 1.9 billion years old. In addition, its structural trends are oriented differently than in the older terrane. To the southeast, the rocks are younger still; the granitic intrusions and metamorphic rocks are as young as 1.0 billion years old.

Continental Growth Rates

The crudely concentric pattern of the basement age provinces in North America is strong evidence that the continent grew by the accretion of material around its margins during a series of mountain-building events. Each province probably represents a period of relatively rapid crustal growth during mountain building, related to convergent margin tectonics and magma production. The growth of North America is typical of other continents. Although the details are still sketchy, the amount of continental crust must have gradually grown during Earth's long history. Geological studies are starting to reveal the rate of growth of the continents (Figure 21.31). Continental crust grew slowly for the first billion years of Earth's history. Initially, much of the crust may have been swept back into the mantle. As Earth matured, its continents appear to have grown rapidly between about 3.5 and 1.5 billion years ago. Subsequently, growth of continental crust was slower. During both of these time periods, however, subduction and plate tectonics seem to be the principal cause of crustal growth.

Thus, the origin of continental crust can be related to the evolution of convergent plate margins. In an analogous manner, the origin and evolution of the oceanic crust is intimately connected to the processes at divergent plate boundaries.

GeoLogic   *Lau Basin: A Back Arc Basin*



Modified from K.E. Zellmer and B. Taylor, 2001, Geochemistry, Geophysics, and Geosystems

The Tonga-Kermadec arc of the southwest Pacific is like many other arcs related to subduction. A deep narrow trench lies to the east of the arc. Subduction carries the Pacific ocean lithosphere beneath the arc and is marked by an inclined zone of intense earthquakes. Explosive andesitic volcanoes dot the length of the arc, but many of the active volcanoes are young and still below sea level. However, unlike some arcs, a broad submarine basin—the Lau Basin—lies west of the island arc. What could have formed this basin?

Observations

1. The broad basin is corrugated with narrow ridges and valleys that more or less parallel the trend of the arc.
2. A narrower, slightly deeper valley runs down the center of the basin and forms a rift (marked in red).
3. Rocks along the submarine rift are mostly basaltic volcanic rocks.
4. The central rift is normally magnetized and rocks on the flanks are reversely magnetized.
5. Shallow earthquakes are clustered along the central valley, but are not as numerous as the deeper earthquakes along the subducting slab.
6. Hot springs are aligned along the central rift valley.
7. Subducting oceanic lithosphere lies about 300 km below the central part of the basin.
8. A high volcanic plateau also marks the western side of the basin. It is made of inactive andesitic volcanoes.

Interpretations

Geologists think the Lau Basin is a back-arc basin that formed by rifting apart an older volcanic arc. The Tonga ridge and the Lau Ridge mark the flanks of this rift. The youngest, most active part of the basin is a rift-valley, marked by extension, volcanism, shallow earthquakes, and hydrothermal vents. It is much like an oceanic ridge because new crust is forming here. The backarc extension may be caused by convection driven by the subducting slab. A new volcanic arc is developing just west of the old high arc. Most of the volcanoes in this new arc are still submarine.

KEY TERMS

accreted terrane (p. 627)
accretionary wedge (p. 606)
backarc (p. 599)
backarc spreading (p. 612)
blueschist facies (p. 624)
continental accretion (p. 625)
continent-continent convergence (p. 600)
convergent plate boundary (p. 598)
dehydrate (p. 616)
folded mountain belt (p. 599)
forearc ridge (p. 599)
inclined seismic zone (p. 604)
melange (p. 606)
nappe (p. 608)
ocean-continent convergence (p. 599)
ocean-ocean convergence (p. 598)
orogenic belt (p. 608)
orogenic metamorphism (p. 600)
outer swell (p. 599)
paired metamorphic belt (p. 623)
subduction zone (p. 599)
trench (p. 599)
volcanic arc (p. 599)

REVIEW QUESTIONS

1. Draw a simple cross section across a subduction zone showing the outer swell, trench, accretionary wedge, and magmatic arc.
2. Contrast the features of a continent-continent convergent plate boundary with those of an ocean-ocean plate boundary.
3. Why are trenches so deep?
4. Where do you expect to find the highest heat flow at a subduction zone? Where would you find the lowest gravity anomalies? Explain.
5. Describe how an accretionary wedge grows. How do the highly deformed rocks in the wedge differ from those in a folded mountain belt?
6. Is extension ever found at convergent plate boundaries?
7. Describe the absolute plate motions along the convergent plate boundaries: Australian-Indonesian, Indian-Eurasian, North American-Pacific, Nazca-South America. How do you think these differences are reflected in the nature of the plate boundary?
8. Explain how magma is generated at subduction zones and contrast that with magma generation at a midoceanic ridge.
9. What explains the origin of paired metamorphic belts at ocean-continent convergence zones?
10. Contrast the type of metamorphism that is found at convergent plate boundaries with metamorphism that occurs at divergent plate boundaries.
11. In which kind of convergent plate boundary would you expect to find the strongest compressional deformation—an ocean-ocean boundary, an ocean-continent boundary, or a continent-continent boundary? Why?
12. Why are earthquakes found deep (more than 300 km) in the mantle only at convergent plate boundaries?
13. Give two reasons why volcanoes at convergent plate margins are so explosive.
14. How would you discriminate an ancient accretionary wedge from the sediments that form on a rifted plate margin?
15. What would a slice of oceanic crust found in the middle of a continent imply about the tectonic history?
16. What are the characteristic sedimentary rocks formed along convergent plate margins? How do they compare with those found in a continental rift?
17. What processes lead to the growth of continents?
18. The average chemical composition of the continental crust corresponds to that of an andesite. Can you explain this observation?

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MULTIMEDIA TOOLS

Earth's Dynamic Systems Website
The Companion Website at **www.prenhall.com/hamblin** provides you with an on-line study guide and additional resources for each chapter, including:

- On-line Quizzes (Chapter Review, Visualizing Geology, Quick Review, Vocabulary Flash Cards) with instant feedback
- Quantitative Problems
- Critical Thinking Exercises
- Web Resources

Earth's Dynamic Systems CD
Examine the CD that came with your text. It is designed to help you visualize and thus understand the concepts in this chapter. It includes:

- Animations of plate motions at convergent margins
- A *flyover* of Earth's major convergent boundaries
- Slide shows of folded mountain belts and island arcs
- A direct link to the Companion Website