1

Plate Tectonics

In this Chapter

The plate tectonic model provides a framework for understanding many geodynamic processes. Earthquakes, volcanism, and mountain building are examples. The plate velocities, 10–100 mm yr^{-1}, imply a fluid-like behavior of the solid Earth. Hot mantle rock can flow (behave as a fluid) on geological time scales due to solid-state creep and thermal convection. The hot mantle rock is cooled by heat loss to the Earth's surface resulting in a *cold thermal "boundary layer."* This boundary layer is rigid and is referred to as the lithosphere. The surface *lithosphere* is broken into a series of plates that are in relative motion with respect to each other. This motion results in "*plate tectonics*."

Plates are created at *mid-ocean ridges*, where hot mantle rock ascends. Partial melting in the ascending rock produces the magmas that form the *basaltic ocean crust*. The surface plates reenter the mantle at *ocean trenches* (*subduction*). The cold rock in the plate (lithosphere) is denser than the adjacent hot mantle rock. This results in a downward gravitational body force that drives the motion of the surface plate. Complex volcanic processes at subduction zones generate the *continental crust*. This crust is thick and light and does not participate in the plate-tectonic cycle. Thus the continental crust is about a factor of 10 older, on average, than the oceanic crust (1 Ga versus 100 Ma).

Interactions between plates at plate boundaries are responsible for a large fraction of the *earthquakes* that occur. Earthquakes are caused by episodic ruptures and displacements on preexisting faults. These displacements provide the relative motions between surface plates. Plate boundary processes are also responsible for a large fraction of the surface *volcanism*.

However, surface volcanism also occurs within plate interiors. At least a fraction of this volcanism can be attributed to *mantle plumes* that impinge on the base of the lithosphere. Mantle plumes are thin conduits of hot solid mantle rock that ascend from great depths.

One important consequence of plate tectonics is *continental drift*. Oceans open and close. The western and eastern boundaries of the Atlantic Ocean fit together like a jigsaw puzzle. New oceans are created at rifts in the continental crust. An example of a young ocean is the Red Sea. Oceans also close, resulting in *continental collisions*. An example is the Himalaya mountain belt that is the result of a continental collision between India and Asia.

A major goal of this book is to provide a fundamental understanding of why our planet has plate tectonics. Heat is being produced within the Earth due to the decay of radioactive isotopes. The interior of the Earth is hot and its surface is cold. The hot rock is less dense than the cold rock, leading to a gravitational instability. Because the hot mantle behaves as a fluid on geological time scales, this instability causes thermal convection. The plate tectonic cycle is one consequence of thermal convection in the Earth's mantle.

We also discuss *comparative planetology* in this chapter. Our knowledge of the structure and behavior of the other terrestrial planets, Mercury, Venus, and Mars, as well as major planetary satellites, the Moon and the satellites of Jupiter and Saturn, is summarized. Two important examples are the constraints on the early solar system obtained from lunar samples returned by the Apollo missions and the lack of plate tectonics on Venus. Considering the similarities in composition and size between the Earth and Venus, the absence of plate tectonics on Venus is a surprise.

1.1 Introduction

Plate tectonics is a model in which the outer shell of the Earth is divided into a number of thin, rigid plates that are in relative motion with respect to one another. The relative velocities of the plates are of the order of a few tens of millimeters per year. A large fraction of all earthquakes, volcanic eruptions, and mountain building occurs at plate boundaries. The distribution of the major surface plates is illustrated in Figure 1.1.

The plates are made up of relatively cool rocks and have an average thickness of about 100 km. The plates are being continually created and consumed. At ocean ridges adjacent plates diverge from each other in a process known as *seafloor spreading*. As the adjacent plates diverge, hot mantle rock ascends to fill the gap. The hot, solid mantle rock behaves like a fluid because of solid-state creep processes. As the hot mantle rock cools, it becomes rigid and accretes to the plates, creating new plate area. For this reason ocean ridges are also known as *accreting plate boundaries*. The accretionary process is symmetric to a first approximation so that the rates of plate formation on the two sides of a ridge are approximately equal. The rate of plate formation on one side of an ocean ridge defines a half-spreading velocity u. The two plates spread with a relative velocity of $2u$. The global system of ocean ridges is denoted by the heavy dark lines in Figure 1.1.

Because the surface area of the Earth is essentially constant, there must be a complementary process of plate consumption. This occurs at *ocean trenches*. The surface plates bend and descend into the interior of the Earth in a process known as *subduction*. At an ocean trench the two adjacent plates converge, and one descends beneath the other. For this reason ocean trenches are also known as *convergent plate boundaries*. The worldwide distribution of trenches is shown in Figure 1.1 by the lines with triangular symbols, which point in the direction of subduction.

A cross-sectional view of the creation and consumption of a typical plate is illustrated in Figure 1.2. That part of the Earth's interior that comprises the plates is referred to as the *lithosphere*. The rocks that make up the lithosphere are relatively cool and rigid; as a result, the interiors of the plates do not deform significantly as they move about the surface of the Earth. As the plates move away from ocean ridges, they cool and thicken. The solid rocks beneath the lithosphere are sufficiently hot to be able to deform freely; these rocks comprise the *asthenosphere*, which lies below the lithosphere. The lithosphere slides over the asthenosphere with relatively little resistance.

As the rocks of the lithosphere become cooler, their density increases because of thermal contraction. As a result, the lithosphere becomes gravitationally unstable with respect to the hot asthenosphere beneath. At the ocean trench the lithosphere bends and sinks into the interior of the Earth because of this negative buoyancy. The downward gravitational body force on the descending lithosphere plays an important role in driving plate tectonics. The lithosphere acts as an elastic plate that transmits large elastic stresses without significant deformation. Thus the gravitational body force can be transmitted directly to the surface plate and this force pulls the plate toward the trench. This body force is known as *trench pull*. Major faults separate descending lithospheres from adjacent overlying lithospheres. These faults are the sites of most great earthquakes. Examples are the Chilean earthquake in 1960, the Alaskan earthquake in 1964,

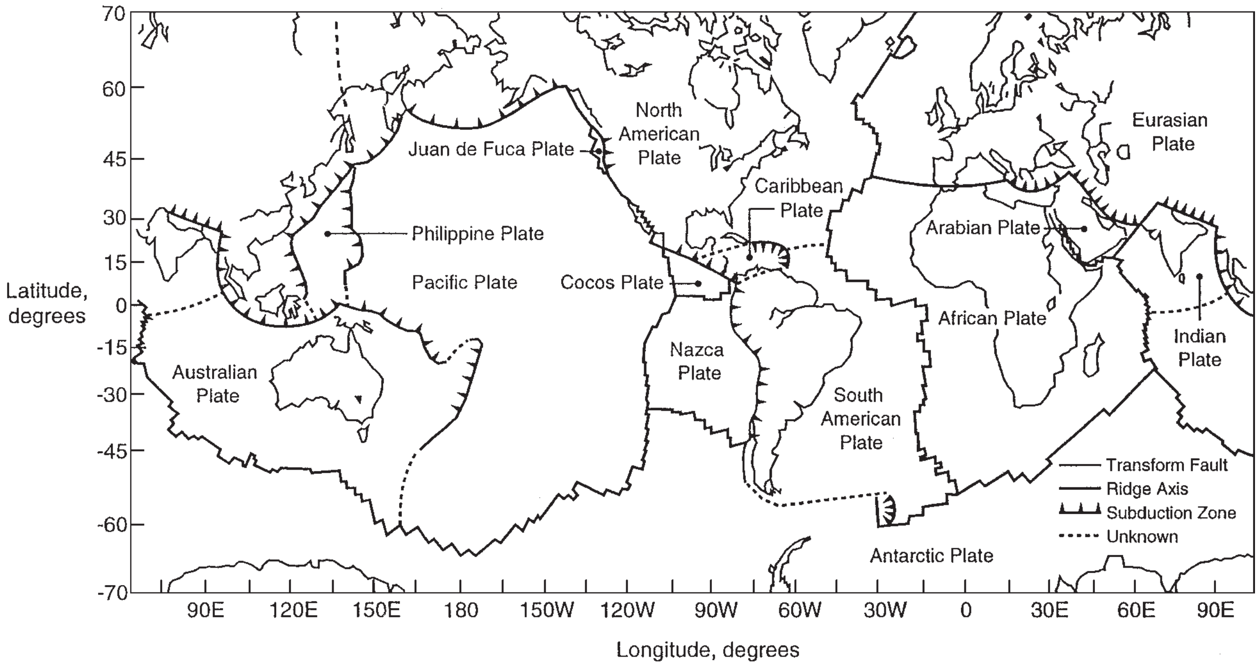

Figure 1.1 Distribution of the major plates. The ocean ridge axis (accretional plate margins), subduction zones (convergent plate margins), and transform faults that make up the plate boundaries are shown.

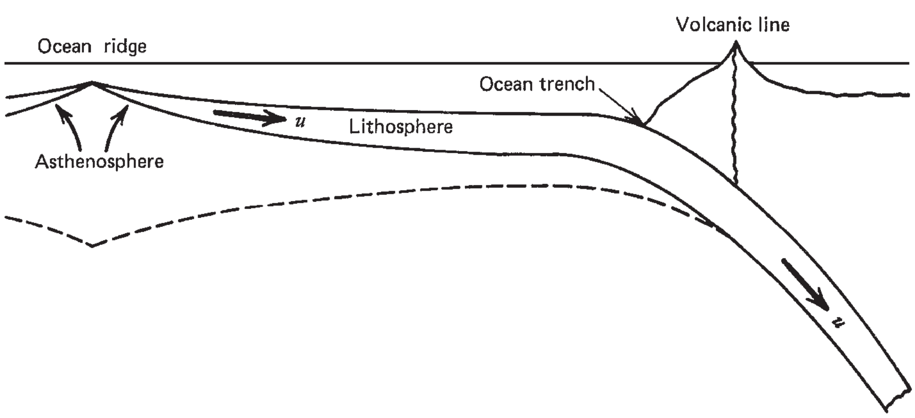

Figure 1.2 Accretion of a lithospheric plate at an ocean ridge and its subduction at an ocean trench. The asthenosphere, which lies beneath the lithosphere, is shown along with the line of volcanic centers associated with subduction.

the Sumatra earthquake in 2004, and the Tohoku, Japan, earthquake in 2011. These are the largest earthquakes that have occurred since modern seismographs have been available. The locations of the descending lithospheres can be accurately determined from the earthquakes occurring in the cold, brittle rocks of the lithospheres. These planar zones of earthquakes associated with subduction are known as *Wadati–Benioff zones*.

Lines of active volcanoes lie parallel to almost all ocean trenches. These volcanoes occur about 125 km above the descending lithosphere. At least a fraction of the magmas that form these volcanoes are produced near the upper boundary of the descending lithosphere and rise some 125 km to the surface. If these volcanoes stand on the seafloor, they form an *island arc*, as typified by the Aleutian Islands in the North Pacific. If the trench lies adjacent to a continent, the volcanoes grow from the land surface. This is the case in the western United States, where a volcanic line extends from Mount Baker in the north to Mount Shasta in the south. Mount St. Helens, the site of a violent eruption in 1980, forms a part of this volcanic line. These volcanoes are the sites of a large fraction of the most explosive and violent volcanic eruptions. The eruption of Mount Pinatubo in the Philippines in 1991, the most violent eruption of the twentieth century, is another example. A typical subduction zone volcano is illustrated in Figure 1.3.

The Earth's surface is divided into continents and oceans. The oceans have an average depth of about 4 km, and the continents rise above sea level. The reason for this difference in elevation is the difference in the thickness of the crust. Crustal rocks have a different composition from that of the mantle rocks beneath and are less dense. The crustal rocks are therefore gravitationally stable with respect to the heavier mantle rocks. There is usually a well-defined boundary, the *Moho* or Mohorovičić discontinuity, between the crust and mantle. A typical thickness for *oceanic crust* is 6 km; *continental crust* is about 35 km thick. Although oceanic crust is gravitationally stable, it is sufficiently thin so that it does not significantly impede

Figure 1.3 Izalco volcano in El Salvador, an example of a subduction zone volcano (NOAA–NGDC Howell Williams).

the subduction of the gravitationally unstable oceanic lithosphere. The oceanic lithosphere is continually cycled as it is accreted at ocean ridges and subducted at ocean trenches. Because of this cycling the average age of the ocean floor is about 10^8 years (100 Ma).

On the other hand, the continental crust is sufficiently thick and gravitationally stable so that it is not subducted at an ocean trench. In some cases the denser lower continental crust, along with the underlying gravitationally unstable continental mantle lithosphere, can be recycled into the Earth's interior in a process known as *delamination*. However, the light rocks of the upper continental crust remain in the continents. For this reason the rocks of the continental crust, with an average age of about 2×10^9 years (2 Ga), are much older than the rocks of the oceanic crust. As the lithospheric plates move across the surface of the Earth, they carry the continents with them. The relative motion of continents is referred to as *continental drift*.

Much of the historical development leading to plate tectonics concerned the validity of the hypothesis of continental drift: that the relative positions of continents change during geologic time. The similarity in shape between the west coast of Africa and the east coast of South America was noted as early as 1620 by Francis Bacon. This "fit" has led many authors to speculate on how these two continents might have been attached. A detailed exposition of the hypothesis of continental drift was put forward by Frank B. Taylor (1910). The hypothesis was further developed by Alfred Wegener beginning in 1912 and summarized in his book *The Origin of Continents and Oceans* (Wegener, 1946). As a meteorologist, Wegener was particularly interested in the observation that glaciation had occurred in equatorial regions at the same time that tropical conditions prevailed at high latitudes. This observation in itself could be explained by *polar wander*, a shift of the rotational axis without other surface deformation. However, Wegener also set forth many of the qualitative arguments that the continents had formerly been attached. In addition to the observed fit of continental margins, these arguments included the correspondence of geological provinces, continuity of structural features such as relict mountain ranges, and the correspondence of fossil types. Wegener argued that a single supercontinent, Pangaea, had formerly existed. He suggested that tidal forces or forces associated with the rotation of the Earth were responsible for the breakup of this continent and the subsequent continental drift.

Further and more detailed qualitative arguments favoring continental drift were presented by Alexander du Toit, particularly in his book *Our Wandering Continents* (du Toit, 1937). Du Toit argued that instead of a single supercontinent, there had formerly been a northern continent, Laurasia, and a southern continent, Gondwanaland, separated by the Tethys Ocean.

During the 1950s extensive exploration of the seafloor led to an improved understanding of the worldwide range of mountains on the seafloor known as mid-ocean ridges. Harry Hess (1962) hypothesized that the seafloor was created at the axis of a ridge and moved away from the ridge to form an ocean in a process now referred to as *seafloor spreading*. This process explains the similarity in shape between continental margins. As a continent breaks apart, a new ocean ridge forms. The ocean floor created is formed symmetrically at this ocean ridge, creating a new ocean. This is how the Atlantic Ocean was formed; the Mid-Atlantic Ridge where the ocean formed now bisects the ocean.

It should be realized, however, that the concept of continental drift won general acceptance by Earth scientists only in the period between 1967 and 1970. Although convincing qualitative, primarily geological, arguments had been put forward to support continental drift, almost all Earth scientists and, in particular, almost all geophysicists had opposed the hypothesis. Their opposition was mainly based on arguments concerning the rigidity of the mantle and the lack of an adequate driving mechanism.

The propagation of seismic shear waves showed beyond any doubt that the mantle was a solid. An essential question was how horizontal displacements of thousands of kilometers could be accommodated by solid rock. The fluidlike behavior of the Earth's mantle had been established in a general way by gravity studies carried out in the latter part of the nineteenth century. Measurements showed that mountain ranges had low-density roots. The lower density of the roots provides a negative relative mass that nearly equals the positive mass of the mountains. This behavior could be explained by the principle of *hydrostatic equilibrium* if the mantle behaved as a fluid. Mountain ranges appear to behave similarly to blocks of wood floating on water.

The fluid behavior of the mantle was established quantitatively by N. A. Haskell (1935). Studies of the elevation of beach terraces in Scandinavia showed that the Earth's surface was still rebounding from the load of the ice during the last ice age. By treating the mantle as a viscous fluid with a viscosity of 10^{20} Pa s, Haskell was able to explain the present uplift of Scandinavia. Although this is a very large viscosity (water has a viscosity of 10^{-3} Pa s), it leads to a fluid behavior for the mantle during long intervals of geologic time.

In the 1950s theoretical studies had established several mechanisms for the very slow creep of crystalline materials. This creep results in a fluid behavior. Robert B. Gordon (1965) showed that solid-state creep quantitatively explained the viscosity determined from observations of postglacial rebound. At temperatures that are a substantial fraction of the melt temperature, thermally activated creep processes allow mantle rock to flow at low stress levels on time scales greater than 10^4 years. The rigid lithosphere includes rock that is sufficiently cold to preclude creep on these long time scales.

The creep of mantle rock was not a surprise to scientists who had studied the widely recognized flow of ice in glaciers. Ice is also a crystalline solid, and gravitational body forces in glaciers cause ice to flow because its temperature is near its melt temperature. Similarly, mantle rocks in the Earth's interior are near their melt temperatures and flow in response to gravitational body forces.

Forces must act on the lithosphere in order to make the plates move. Wegener suggested that either tidal forces or forces associated with the rotation of the Earth caused the motion responsible for continental drift. However, in the 1920s Sir Harold Jeffreys, as summarized in his book *The Earth* (Jeffreys, 1924), showed that these forces were insufficient. Some other mechanism had to be found to drive the motion of the plates. Any reasonable mechanism must also have sufficient energy available to provide the energy being dissipated in earthquakes, volcanoes, and mountain building. Arthur Holmes (1931) hypothesized that thermal convection was capable of driving mantle convection and continental drift. If a fluid is heated from below, or from within, and is cooled from above in the presence of a gravitational field, it becomes gravitationally unstable, and thermal convection can occur. The hot mantle rocks at depth are gravitationally unstable with respect to the colder, denser rocks in the lithosphere. The result is thermal convection in which the colder rocks descend into the mantle and the hotter rocks ascend toward the surface. The ascent of mantle material at ocean ridges and the descent of the lithosphere into the mantle at ocean trenches are parts of this process. The Earth's mantle is being heated by the decay of the radioactive isotopes uranium 235 (^{235}U), uranium 238 (^{238}U), thorium 232 (^{232}Th), and potassium 40 (^{40}K). The volumetric heating from these isotopes and the *secular cooling* of the Earth drive mantle convection. The heat generated by the radioactive isotopes decreases with time as they decay. Two billion years ago the heat generated was about twice the present value. Because the amount of heat generated is less today, the vigor of the mantle convection required today to extract the heat is also less. The vigor of mantle convection depends on the mantle viscosity. Less vigorous mantle convection implies a higher viscosity. But the mantle viscosity is a strong function of mantle temperature; a higher mantle viscosity implies a cooler mantle. Thus as mantle convection becomes less vigorous, the mantle cools; this is secular cooling. As a result, about 80% of the heat lost from the interior of the Earth is from the decay of the radioactive isotopes and about 20% is due to the cooling of the Earth (secular cooling).

During the 1960s independent observations supporting continental drift came from paleomagnetic studies. When magmas solidify and cool, their iron component is magnetized by the Earth's magnetic field. This remanent magnetization provides a fossil record of the orientation of the magnetic field at that time. Studies of the orientation of this field can be used to determine the movement of the rock relative to the Earth's magnetic poles since the rock's formation. Rocks in a single surface plate that have not been deformed locally show the same position for the Earth's magnetic poles. Keith Runcorn (1956) showed that rocks in North America and Europe gave different positions for the magnetic poles. He concluded that the differences were the result of continental drift between the two continents.

Paleomagnetic studies also showed that the Earth's magnetic field has been subject to episodic reversals. Observations of the magnetic field over the oceans indicated a regular striped pattern of *magnetic anomalies* (regions of magnetic field above and below the average field value) lying parallel to the ocean ridges. Frederick Vine and Drummond Matthews (1963) correlated the locations of the edges of the striped pattern of magnetic anomalies with the times of magnetic field reversals and were able to obtain quantitative values for the rate of seafloor spreading. These observations have provided the basis for accurately determining the relative velocities at which adjacent plates move with respect to each other.

By the late 1960s the framework for a comprehensive understanding of the geological phenomena and processes of continental drift had been built. The basic hypothesis of plate tectonics was given by Jason Morgan (1968). The concept of a mosaic of rigid plates in relative motion with respect to one another was a natural consequence of thermal convection in the mantle. A substantial fraction of all earthquakes, volcanoes, and mountain building can be attributed to the interactions among the lithospheric plates at their boundaries (Isacks *et al.*, 1968). Continental drift is an inherent part of plate tectonics. The continents are carried with the plates as they move about the surface of the Earth.

Problem 1.1

If the area of the oceanic crust is 3.2×10^8 km^2 and new seafloor is now being created at the rate of 2.8 km^2 yr^{-1}, what is the mean age of the oceanic crust? Assume that the rate of seafloor creation has been constant in the past.

1.2 The Lithosphere

An essential feature of plate tectonics is that only the outer shell of the Earth, the *lithosphere*, remains rigid during intervals of geologic time. Because of their low temperature, rocks in the lithosphere do not significantly deform on time scales of up to 10^9 years. The rocks beneath the lithosphere are sufficiently hot so that solid-state creep can occur. This creep leads to a fluid-like behavior on geologic time scales. In response to forces, the rock beneath the lithosphere flows like a fluid.

The lower boundary of the lithosphere is defined to be an isotherm (surface of constant temperature). A typical value is approximately 1600 K. Rocks lying above this isotherm are sufficiently cool to behave rigidly, whereas rocks below this isotherm are sufficiently hot to readily deform. Beneath the ocean basins the lithosphere has a thickness of about 100 km; beneath the continents the thickness is about twice this value. Because the thickness of the lithosphere is only 2 to 4% of the radius of the Earth, the lithosphere is a thin shell. This shell is broken up into a number of plates that are in relative motion with respect to one another. This relative motion is primarily accommodated at plate boundaries: ocean ridges, ocean trenches, and transform faults. However, as the plates evolve in time broader zones of deformation are required. These zones of deformation are usually adjacent to the plate boundaries. Examples include the western United States and eastern China.

The rigidity of the lithosphere allows the plates to transmit elastic stresses during geologic intervals. The plates act as stress guides. Stresses that are applied at the boundaries of a plate can be transmitted throughout the interior of the plate. The ability of the plates to transmit stress over large distances has important implications with regard to the driving mechanism of plate tectonics.

The rigidity of the lithosphere also allows it to bend when subjected to a load. An example is the load applied by a volcanic island. The load of the Hawaiian Islands causes the lithosphere to bend downward around the load, resulting in a region of deeper water around the islands. The elastic bending of the lithosphere under vertical loads can also explain the structure of ocean trenches and some sedimentary basins.

However, the entire lithosphere is not effective in transmitting elastic stresses. Only about the upper half of it is sufficiently rigid so that elastic stresses are not relaxed on time scales of 10^9 years. This fraction of the lithosphere is referred to as the *elastic lithosphere*. Solid-state creep processes relax stresses in the lower, hotter part of the lithosphere. However, this part of the lithosphere remains a coherent part of the plates. A detailed discussion of the difference between the thermal and elastic lithospheres is given in Section 7.10.

1.3 Accreting Plate Boundaries

Lithospheric plates are created at ocean ridges. The two plates on either side of an ocean ridge move away from each other with near constant velocities of a few tens of millimeters per year. As the two plates diverge, hot mantle rock flows upward to fill the gap. The upwelling mantle rock cools by conductive heat loss to the surface. The cooling rock accretes to the base of the spreading plates, becoming part of them; the structure of an *accreting plate boundary* is illustrated in Figure 1.4.

As the plates move away from the ocean ridge, they continue to cool and the lithosphere thickens. The elevation of the ocean ridge as a function of distance from the ridge crest can be explained in terms of the temperature distribution in the lithosphere. As the lithosphere cools, it becomes denser; as a result it sinks downward into the underlying mantle rock. The topographic elevation of the ridge is due to the greater buoyancy of the thinner, hotter lithosphere near the axis of accretion at the ridge crest. The elevation of the ocean ridge also provides a body force that causes the plates to move away from the ridge crest. A component of the gravitational body force on the elevated lithosphere drives the lithosphere away from the accretional boundary; it is one of the important forces driving the plates. This force on the lithosphere is known as *ridge push* and is a form of *gravitational sliding*.

The volume occupied by the ocean ridge displaces seawater. Rates of seafloor spreading vary in time.

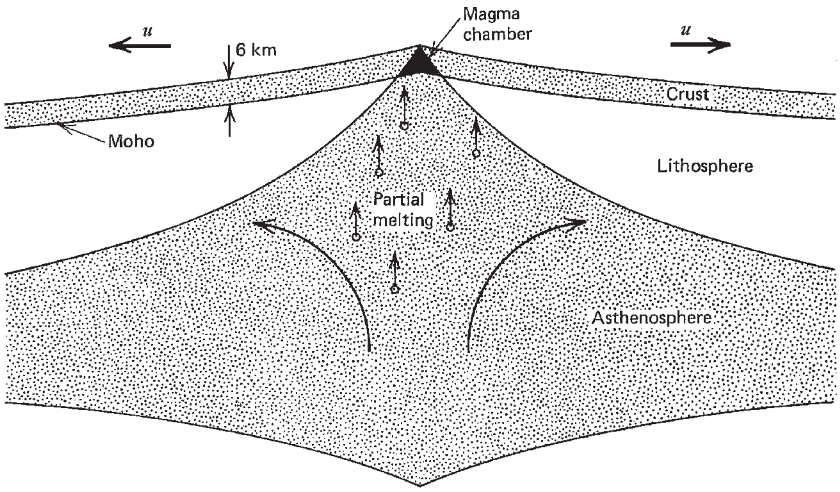

Figure 1.4 An accreting plate margin at an ocean ridge.

When rates of seafloor spreading are high, ridge volume is high, and seawater is displaced. The result is an increase in the global sea level. Variations in the rates of seafloor spreading are the primary cause for changes in sea level on geological time scales. In the Cretaceous (≈80 Ma) the rate of seafloor spreading was about 30% greater than at present and sea level was about 200 m higher than today. One result was that a substantial fraction of the continental interiors was covered by shallow seas.

Ocean ridges are the sites of a large fraction of the Earth's volcanism. Because almost all the ridge system is under water, only a small part of this volcanism can be readily observed. The details of the volcanic processes at ocean ridges have been revealed by exploration using submersible vehicles. Ridge volcanism can also be seen in Iceland, where the oceanic crust is sufficiently thick so that the ridge crest rises above sea level. The volcanism at ocean ridges is caused by *pressure-release melting*. As the two adjacent plates move apart, hot mantle rock ascends to fill the gap. The temperature of the ascending rock is nearly constant, but its pressure decreases. The pressure p of rock in the mantle is given by the simple hydrostatic equation

$$p = \rho g y, \tag{1.1}$$

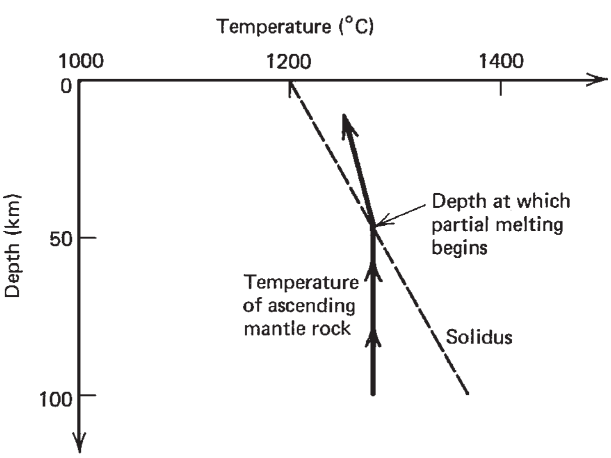

Figure 1.5 The process of pressure-release melting is illustrated. Melting occurs because the nearly isothermal ascending mantle rock encounters pressures low enough so that the associated solidus temperatures are below the rock temperatures.

where ρ is the density of the mantle rock, g is the acceleration of gravity, and y is the depth. The solidus temperature (the temperature at which the rock first melts) decreases with decreasing pressure. When the temperature of the ascending mantle rock equals the solidus temperature, melting occurs, as illustrated in Figure 1.5. The ascending mantle rock contains a low-melting-point, basaltic component. This component melts to form the oceanic crust.

Problem 1.2

At what depth will ascending mantle rock with a temperature of 1600 K melt if the equation for the solidus temperature T is

$$T(\mathrm{K}) = 1500 + 0.12p\ (\mathrm{MPa}).$$

Assume $\rho = 3300$ kg m^{-3}, $g = 10$ m s^{-2}, and the mantle rock ascends at constant temperature.

The *magma* (melted rock) produced by partial melting beneath an ocean ridge is lighter than the residual mantle rock, and buoyancy forces drive it upward to the surface in the vicinity of the ridge crest. Magma chambers form, heat is lost to the seafloor, and this magma solidifies to form the oceanic crust. In some localities slices of oceanic crust and underlying mantle have been brought to the surface. These are known as *ophiolites*; they occur in such locations as Cyprus, Newfoundland, Oman, and New Guinea. Field studies of ophiolites have provided a detailed understanding of the oceanic crust and underlying mantle. Typical oceanic crust is illustrated in Figure 1.6. The crust is divided into layers 1, 2, and 3, which were originally associated with different seismic velocities but subsequently identified compositionally. Layer 1 is composed of sediments that are deposited on the volcanic rocks of layers 2 and 3. The thickness of sediments increases with distance from the ridge crest; a typical thickness is 1 km. Layers 2 and 3 are composed of basaltic rocks of nearly uniform composition. A typical composition of an ocean basalt is given in Table 1.1. The basalt is composed primarily of two rock-forming minerals, plagioclase feldspar and pyroxene. The plagioclase feldspar is 50 to 85% anorthite ($CaAl_2Si_2O_8$) component and 15 to 50% albite ($NaAlSi_3O_8$) component. The principal pyroxene is rich in the diopside ($CaMgSi_2O_6$) component. Layer 2 of the oceanic crust is composed of extrusive volcanic flows that have interacted with the seawater to form pillow lavas and intrusive flows primarily in the form of sheeted dikes. A typical thickness for layer 2 is 1.5 km. Layer 3 is made up of gabbros and related cumulate rocks that crystallized directly from the magma chamber. Gabbros are coarse-grained basalts; the larger grain size is due to slower cooling rates at greater depths. The thickness of layer 3 is typically 4.5 km.

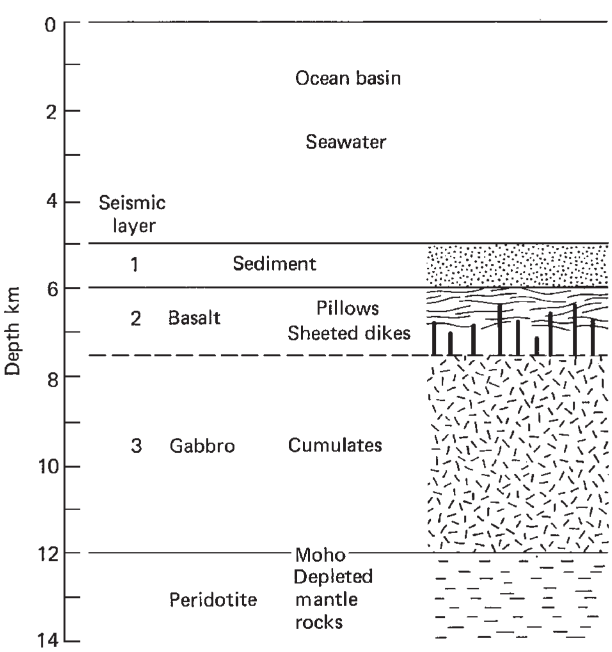

Figure 1.6 Typical structure of the oceanic crust, overlying ocean basin, and underlying depleted mantle rock.

Studies of ophiolites show that oceanic crust is underlain primarily by a peridotite called harzburgite. A typical composition of a harzburgite is given in Table 1.1. This peridotite is primarily composed of olivine and orthopyroxene. The olivine consists of about 90% forsterite component (Mg_2SiO_4) and about 10% fayalite component (Fe_2SiO_4). The orthopyroxene is less abundant and consists primarily of the enstatite component ($MgSiO_3$). Relative to basalt, harzburgite contains lower concentrations of calcium and aluminum and much higher concentrations of magnesium. The basalt of the oceanic crust with a density of 2900 kg m^{-3} is gravitationally stable with respect to the underlying peridotite with a density of 3300 kg m^{-3}. The harzburgite has a greater melting temperature ($\simeq$500 K higher) than basalt and is therefore more refractory.

Field studies of ophiolites indicate that the harzburgite did not crystallize from a melt. Instead, it is the crystalline residue left after partial melting produced the basalt. The process by which partial melting

Table 1.1 Typical Compositions (wt%) of Important Rock Types

	Granite	Diorite	Clastic Sediments	Continental Crust	Basalt	Harzburgite	"Pyrolite"	Chondrite
SiO_2	70.8	57.6	70.4	61.7	50.3	45.3	46.1	33.3
Al_2O_3	14.6	16.9	14.3	15.8	16.5	1.8	4.3	2.4
Fe_2O_3	1.6	3.2	——	——	——	——	——	——
FeO	1.8	4.5	5.3	6.4	8.5	8.1	8.2	35.5
MgO	0.9	4.2	2.3	3.6	8.3	43.6	37.6	23.5
CaO	2.0	6.8	2.0	5.4	12.3	1.2	3.1	2.3
Na_2O	3.5	3.4	1.8	3.3	2.6	——	0.4	1.1
K_2O	4.2	3.4	3.0	2.5	0.2	——	0.03	——
TiO_2	0.4	0.9	0.7	0.8	1.2	——	0.2	——

produces the basaltic oceanic crust, leaving a refractory residuum of peridotite, is an example of igneous *fractionation*.

Molten basalts are less dense than the solid, refractory harzburgite and ascend to the base of the oceanic crust because of their buoyancy. At the base of the crust they form a magma chamber. Since the forces driving plate tectonics act on the oceanic lithosphere, they produce a fluid-driven fracture at the ridge crest. The molten basalt flows through this fracture, draining the magma chamber and resulting in surface flows. These surface flows interact with the seawater to generate pillow basalts. When the magma chamber is drained, the residual molten basalt in the fracture solidifies to form a dike. The solidified rock in the dike prevents further migration of molten basalt, the magma chamber refills, and the process repeats. A typical thickness of a dike in the vertical sheeted dike complex is 1 m.

Other direct evidence for the composition of the mantle comes from *xenoliths* that are carried to the surface in various volcanic flows. Xenoliths are solid rocks that are entrained in erupting magmas. Xenoliths of mantle peridotites are found in some basaltic flows in Hawaii and elsewhere. Mantle xenoliths are also carried to the Earth's surface in kimberlitic eruptions. These are violent eruptions that form the kimberlite pipes where diamonds are found.

It is concluded that the composition of the upper mantle is such that basalts can be fractionated leaving harzburgite as a residuum. One model composition for the parent undepleted mantle rock is called *pyrolite* and its chemical composition is given in Table 1.1. In order to produce the basaltic oceanic crust, about 20% partial melting of pyrolite must occur. Incompatible elements such as the heat-producing elements uranium, thorium, and potassium do not fit into the crystal structures of the principal minerals of the residual harzburgite; they are therefore partitioned into the basaltic magma during partial melting.

Support for a pyrolite composition of the mantle also comes from studies of meteorites. A pyrolite composition of the mantle follows if it is hypothesized that the Earth was formed by the accretion of parental material similar to Type 1 carbonaceous chondritic meteorites. An average composition for a Type 1 carbonaceous chondrite is given in Table 1.1. In order to generate a pyrolite composition for the mantle, it is necessary to remove an appropriate amount of iron to form the core as well as some volatile elements such as potassium.

A 20% fractionation of pyrolite to form the basaltic ocean crust and a residual harzburgite mantle explains the major element chemistry of these components. The basalts generated over a large fraction of the ocean ridge system have near-uniform compositions in both major and trace elements. This is evidence that the parental mantle rock from which the basalt is fractionated also has a near-uniform composition. However, both the basalts of normal ocean crust and their parental mantle rock are systematically depleted in incompatible elements compared with the