GEODYNAMICS

Second Edition

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Plate Tectonics

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.

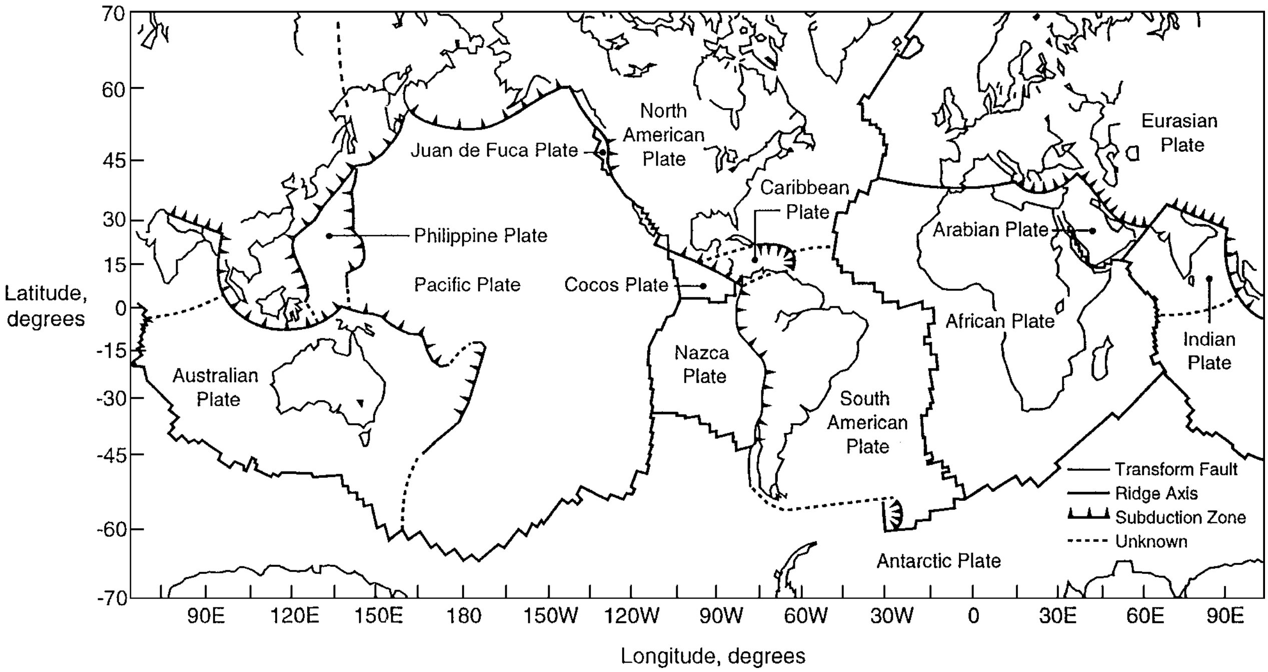
The plates are made up of relatively cool rocks and have an average thickness of about 100 km. The plates are being continually created and consumed. At ocean ridges adjacent plates diverge from each other in a process known as *seafloor spreading*. As the adjacent plates diverge, hot mantle rock ascends 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 and the Alaskan earthquake in 1964. 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

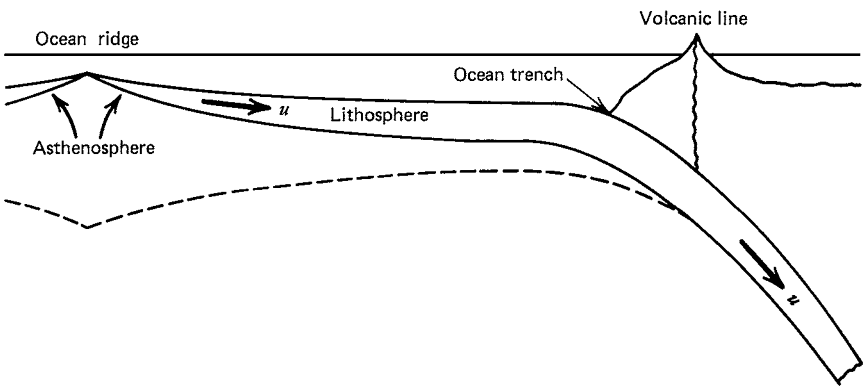


1–1 Distribution of the major plates. The ocean ridge axis (accretional plate margins), subduction zones (convergent plate margins), and transform faults that make up the plate boundaries are shown.

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 Mt. Baker in the north to Mt. Shasta in the south. Mt. 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 Mt. Pinatubo in the Philippines in 1991, the most violent eruption of the 20th 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

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.



1–3 Izalco volcano in El Salvador, an example of a subduction zone volcano (NOAA–NGDC Howell Williams).

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 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 10^9 years (1 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, more dense 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 lower viscosity. But the mantle viscosity is a strong function of mantle temperature; a lower 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 fluidlike 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. The rigidity of the lithosphere ensures, however, that the interiors of the plates do not deform significantly.

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

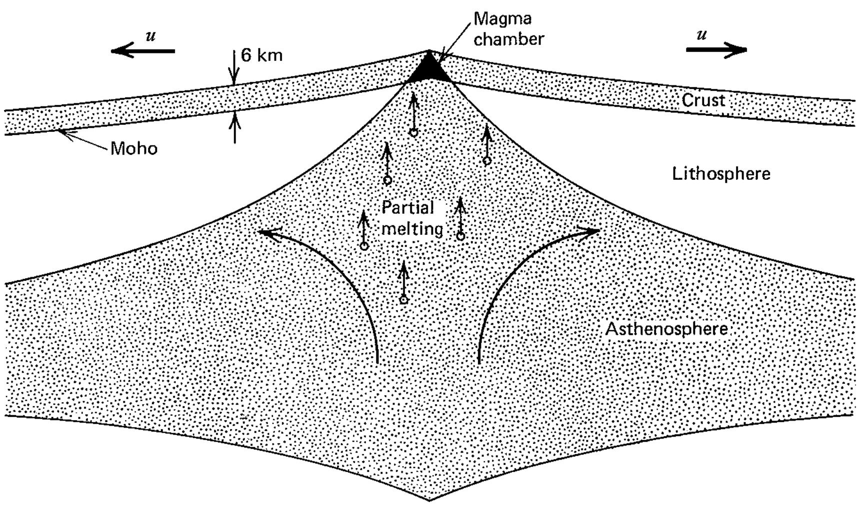
Lithospheric plates are created at ocean ridges. The two plates on either side of an ocean ridge move away from each other with near constant velocities of a few tens of millimeters per year. As the two plates diverge, hot mantle rock flows upward to fill the gap. The upwelling mantle rock cools by conductive 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.

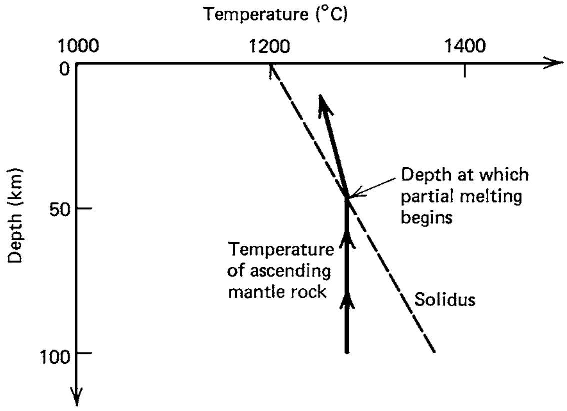
As the plates move away from the ocean ridge, they continue to cool and the lithosphere thickens. The elevation of the ocean ridge as a function of distance from the ridge crest can be explained in terms of the temperature distribution in the lithosphere. As the lithosphere cools, it becomes more dense; 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. 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 ($\approx$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

1–4 An accreting plate margin at an ocean ridge.





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.

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

$$p = \rho g y, \tag{1–1}$$

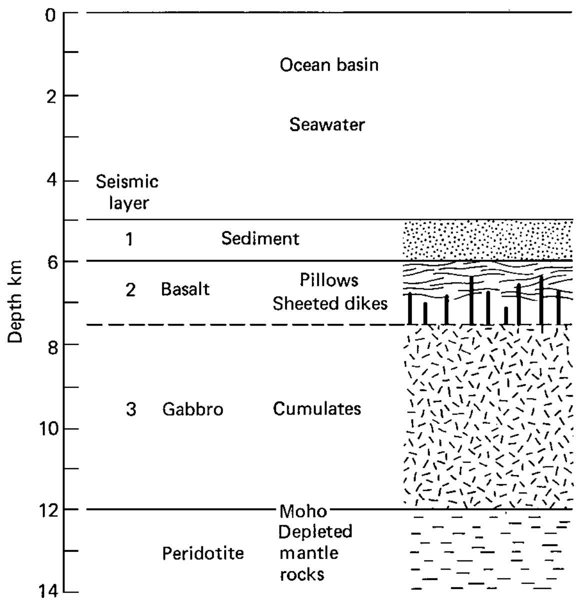
where ρ is the density of the mantle rock, g is the acceleration of gravity, and y is the depth. The solidus temperature (the temperature at which the rock first melts) decreases with decreasing pressure. When the temperature of the ascending mantle rock equals the solidus temperature, melting occurs, 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

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

Assume $\rho = 3300\ \text{kg m}^{-3}$, $g = 10\ \text{m s}^{-2}$, 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



1–6 Typical structure of the oceanic crust, overlying ocean basin, and underlying depleted mantle rock.

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

TABLE 1–1 **Typical Compositions 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	—

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.

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 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 model chondritic abundances. The missing incompatible elements are found to reside in the continental crust.

Seismic studies have been used to determine the thickness of the oceanic crust on a worldwide basis. The thickness of the basaltic oceanic crust has a nearly constant value of about 6 km throughout much of the area of the oceans. Exceptions are regions of abnormally shallow bathymetry such as the North Atlantic near Iceland, where the oceanic crust may be as thick as 25 km. The near-constant thickness of the basaltic oceanic crust places an important constraint on mechanisms of partial melting beneath the ridge crest. If the basalt of the oceanic crust represents a 20% partial melt, the thickness of depleted mantle beneath the oceanic crust is about 24 km. However, this depletion is gradational so the degree of depletion decreases with depth.

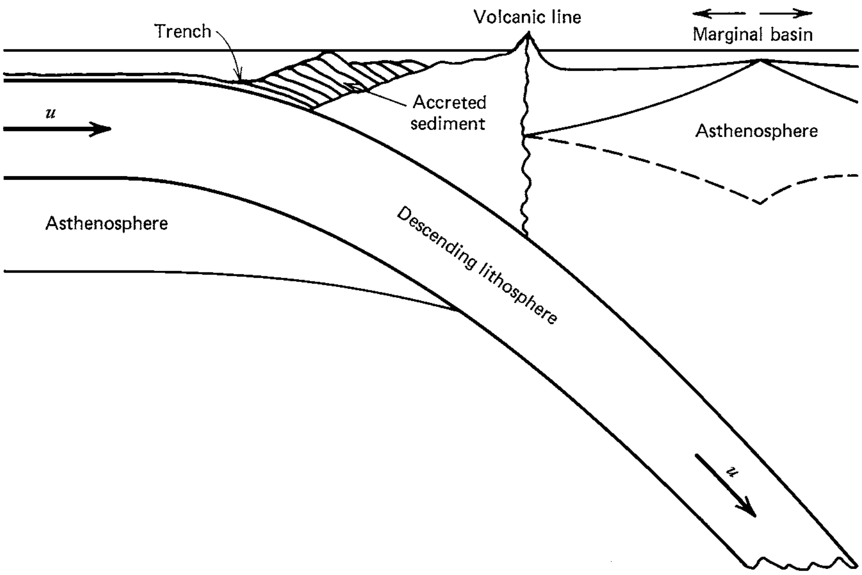
1–4 Subduction

As the oceanic lithosphere moves away from an ocean ridge, it cools, thickens, and becomes more dense because of thermal contraction. Even though the basaltic rocks of the oceanic crust are lighter than the underlying mantle rocks, the colder subcrustal rocks in the lithosphere become sufficiently dense to make old oceanic lithosphere heavy enough to be gravitationally unstable with respect to the hot mantle rocks immediately underlying the lithosphere. As a result of this gravitational instability the oceanic lithosphere founders and begins to sink into the interior of the Earth at ocean trenches. As the lithosphere descends into the mantle, it encounters increasingly dense rocks. However, the rocks of the lithosphere also become increasingly dense as a result of the increase of pressure with depth (mantle rocks are compressible), and they continue to be heavier than the adjacent mantle rocks as they descend into the mantle so long as they remain colder than the surrounding mantle rocks at any depth. Phase changes in the descending lithosphere and adjacent mantle and compositional variations with depth in the ambient mantle may complicate this simple picture of thermally induced gravitational instability. Generally speaking, however, the descending lithosphere continues to subduct as long as it remains denser than the immediately adjacent mantle rocks at any depth. The subduction of the oceanic lithosphere at an ocean trench is illustrated schematically in Figure 1–7.

The negative buoyancy of the dense rocks of the descending lithosphere results in a downward body force. Because the lithosphere behaves elastically, it can transmit stresses and acts as a stress guide. The body force acting on the descending plate is transmitted to the surface plate, which is pulled toward the ocean trench. This is one of the important forces driving plate tectonics and continental drift. It is known as *slab pull*.

Prior to subduction the lithosphere begins to bend downward. The convex curvature of the seafloor defines the seaward side of the ocean trench. The oceanic lithosphere bends continuously and maintains its structural integrity as it passes through the subduction zone. Studies of elastic bending at subduction zones are in good agreement with the morphology of some subduction zones seaward of the trench axis (see Section 3–17). However, there are clearly significant deviations from a simple elastic rheology. Some trenches exhibit a sharp "hinge" near the trench axis and this has been attributed to an elastic–perfectly plastic rheology (see Section 7–11).

As a result of the bending of the lithosphere, the near-surface rocks are placed in tension, and block faulting often results. This block faulting allows some of the overlying sediments to be entrained in the upper part of the basaltic crust. Some of these sediments

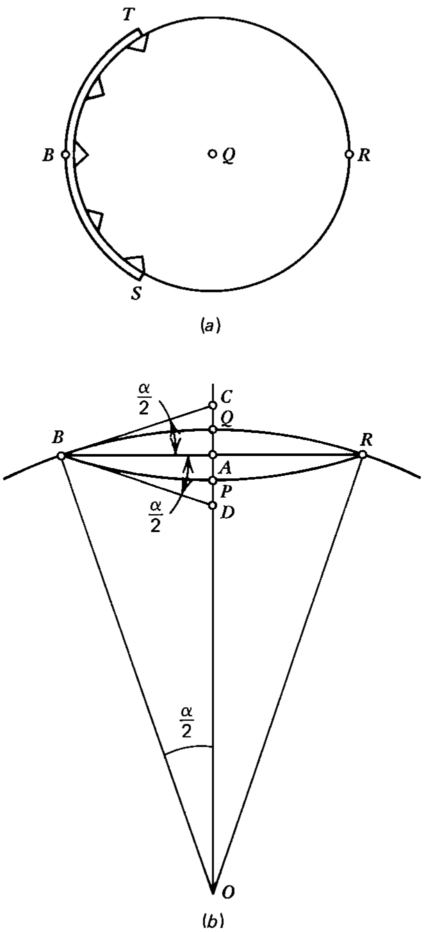


1–7 Subduction of oceanic lithosphere at an ocean trench. Sediments forming layer 1 of the oceanic crust are scraped off at the ocean trench to form the accretionary prism of sediments. The volcanic line associated with subduction and the marginal basin sometimes associated with subduction are also illustrated.

are then subducted along with the basaltic rocks of the oceanic crust, but the remainder of the sediments are scraped off at the base of the trench. These sediments form an *accretionary prism* (Figure 1–7) that defines the landward side of many ocean trenches. Mass balances show that only a fraction of the sediments that make up layer 1 of the oceanic crust are incorporated into accretionary prisms. Since these sediments are derived by the erosion of the continents, the subduction of sediments is a mechanism for subducting continental crust and returning it to the mantle.

The arclike structure of many ocean trenches (see Figure 1–1) can be qualitatively understood by the *ping-pong ball analogy*. If a ping-pong ball is indented, the indented portion will have the same curvature as the original ball, that is, it will lie on the surface of an imaginary sphere with the same radius as the ball, as illustrated in Figure 1–8. The lithosphere as it bends downward might also be expected to behave as a flexible but inextensible thin spherical shell. In this case the angle of dip α of the lithosphere at the trench can be related to the radius of curvature of the island arc. A cross section of the subduction zone is shown in Figure 1–8*b*. The triangles OAB, BAC, and BAD are similar right triangles so that the angle subtended by the indented section of the sphere at the center of the Earth is equal to the angle of dip. The radius of curvature of the indented section, defined as the great circle distance BQ, is thus $a\alpha/2$, where a is the radius of the Earth. The radius of curvature of the arc of the Aleutian trench is about 2200 km. Taking $a = 6371$ km, we find that $\alpha = 39.6°$. The angle of dip of the descending lithosphere along much of the Aleutian trench is near 45°. Although the ping-pong ball analogy provides a framework for understanding the arclike structure of some trenches, it should be emphasized that other trenches do not have an arclike form and have radii of curvature that are in poor agreement with this relationship. Interactions of the descending lithosphere with an adjacent continent may cause the descending lithosphere to deform so that the ping-pong ball analogy would not be valid.

Ocean trenches are the sites of many of the largest earthquakes. These earthquakes occur on the fault zone separating the descending lithosphere from the overlying lithosphere. Great earthquakes, such as the 1960 Chilean earthquake and the 1964 Alaskan earthquake, accommodate about 20 m of downdip motion of the oceanic lithosphere and have lengths of about 350 km along the trench. A large fraction of the relative displacement between the descending lithosphere and the overlying mantle wedge appears to be accommodated by great earthquakes of this type. A typical velocity of



1–8 The ping-pong ball analogy for the arc structure of an ocean trench. (*a*) Top view showing subduction along a trench extending from *S* to *T*. The trench is part of a small circle centered at *Q*. (*b*) Cross section of indented section. *BQR* is the original sphere, that is, the surface of the Earth. *BPR* is the indented sphere, that is, the subducted lithosphere. The angle of subduction α is *CBD*. *O* is the center of the Earth.

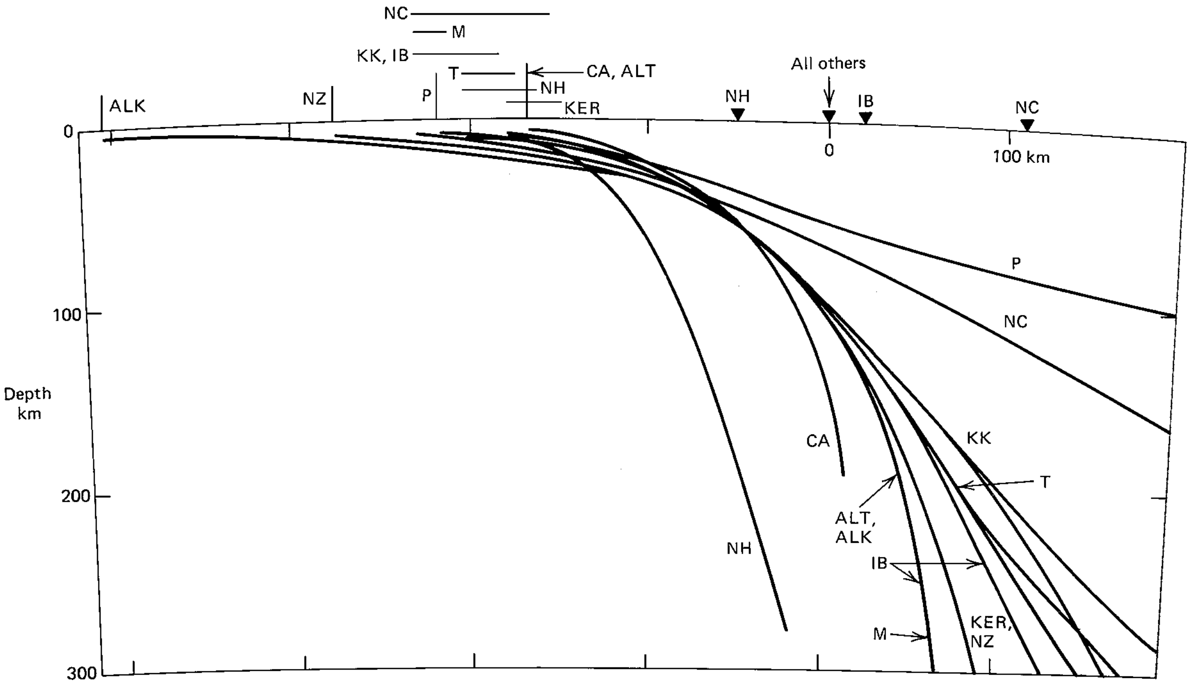
subduction is 0.1 m yr^{-1} so that a great earthquake with a displacement of 20 m would be expected to occur at intervals of about 200 years.

Earthquakes within the cold subducted lithosphere extend to depths of about 660 km. The locations of these earthquakes delineate the structure of the descending plate and are known as the *Wadati-Benioff zone*. The shapes of the upper boundaries of several descending lithospheres are given in Figure 1–9. The positions of the trenches and the volcanic lines are also shown. Many subducted lithospheres have an angle of dip near 45°. In the New Hebrides the dip is significantly larger, and in Peru and North Chile the angle of dip is small.

The lithosphere appears to bend continuously as it enters an ocean trench and then appears to straighten out and descend at a near-constant dip angle. A feature of some subduction zones is paired belts of deep seismicity. The earthquakes in the upper seismic zone, near the upper boundary of the descending lithosphere, are associated with compression. The earthquakes within the descending lithosphere are associated with tension. These double seismic zones are attributed to the "unbending," i.e., straightening out, of the descending lithosphere. The double seismic zones are further evidence of the rigidity of the subducted lithosphere. They are also indicative of the forces on the subducted lithosphere that are straightening it out so that it descends at a typical angle of 45°.

Since the gravitational body force on the subducted lithosphere is downward, it would be expected that the subduction dip angle would be 90°. In fact, as shown in Figure 1–9, the typical dip angle for a subduction zone is near 45°. One explanation is that the oceanic lithosphere is "foundering" and the trench is migrating oceanward. In this case the dip angle is determined by the flow kinematics. While this explanation is satisfactory in some cases, it has not been established that all slab dips can be explained by the kinematics of mantle flows. An alternative explanation is that the subducted slab is supported by the induced flow above the slab. The descending lithosphere induces a corner flow in the mantle wedge above it, and the pressure forces associated with this corner flow result in a dip angle near 45° (see Section 6–11).

One of the key questions in plate tectonics is the fate of the descending plates. Earthquakes terminate at a depth of about 660 km, but termination of seismicity does not imply cessation of subduction. This is the depth of a major seismic discontinuity associated with the solid–solid phase change from spinel to perovskite and magnesiowüstite; this phase change could act to deter penetration of the descending lithosphere. In some cases seismic activity spreads out at this depth, and in some cases it does not. Shallow subduction earthquakes generally indicate extensional stresses whereas the deeper earthquakes indicate compressional stresses. This is also an indication of a resistance to subduction. Seismic velocities in the cold descending



1–9 The shapes of the upper boundaries of descending lithospheres at several oceanic trenches based on the distributions of earthquakes. The names of the trenches are abbreviated for clarity (NH = New Hebrides, CA = Central America, ALT = Aleutian, ALK = Alaska, M = Mariana, IB = Izu–Bonin, KER = Kermadec, NZ = New Zealand, T = Tonga, KK = Kurile–Kamchatka, NC = North Chile, P = Peru). The locations of the volcanic lines are shown by the solid triangles. The locations of the trenches are shown either as a vertical line or as a horizontal line if the trench–volcanic line separation is variable (Isacks and Barazangi, 1977).

lithosphere are significantly higher than in the surrounding hot mantle. Systematic studies of the distribution of seismic velocities in the mantle are known as *mantle tomography*. These studies have provided examples of the descending plate penetrating the 660-km depth.

The fate of the descending plate has important implications regarding mantle convection. Since plates descend into the *lower mantle*, beneath a depth of 660 km, some form of *whole mantle* convection is required. The entire upper and at least a significant fraction of the lower mantle must take part in the plate tectonic cycle. Although there may be a resistance to convection at a depth of 660 km, it is clear that the plate tectonic cycle is not restricted to the *upper mantle* above 660 km.

Volcanism is also associated with subduction. A line of regularly spaced volcanoes closely parallels the trend of the ocean trench in almost all cases. These volcanics may result in an island arc or they may occur on the continental crust (Figure 1–10). The volcanoes lie 125 to 175 km above the descending plate, as illustrated in Figure 1–9.

It is far from obvious why volcanism is associated with subduction. The descending lithosphere is cold compared with the surrounding mantle, and thus it

1–10 Eruption of ash and steam from Mount St. Helens, Washington, on April 3, 1980. Mount St. Helens is part of a volcanic chain, the Cascades, produced by subduction of the Juan de Fuca plate beneath the western margin of the North American plate (Washington Department of Natural Resources).

should act as a heat sink rather than as a heat source. Because the flow is downward, magma cannot be produced by pressure-release melting. One source of heat is frictional dissipation on the fault zone between the descending lithosphere and the overlying mantle. However, there are several problems with generating island-arc magmas by frictional heating. When rocks are cold, frictional stresses can be high, and significant heating can occur. However, when the rocks become hot, the stresses are small, and it appears to be impossible to produce significant melting simply by frictional heating.

It has been suggested that interactions between the descending slab and the induced flow in the overlying mantle wedge can result in sufficient heating of the descending oceanic crust to produce melting. However, thermal models of the subduction zone show that there is great difficulty in producing enough heat to generate the observed volcanism. The subducted cold lithospheric slab is a very large heat sink and strongly depresses the isotherms above the slab. It has also been argued that water released from the heating of hydrated minerals in the subducted oceanic crust can contribute to melting by depressing the solidus of the crustal rocks and adjacent mantle wedge rocks. However, the bulk of the volcanic rocks at island arcs have near-basaltic compositions and erupt at temperatures very similar to eruption temperatures at accretional margins. Studies of the petrology of island-arc magmas indicate that they are primarily the result of the partial melting of rocks in the mantle wedge above the descending lithosphere. Nevertheless, geochemical evidence indicates that partial melting of subducted sediments and oceanic crust does play an important role in island-arc volcanism. Isotopic studies have shown conclusively that subducted sediments participate in the melting process. Also, the locations of the surface volcanic lines have a direct geometrical relationship to the geometry of subduction. In some cases two adjacent slab segments subduct at different angles, and an offset occurs in the volcanic line; for the shallower dipping slab, the volcanic line is farther from the trench keeping the depth to the slab beneath the volcanic line nearly constant.

Processes associated with the subducted oceanic crust clearly trigger subduction zone volcanism. However, the bulk of the volcanism is directly associated with the melting of the mantle wedge in a way similar to the melting beneath an accretional plate margin. A possible explanation is that "fluids" from the descending oceanic crust induce melting and create sufficient buoyancy in the partially melted mantle wedge rock to generate an ascending flow and enhance melting through pressure release. This process may be three-dimensional with ascending diapirs associated with individual volcanic centers.

In some trench systems a secondary accretionary plate margin lies behind the volcanic line, as illustrated in Figure 1–7. This *back-arc spreading* is very similar to the seafloor spreading that is occurring at ocean ridges. The composition and structure of the ocean crust that is being created are nearly identical. Back-arc spreading creates *marginal basins* such as the Sea of Japan. A number of explanations have been given for back-arc spreading. One hypothesis is that the descending lithosphere induces a secondary convection cell, as illustrated in Figure 1–11*a*. An alternative hypothesis is that the ocean trench migrates away from an adjacent continent because of the "foundering" of the descending lithosphere. Back-arc spreading is required to fill the gap, as illustrated in Figure 1–11*b*. If the adjacent continent is being driven up against the trench, as in South America, marginal basins do not develop. If the adjacent continent is stationary, as in the western Pacific, the foundering of the lithosphere leads to a series of marginal basins as the trench migrates seaward. There is observational evidence that back-arc spreading centers are initiated at volcanic lines. Heating of the lithosphere at the volcanic line apparently weakens it sufficiently so that it fails under tensional stresses.

PROBLEM 1–3 If we assume that the current rate of subduction, 0.09 $m^2\ s^{-1}$, has been applicable in the past, what thickness of sediments would have to have been subducted in the last 3 Gyr if the mass of subducted sediments is equal to one-half the present mass of the continents? Assume the density of the continents ρ_c is 2700 kg m^{-3}, the density of the sediments ρ_s is 2400 kg m^{-3}, the continental area A_c is 1.9×10^8 km^2, and the mean continental thickness h_c is 35 km.

1–5 Transform Faults

In some cases the rigid plates slide past each other along *transform faults*. The ocean ridge system is not