Chapter 1

Earth materials and the geosphere

1.1 EARTH MATERIALS

This book concerns the nature, origin, evolution and significance of Earth materials. Earth is composed of a variety of naturally occurring and synthetic materials whose composition can be expressed in many ways. Solid Earth materials are described by their chemical, mineral and rock composition. Atoms combine to form minerals and minerals combine to form rocks. Discussion of the relationships between atoms, minerals and rocks is fundamental to an understanding of Earth materials and their behavior.

The term mineral is used in a number of ways. For example, elements on your typical breakfast cereal box are listed as minerals. Oil and gas are considered mineral resources. All these are loose interpretations of the term mineral. In the narrowest sense, minerals are defined by the following five properties:

1 Minerals are **solid**, so they do not include liquids and gases. Minerals are solid

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because all the atoms in them are held together in fixed positions by forces called chemical bonds (Chapter 2).

2 Minerals are **naturally occurring**. This definition excludes synthetic solids produced through technology. Many solid Earth materials are produced by both natural and synthetic processes. Natural and synthetic diamonds are a good example. Another example is the solid materials synthesized in high temperature and high pressure laboratory experiments that are thought to be analogous to real minerals that occur only in the deep interior of Earth.

3 Minerals usually form by **inorganic** processes. Some solid Earth materials form by both inorganic and organic processes. For example, the mineral calcite ($CaCO_3$) forms by inorganic processes (stalactites and other cavestones) and is also precipitated as shell material by organisms such as clams, snails and corals.

4 Each mineral species has a **specific chemical composition** which can be expressed by a chemical formula. An example is common table salt or halite which is

composed of sodium and chlorine atoms in a 1:1 ratio (NaCl). Chemical compositions may vary within well-defined limits because minerals incorporate impurities, have atoms missing, or otherwise vary from their ideal compositions. In addition some types of atoms may substitute freely for one another when a mineral forms, generating a well-defined range of chemical compositions. For example, magnesium (Mg) and iron (Fe) may substitute freely for one another in the mineral olivine whose composition is expressed as $(Mg,Fe)_2SiO_4$. The parentheses are used to indicate the variable amounts of Mg and Fe that may substitute for each other in olivine group minerals (Chapter 3).

5   Every mineral species possesses a **long-range, geometric arrangement of constituent atoms or ions**. This implies that the atoms in minerals are not randomly arranged. Instead minerals crystallize in geometric patterns so that the same pattern is repeated throughout the mineral. In this sense, minerals are like three-dimensional wall paper. A basic pattern of atoms, a motif, is repeated systematically to produce the entire geometric design. This long-range pattern of atoms characteristic of each mineral species is called its **crystal structure**. All materials that possess geometric crystal structures are **crystalline** materials. Solid materials that lack a long-range crystal structure are **amorphous** materials, where amorphous means without form; without a long-range geometric order.

Over 3500 minerals have been discovered to date (Wenk and Bulakh, 2004) and each is distinguished by a unique combination of crystal structure and chemical composition. Strictly speaking, naturally-occurring, solid materials that lack one of the properties described above are commonly referred to as **mineraloids**. Common examples include amorphous materials such as volcanic glass and organic crystalline materials such as those in organic sedimentary rocks such as coal.

Most of the solid Earth is composed of various types of rock. A rock is an **aggregate of mineral crystals and/or mineraloids**. A **monominerallic** rock consists of multiple crystals of a single mineral. Examples include the sedimentary rock quartz sandstone, which may consist of nothing but grains of quartz held together by quartz cement, and the igneous rock dunite, which can consist entirely of olivine crystals. Most rocks are **polyminerallic**; they are composed of many types of mineral crystals. For example, granite commonly contains quartz, potassium feldspar, plagioclase, hornblende and biotite and may include other mineral species.

Mineral composition is one of the major defining characteristics of rocks. Rock textures and structures are also important defining characteristics. It is not surprising that the number of rock types is very large indeed, given the large number of different minerals that occur in nature, the different conditions under which they form, and the different proportions in which they can combine to form aggregates with various textures and structures. Helping students to understand the properties, classification, origin and significance of rocks is the major emphasis of this text.

1.2   THE GEOSPHERE

Earth materials can occur anywhere within the geosphere, whose radius is approximately 6380 km (Figure 1.1). In static standard models of the geosphere, Earth is depicted with a number of roughly concentric layers. Some of these layers are distinguished primarily on the basis of differences in composition and others by differences in their state or mechanical properties. These two characteristics by which the internal layers of Earth are distinguished are not totally independent, because differences in chemical, mineralogical and/or rock composition influence mechanical properties and state.

1.2.1   Compositional layers

The layers within Earth that are defined largely on the basis of chemical composition (Figure 1.1; left side) include: (1) the **crust**, which is subdivided into **continental** and **oceanic** crust, (2) the **mantle**, and (3) the **core**. Each of these layers has a distinct combination of chemical, mineral and rock compositions that distinguishes it from the others as

Compositional layers   **Mechanical layers**
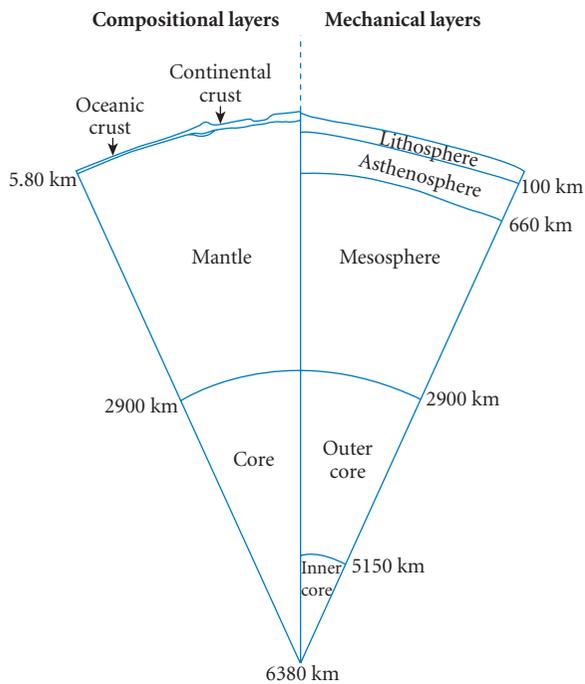
Figure 1.1   Standard cross-section model of the geosphere showing the major compositional layers on the left and the major mechanical layers on the right.

described in the next section. The thin crust ranges from 5 to 80 km thick and occupies <1% of Earth's volume. The much thicker mantle has an average radius of ~2885 km and occupies ~83% of Earth's volume. The core has a radius of ~3480 km and comprises ~16% of Earth's volume.

1.2.2   Mechanical layers

The layers within Earth defined principally on the basis of mechanical properties (Figure 1.1; right side) include: (1) a strong lithosphere to an average depth of ~100 km that includes all of the crust and the upper part of the mantle; (2) a weaker asthenosphere extending to depths ranging from 100 to 660 km and including a transition zone from ~400 to 660 km; and (3) a mesosphere or lower mantle from ~660 to 2900 km. The core is divided into a liquid outer core (~2900–5150 km) and a solid inner core, below ~5150 km to the center of Earth. Each of these layers is distinguished from the layers above and below by its unique mechanical properties. The major

features of each of these layers are summarized in the next section.

1.3.1   Earth's crust

The outermost layer of the geosphere, Earth's crust, is extremely thin; in some ways it is analogous to the very thin skin on an apple. The crust is separated from the underlying mantle by the **Mohorovičić (Moho) discontinuity**. Two major types of crust occur.

Oceanic crust

Oceanic crust is composed largely of dark-colored, mafic rocks enriched in oxides of magnesium, iron and calcium (MgO, FeO and CaO) relative to average crust. The elevated iron (Fe) content is responsible for both the dark color and the elevated density of oceanic crust. Oceanic crust is thin; the depth to the Moho averages 5–7 km. Under some oceanic islands, its thickness reaches 18 km. The elevated density and small thickness of oceanic crust cause it to be less buoyant than continental crust, so that it occupies areas of lower elevation on Earth's surface. As a result, most oceanic crust of normal thickness is located several thousand meters below sea level and is covered by oceans. Oceanic crust consists principally of rocks such as basalt and gabbro, composed largely of the minerals pyroxene and calcic plagioclase. These mafic rocks comprise layers 2 and 3 of oceanic crust and are generally topped with sediments that comprise layer 1 (Table 1.1). An idealized stratigraphic column (see Figure 1.8) of ocean crust consists of three main layers, each of which can be subdivided into sublayers.

Oceanic crust is young relative to the age of the Earth (~4.55 Ga = 4550 Ma). The oldest ocean crust, less than 180 million years old (180 Ma), occurs along the western and eastern borders of the north Atlantic Ocean and in the western Pacific Ocean. Older oceanic crust has largely been destroyed by subduction, but fragments of oceanic crust, perhaps as old as 2.5 Ga, may be preserved on land in the form of ophiolites. **Ophiolites** may be slices of ocean crust thrust onto continental margins and, if so, provide evidence for the existence of Precambrian oceanic crust.

Table 1.1   A comparison of oceanic and continental crust characteristics.

Properties	Oceanic crust	Continental crust
Composition	Dark-colored, mafic rocks enriched in MgO, FeO and CaO Averages ~50% SiO_2	Complex; many lighter colored felsic rocks Enriched in K_2O, Na_2O and SiO_2 Averages ~60% SiO_2
Density	Higher; less buoyant Average 2.9–3.1 g/cm^3	Lower; more buoyant Average 2.6–2.9 g/cm^3
Thickness	Thinner; average 5–7 km thickness Up to 15 km under islands	Thicker; average 30 km thickness Up to 80 km under mountains
Elevation	Low surface elevation; mostly submerged below sea level	Higher surface elevations; mostly emergent above sea level
Age	Up to 180 Ma for in-place crust ~3.5% of Earth history	Up to 4000 Ma 85–90% of Earth history

Continental crust

Continental crust has a much more variable composition than oceanic crust. Continental crust can be generalized as "granitic" in composition, enriched in K_2O, Na_2O and SiO_2 relative to average crust. Although igneous and metamorphic rocks of granitic composition are common in the upper portion of continental crust, lower portions contain more rocks of dioritic and/or gabbroic composition. Granites and related rocks tend to be light-colored, lower density felsic rocks rich in quartz and potassium and sodium feldspars. Continental crust is generally much thicker than oceanic crust; the depth to the Moho averages 30 km. Under areas of very high elevation, such as the Himalayas, its thickness approaches 80 km. The greater thickness and lower density of continental crust make it more buoyant than oceanic crust. As a result, the top of continental crust is generally located at higher elevations and the surfaces of the continents tend to be above sea level. The distribution of land and sea on Earth is largely dictated by the distribution of continental and oceanic crust. Only the thinnest portions of continental crust, most frequently along thinned continental margins and rifts, occur below sea level.

Whereas modern oceans are underlain by oceanic crust younger than 180 Ma, the oldest well-documented continental crust includes 4.03 Ga rocks from the Northwest Territories of Canada (Stern & Bleeker, 1998). Approximately 4 Ga rocks also occur in Greenland and Australia. Greenstone belts (Chapter 18) may date back as far as 4.28 Ga (O'Neill et al., 2008) suggesting that crust began forming within 300 million years of Earth's birth. Individual zircon grains from metamorphosed sedimentary rocks in Australia have been dated at 4.4 Ga (Wilde et al., 2001). The great age of some continental crust results from its relative buoyancy. In contrast to ocean crust, continental crust is largely preserved as its density is too low for it to be readily subducted. Table 1.1 summarizes the major differences between oceanic and continental crust.

1.3.2   Earth's mantle

The mantle is thick (~2900 km) relative to the radius of Earth (~6370 km) and constitutes ~83% of Earth's total volume. The mantle is distinguished from the crust by being very rich in MgO (30–40%) and, to a lesser extent, in FeO. It contains an average of approximately 40–45% SiO_2 which means it has an **ultrabasic composition** (Chapter 7). Some basic rocks such as eclogite occur in smaller proportions. In the upper mantle (depths to 400 km), the silicate minerals olivine and pyroxene are dominant; spinel, plagioclase and garnet are locally common. These minerals combine to produce dark-colored **ultramafic rocks** (Chapter 7) such as peridotite, the dominant group of rocks in the upper mantle. Under the higher pressure conditions deeper in the mantle similar chemical components combine to produce dense minerals with tightly packed structures. These high pressure mineral transformations are largely indicated
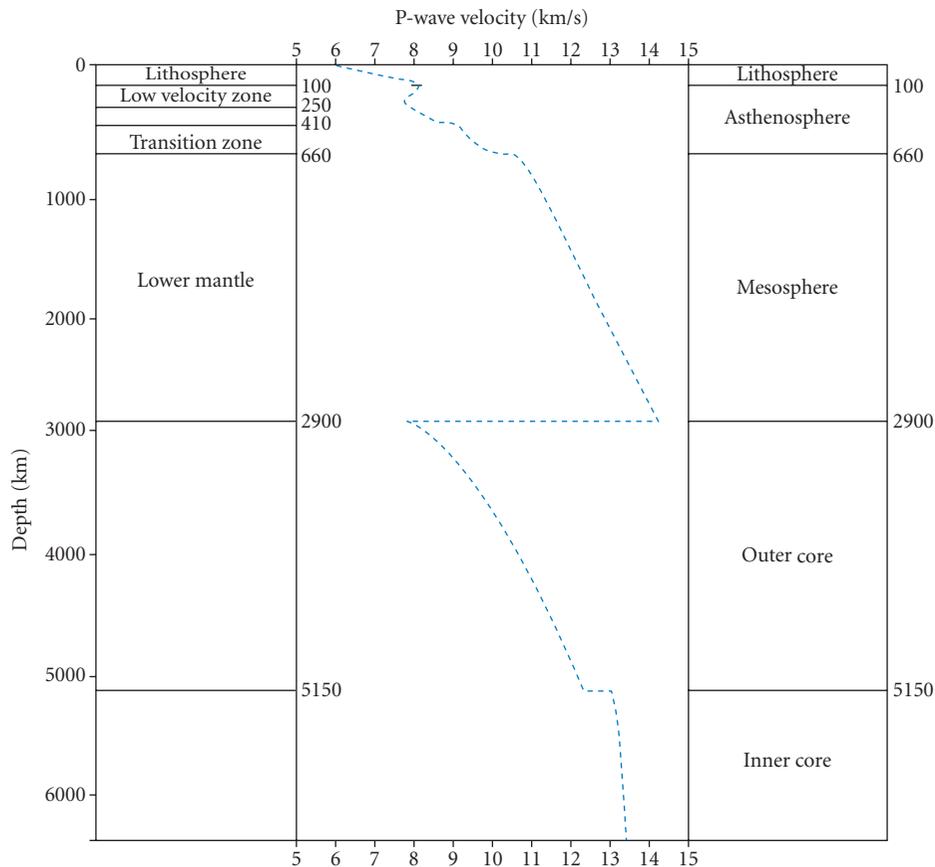
Figure 1.2 Major layers and seismic (P-wave) velocity changes within Earth, with details of the upper mantle layers.

by changes in seismic wave velocity, which reveal that the mantle contains a number of sublayers (Figure 1.2) as discussed below.

Upper mantle and transition zone

The uppermost part of the mantle and the crust together constitute the relatively rigid **lithosphere,** which is strong enough to rupture in response to stress. Because the lithosphere can rupture in response to stress, it is the site of most earthquakes and is broken into large fragments called plates, as discussed later in this chapter.

A discrete **low velocity zone** (LVZ) occurs within the upper mantle at depths of ~100–250 km below the surface. The top of LVZ marks the contact between the strong lithosphere and the weak asthenosphere (Figure 1.2). The **asthenosphere** is more plastic and

flows slowly, rather than rupturing, when subjected to stress. The anomalously low P-wave velocity of the LVZ has been explained by small amounts of partial melting (Anderson et al., 1971). This is supported by laboratory studies suggesting that peridotite should be very near its melting temperature at these depths due to high temperature and small amounts of water or water-bearing minerals. Below the base of the LVZ (250–410 km), seismic wave velocities increase (Figure 1.2) indicating that the materials are more rigid solids. These materials are still part of the relatively weak asthenosphere which extends to the base of the transition zone at 660 km.

Seismic discontinuities marked by increases in seismic velocity occur within the upper mantle at depths of ~410 and ~660 km (Figure 1.2). The interval between the depths

of 410 and 660 km is called the **transition zone** between the upper and lower mantle. The sudden jumps in seismic velocity record sudden increases in rigidity and incompressibility. Laboratory studies suggest that the minerals in peridotite undergo transformations into new minerals at these depths. At approximately 410 km depth (~14 GPa), olivine (Mg_2SiO_4) is transformed to more rigid, incompressible beta spinel (β-spinel), also known as wadleysite (Mg_2SiO_4). Within the transition zone, wadleysite is transformed into the higher pressure mineral ringwoodite (Mg_2SiO_4). At ~660 km depth (~24 GPa), ringwoodite and garnet are converted to very rigid, incompressible perovskite [$(Mg,Fe,Al)SiO_3$] and oxide phases such as periclase (MgO). The mineral phase changes from olivine to wadleysite and from ringwoodite to perovskite are inferred to be largely responsible for the seismic wave velocity changes that occur at 410 and 660 km (Ringwood, 1975; Condie, 1982; Anderson, 1989). Inversions of pyroxene to garnet and garnet to minerals with ilmenite and perovskite structures may also be involved. The base of the transition zone at 660 km marks the base of the asthenosphere in contact with the underlying mesosphere or lower mantle (see Figure 1.2).

Lower mantle (mesosphere)

Perovskite, periclase [$(Mg,Fe)O$], magnesiowustite [$(Mg,Fe)O$], stishovite (SiO_2), ilmenite [$(Fe,Mg)TiO_2$] and ferrite [$(Ca,Na,Al)Fe_2O_4$] are thought to be the major minerals in the lower mantle or **mesosphere**, which extends from depths of 660 km to the mantle–core boundary at ~2900 km depth. Our knowledge of the deep mantle continues to expand, largely based on anomalous seismic signals deep within Earth. These are particularly common in a complex zone near the core–mantle boundary called the **D″ layer**. The D″ discontinuity ranges from ~130 to 340 km above the core–mantle boundary. Williams and Garnero (1996) proposed an **ultra low velocity zone** (ULVZ) in the lowermost mantle on seismic evidence. These sporadic ULVZs may be related to the formation of deep mantle plumes within the lower mantle. Other areas near the core–mantle boundary are characterized by anomalously fast velocities. Hutko et al.

(2006) detected subducted lithosphere that had sunk all the way to the D″ layer and may be responsible for the anomalously fast velocities. Deep subduction and deeply rooted mantle plumes support the concept of whole mantle convection and may play a significant role in the evolution of a highly heterogeneous mantle, but these concepts are highly controversial (Foulger et al., 2005).

1.3.3   Earth's core

Earth's core consists primarily of **iron** (~85%), with smaller, but significant amounts of nickel (~5%) and lighter elements (~8–10%) such as oxygen, sulfur and/or hydrogen. A dramatic decrease in P-wave velocity and the termination of S-wave propagation occurs at the 2900 km discontinuity (Gutenberg discontinuity or core–mantle boundary). Because S-waves are not transmitted by non-rigid substances such as fluids, the **outer core** is inferred to be a liquid. Geophysical studies suggest that Earth's outer core is a highly compressed liquid with a density of ~10–12 g/cm³. Circulating molten iron in Earth's outer core is responsible for the production of most of Earth's magnetic field.

The outer/inner core boundary, the Lehman discontinuity, at 5150 km, is marked by a rapid increase in P-wave velocity and the occurrence of low velocity S-waves. The solid inner core has a density of ~13 g/cm³. Density and magnetic studies suggest that Earth's **inner core** also consists largely of iron, with nickel and less oxygen, sulfur and/or hydrogen than in the outer core. Seismic studies have shown that the inner core is seismically anisotropic; that is, seismic velocity in the inner core is faster in one direction than in others. This has been interpreted to result from the parallel alignment of iron-rich crystals or from a core consisting of a single crystal with a fast velocity direction.

In this section, we have discussed the major layers of the geosphere, their composition and their mechanical properties. This model of a layered geosphere provides us with a spatial context in which to visualize where the processes that generate Earth materials occur. In the following sections we will examine the ways in which all parts of the geosphere interact to produce global tectonics. The ongoing story of global tectonics is one of the most
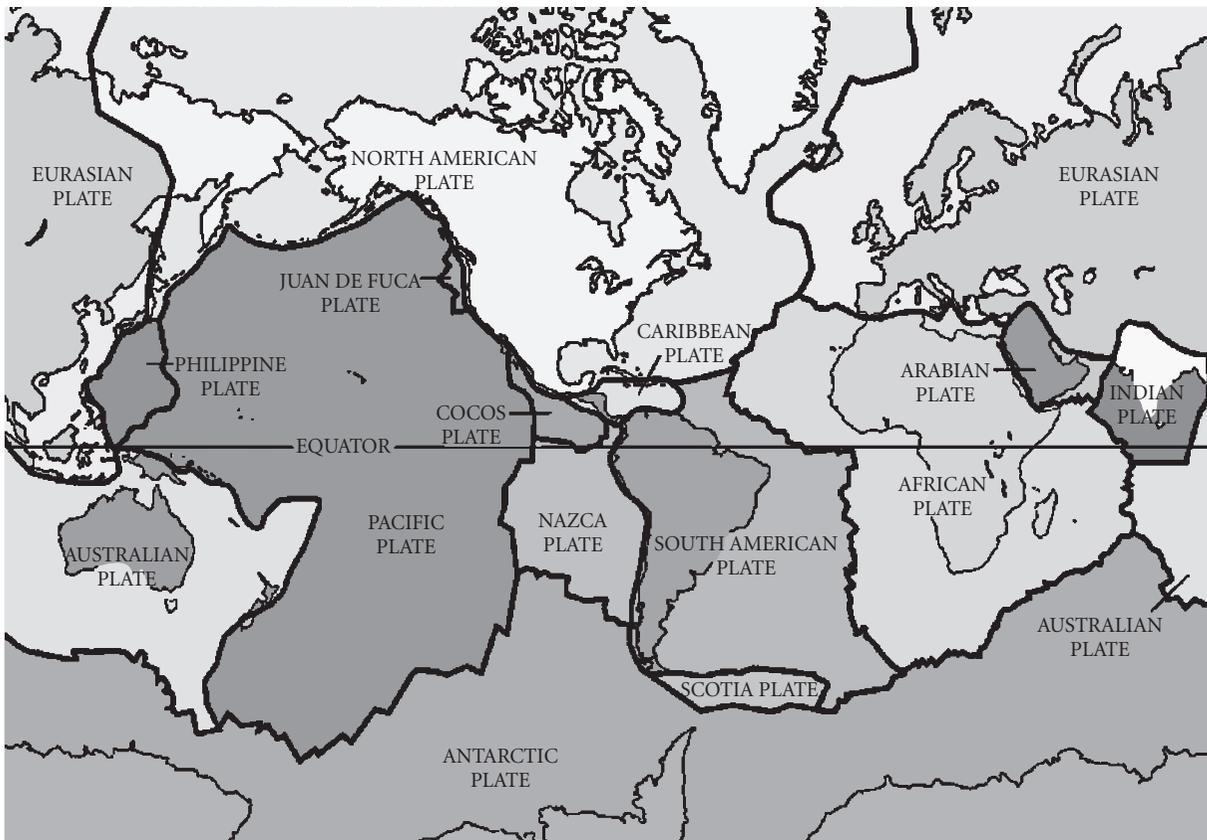
Figure 1.3   World map showing the distribution of the major plates separated by boundary segments that end in triple junctions. (Courtesy of the US Geological Survey.)

fascinating tales of scientific discovery in the last century.

1.4   GLOBAL TECTONICS

Plate tectonic theory has profoundly changed the way geoscientists view Earth and provides an important theoretical and conceptual framework for understanding the origin and global distribution of igneous, sedimentary and metamorphic rock types. It also helps to explain the distribution of diverse phenomena that include faults, earthquakes, volcanoes, mountain belts and mineral deposits.

The fundamental tenet of plate tectonics (Isacks et al., 1968; Le Pichon, 1968) is that the lithosphere is broken along major fault systems into large pieces called **plates** that move relative to one another. The existence of the strong, breakable lithosphere permits plates to form. The fact that they overlie a weak, slowly flowing asthenosphere permits them to move. Each plate is separated from adjacent plates by plate boundary segments ending in **triple junctions** (McKenzie and Morgan, 1969) where three plates are in contact (Figure 1.3).

The relative movement of plates with respect to the boundary that separates them defines three major types of plate boundary segments (Figure 1.4) and two hybrids: (1) divergent plate boundaries, (2) convergent plate boundaries, (3) transform plate boundaries, and (4) divergent–transform and convergent–transform hybrids (shown).

Each type of plate boundary produces a characteristic suite of features composed of a characteristic suite of Earth materials. This relationship between the kinds of Earth materials formed and the plate tectonic settings in
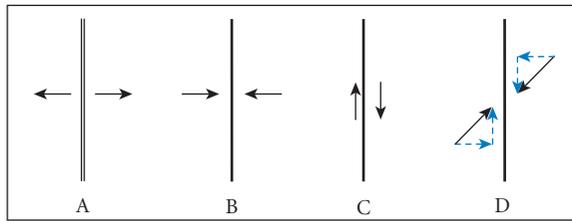
Figure 1.4   Principal types of plate boundaries: A, divergent; B, convergent; C, transform; D, hybrid convergent–transform boundary. Thick black lines represent plate boundaries and arrows indicate relative motion between the plates; blue dashed arrows show components of convergent and transform relative motion.

^ Volcanism   ▮ Rising magma   ⌢ Normal faults

Figure 1.5   Major features of continental rifts include rift valleys, thinned continental crust and lithosphere and volcanic–magmatic activity from melts generated in the rising asthenosphere.

which they are produced provides a major theme of the chapters that follow.

1.4.1   Divergent plate boundaries

Divergent plate boundaries occur where two plates are moving apart relative to their boundary (Figure 1.4a). Such areas are characterized by horizontal extension and vertical thinning of the lithosphere. Horizontal extension in continental lithosphere is marked by continental rift systems and in oceanic lithosphere by the oceanic ridge system.

Continental rifts

Continental rift systems form where horizontal extension occurs in continental lithosphere (Figure 1.5). In such regions, the lithosphere is progressively stretched and thinned, like a candy bar being stretched in two. This stretching occurs by brittle, normal faulting near the cooler surface and by ductile flow at deeper, warmer levels. Extension is accompanied by uplift of the surface as the hot asthenosphere rises under the thinned lithosphere. Rocks near the surface of the lithosphere eventually rupture along normal faults to produce continental rift valleys. The East African Rift, the Rio Grande Rift in the United States and the Dead Sea Rift in the Middle East are modern examples of continental rift valleys.

If horizontal extension and vertical thinning occur for a sufficient period of time, the continental lithosphere may be completely rifted into two separate continents. Complete **continental rifting** is the process by which supercontinents such as Pangea and Rodinia were broken into smaller continents such as those we see on Earth's surface at present. When this happens, a new and growing ocean basin begins to form between the two continents by the process of **sea floor spreading** (Figure 1.6). The most recent example of this occurred when the Arabian Peninsula separated from the rest of Africa to produce the Red Sea basin some 5 million years ago. Older examples include the separation of India from Africa to produce the northwest Indian Ocean basin (~115 Ma) and the separation of North America from Africa to produce the north Atlantic Ocean basin (~180 Ma). Once the continental lithosphere has rifted completely, the divergent plate boundary is no longer situated within continental lithosphere. Its position is instead marked by a portion of the oceanic ridge system where oceanic crust is produced and grows by sea floor spreading (Figure 1.6).

Oceanic ridge system

The **oceanic ridge system** (ridge) is Earth's largest mountain range and covers roughly 20% of Earth's surface (Figure 1.7). The ridge is >65,000 km long, averages ~1500 km in width, and rises to a crest with an average

Continents separate, ridge forms,
initiating sea floor spreading and
ocean basin creation

Ridge

Rising
magma

Sea floor spreading widens
ocean basins as sediments
cover continental margins

Ridge

Rising
magma

Sediments   Oceanic crust   Continental   Normal
                                        crust   faults
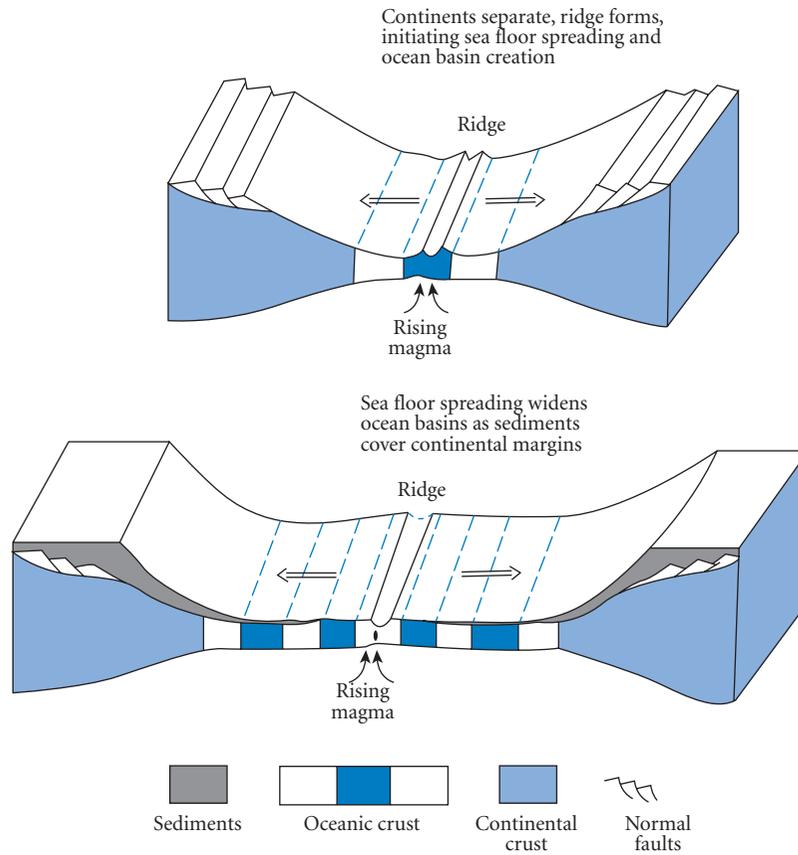
Figure 1.6   Model showing the growth of ocean basins by sea floor spreading from the ridge system following the complete rifting of continental lithosphere along a divergent plate boundary.

Figure 1.7   Map of the ocean floor showing the distribution of the oceanic ridge system. (Courtesy of Marie Tharp, with permission of Bruce C. Heezen and Marie Tharp, 1977; © Marie Tharp 1977/2003. Reproduced by permission of Marie Tharp Maps, LLC, 8 Edward Street, Sparkhill, NT 10976, USA.) (For color version, see Plate 1.7, opposite p. 248.)

elevation of ~3 km above the surrounding sea floor. A moment's thought will show that the ridge system is only a broad swell on the ocean floor, whose slopes on average are very gentle. Since it rises only 3 km over a horizontal distance of 750 km, then the average slope is 3 km/750 km which is about 0.4%; the average slope is about 0.4°. We exaggerate the vertical dimension on profiles and maps in order to make the subtle stand out. Still there are differences in relief along the ridge system. In general, warmer, faster spreading portions of the ridge such as the East Pacific Rise (~6–18 cm/yr) have gentler slopes than colder, slower spreading portions such as the Mid-Atlantic Ridge (~2–4 cm/yr). The central or axial portion of the ridge system is marked by a rift valley, especially along slower spreading segments, or other rift features, and marks the position of the divergent plate boundary in oceanic lithosphere.

One of the most significant discoveries of the 20th century (Dietz, 1961; Hess, 1962) was that oceanic crust and lithosphere form along the axis of the ridge system, then spreads away from it in both directions, causing ocean basins to grow through time. The details of this process are illustrated by Figure 1.8. As the lithosphere is thinned, the asthenosphere rises toward the surface generating basaltic–gabbroic melts. Melts that crystallize in magma bodies well below the surface form plutonic rocks such as gabbros that become layer 3 in oceanic crust. Melts intruded into near-vertical fractures above the chamber form the basaltic–gabbroic sheeted dikes that become layer 2b. Lavas that flow onto the ocean floor commonly form basaltic pillow lavas that become layer 2a. The marine sediments of layer 1 are deposited atop the basalts. In this way layers 1, 2 and 3 of the oceanic crust are formed. The underlying mantle consists of ultramafic rocks (layer 4). Layered ultramafic rocks form by differentiation near the base of the basaltic–gabbroic magma bodies, whereas the remainder of layer 4 represents the unmelted, refractory residue that accumulates below the magma body.

Because the ridge axis marks a divergent plate boundary, the new sea floor on one side moves away from the ridge axis in one direction and the new sea floor on the other side moves in the opposite direction relative to the ridge axis. More melts rise from the astheno-
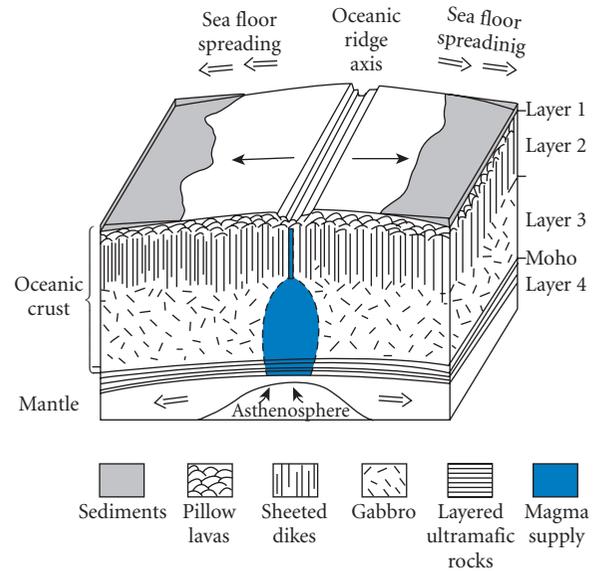
Figure 1.8   The formation of oceanic crust along the ridge axis generates layer 2 pillow basalts and dikes, layer 3 gabbros of the oceanic crust and layer 4 mantle peridotites. Sediment deposition on top of these rocks produces layer 1 of the crust. Sea floor spreading carries these laterally away from the ridge axis in both directions.

sphere and the process is repeated, sometimes over >100 Ma. In this way ocean basins grow by sea floor spreading as though new sea floor is being added to two conveyor belts that carry older sea floor in opposite directions away from the ridge where it forms (Figure 1.8). Because most oceanic lithosphere is produced along divergent plate boundaries marked by the ridge system, they are also called **constructive** plate boundaries.

As the sea floor spreads away from the ridge axis, the crust thickens from above by the accumulation of additional marine sediments and the lithosphere thickens from below by a process called **underplating**, which occurs as the solid, unmelted portion of the asthenosphere spreads laterally and cools through a critical temperature below which it becomes strong enough to fracture. As the entire lithosphere cools, it contracts, becomes denser and sinks so that the floors of the ocean gradually deepen away from the thermally elevated ridge axis. As explained in the next section, if the density of oceanic litho-
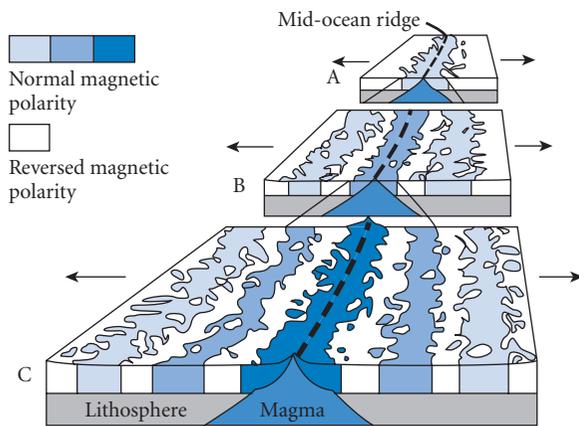
Figure 1.9 Model depicting the production of alternating normal (colored) and reversed (white) magnetic bands in oceanic crust by progressive sea floor spreading and alternating normal and reversed periods of geomagnetic polarity (A through C). The age of such bands should increase away from the ridge axis. (Courtesy of the US Geological Survey.)

sphere exceeds that of the underlying asthenosphere, subduction occurs.

The formation of oceanic lithosphere by sea floor spreading implies that the age of oceanic crust should increase systematically away from the ridge in opposite directions. Crust produced during a period of time characterized by normal magnetic polarity should split in two and spread away from the ridge axis as new crust formed during the subsequent period of reversed magnetic polarity forms between it. As indicated by Figure 1.9, repetition of this splitting process produces oceanic crust with bands (linear magnetic anomalies) of alternating normal and reversed magnetism whose age increases systematically away from the ridge (Vine and Matthews, 1963).

Sea floor spreading was convincingly demonstrated in the middle to late 1960s by paleomagnetic studies and radiometric dating that showed that the age of ocean floors systematically increases in both directions away from the ridge axis, as predicted by sea floor spreading (Figure 1.10).

Hess (1962), and those who followed, realized that sea floor spreading causes the outer layer of Earth to grow substantially over time.

If Earth's circumference is relatively constant and Earth's lithosphere is growing horizontally at divergent plate boundaries over a long period of time, then there must be places where it is undergoing long-term horizontal shortening of similar magnitude. As ocean lithosphere ages and continues to move away from ocean spreading centers, it cools, subsides and becomes more dense over time. The increased density causes the ocean lithosphere to become denser than the underlying asthenosphere. As a result, a plate carrying old, cold, dense oceanic lithosphere begins to sink downward into the asthenosphere, creating a convergent plate boundary.

1.4.2 Convergent plate boundaries

Convergent plate boundaries occur where two plates are moving toward one another relative to their mutual boundary (Figure 1.11). The scale of such processes and the features they produce are truly awe inspiring.

Subduction zones

The process by which the leading edge of a denser lithospheric plate is forced downward into the underlying asthenosphere is called **subduction**. The downgoing plate is called the subducted plate or downgoing slab; the less dense plate is called the overriding plate. The area where this process occurs is a subduction zone. The subducted plate, whose thickness averages 100 km, is always composed of oceanic lithosphere. Subduction is the major process by which oceanic lithosphere is destroyed and recycled into the asthenosphere at rates similar to oceanic lithosphere production along the oceanic ridge system. For this reason, subduction zone plate boundaries are also called **destructive** plate boundaries.

The surface expressions of subduction zones are **trench–arc systems** of the kind that encircle most of the shrinking Pacific Ocean. Trenches are deep, elongate troughs in the ocean floors marked by water depths that can approach 11 km. They are formed as the downgoing slab forces the overriding slab to bend downward forming a long trough along the boundary between them.

Because the asthenosphere is mostly solid, it resists the downward movement of the

| Chron | 5 | 6 | 13 | 18 | 21 | 25 | 31 | 34 | M0 | M4 | M10 | M16 | M21 | M25 |

Age 0   9.7   20.1   33.1 40.1 47.9 55.9   67.7   83.5   120.4 126.7   139.6 147.7   180.0 Ma
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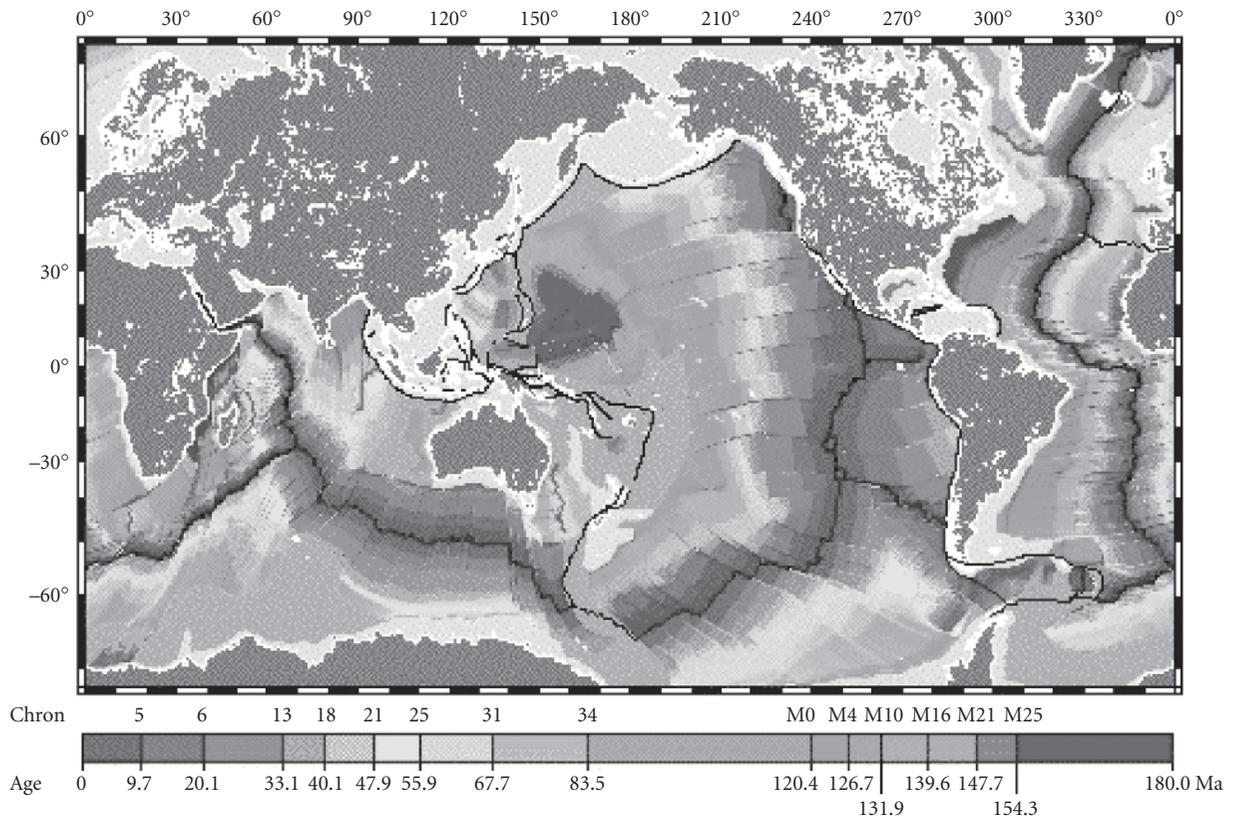
Figure 1.10   World map showing the age of oceanic crust; such maps confirmed the origin of oceanic crust by sea floor spreading. (From Muller et al., 1997; with permission of the American Geophysical Union.) (For color version, see Plate 1.10, opposite p. 248.)
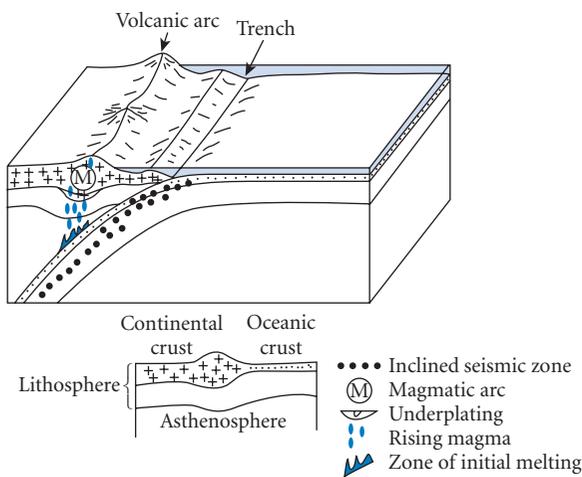
Figure 1.11   Convergent plate boundary, showing a trench–arc system, inclined seismic zone and subduction of oceanic lithosphere.

subducted plate. This produces stresses in the cool interior of the subducted lithosphere that generate earthquakes (Figure 1.11) along an **inclined seismic (Wadati–Benioff) zone** that marks the path of the subducted plate as it descends into the asthenosphere. The three largest magnitude earthquakes in the past century occurred along inclined seismic zones beneath Chile (1909), Alaska (1964) and Sumatra (2004). The latter event produced the devastating Banda Aceh tsunami which killed some 300,000 people in the Indian Ocean region.

What is the ultimate fate of subducted slabs? Earthquakes occur in subducted slabs to a depth of 660 km, so we know slabs reach the base of the asthenosphere transition zone. Earthquake records suggest that some slabs flatten out as they reach this boundary,

indicating that they may not penetrate below this. Seismic tomography, which images three-dimensional variations in seismic wave velocity within the mantle, has shed some light on this question, while raising many questions. A consensus is emerging (Hutko et al., 2006) that some subducted slabs become dense enough to sink all the way to the core–mantle boundary where they contribute material to the D″ layer. These recycled slabs may ultimately be involved in the formation of mantle plumes, as suggested by Jeanloz (1993).

Subduction zones produce a wide range of distinctive Earth materials. The increase in temperature and pressure within the subducted plate causes it to undergo significant metamorphism. The upper part of the subducted slab, in contact with the hot asthenosphere, releases fluids as it undergoes metamorphism which triggers partial melting. A complex set of melts rise from this region to produce **volcanic–magmatic arcs**. These melts range in composition from basaltic–gabbroic through dioritic–andesitic and may differentiate or be contaminated to produce melts of granitic–rhyolitic composition. Melts that reach the surface produce volcanic arcs such as those that characterize the "ring of fire" of the Pacific Ocean basin. Mt St. Helens in Washington, Mt Pinatubo in the Philippines, Mt Fuji in Japan and Krakatau in Indonesia are all examples of composite volcanoes that mark the volcanic arcs that form over Pacific Ocean subduction zones.

When magmas intrude the crust they also produce plutonic igneous rocks that add new continental crust to the Earth. Most of the world's major **batholith belts** represent plutonic magmatic arcs, subsequently exposed by erosion of the overlying volcanic arc. In addition, many of Earth's most important ore deposits are produced in association with volcanic–magmatic arcs over subduction zones.

Many of the magmas generated over the subducted slab cool and crystallize at the base of the lithosphere, thickening it by underplating. Underplating and intrusion are two of the major sets of processes by which new continental crust is generated by the solidification of melts. Once produced, the density of continental crust is generally too low for it to be subducted. This helps to explain the great age that continental crust can achieve (>4.0 Ga).

Areas of significant relief, such as trench–arc systems, are ideal sites for the production and accumulation of detrital (epiclastic) sedimentary rocks. Huge volumes of detrital sedimentary rocks produced by the erosion of volcanic and magmatic arcs are deposited in forearc and backarc basins (Figure 1.12). They also occur with deformed abyssal sediments in the forearc subduction complex. As these sedimentary rocks are buried and deformed, they are metamorphosed.

Continental collisions

As ocean basins shrink by subduction, portions of the ridge system may be subducted. Once the ridge is subducted, growth of the ocean basin by sea floor spreading ceases, the ocean basin continues to shrink by subduction, and the continents on either side are brought closer together as subduction proceeds. Eventually they converge to produce a continental collision.

When a **continental collision** occurs (Dewy and Bird, 1970), subduction ceases, because continental lithosphere is too buoyant to be subducted to great depths. The continental lithosphere involved in the collision may be part of a continent, a microcontinent or a volcanic–magmatic arc. As convergence continues, the margins of both continental plates are compressed and shortened horizontally and thickened vertically in a manner analogous to what happens to two vehicles in a head-on collision. In the case of continents colliding at a convergent plate boundary, however, the convergence continues for millions of years resulting in a severe horizontal shortening and vertical thickening which results in the progressive uplift of a mountain belt and/or extensive elevated plateau that mark the closing of an ancient ocean basin (Figure 1.13).

Long mountain belts formed along convergent plate boundaries are called **orogenic belts**. The increasing weight of the thickening orogenic belt causes the adjacent continental lithosphere to bend downward to produce **foreland basins**. Large amounts of detrital sediments derived from the erosion of the mountain belts are deposited in such basins. In addition, increasing temperatures and pressures within the thickening orogenic belt cause regional metamorphism of the
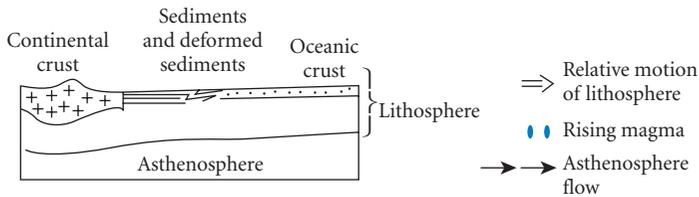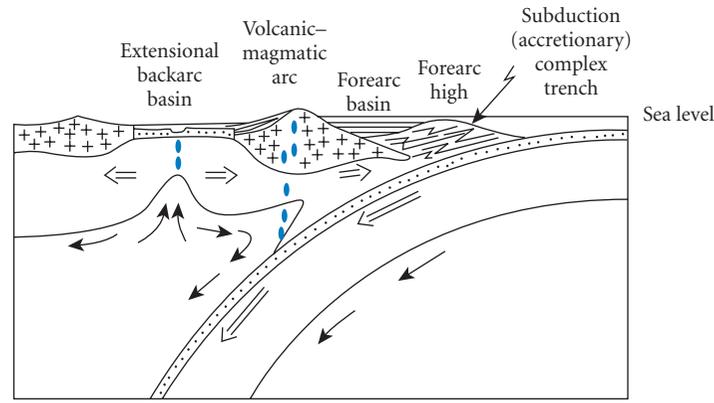
Figure 1.12 Subduction zone depicting details of sediment distribution, sedimentary basins and volcanism in trench–arc system forearc and backarc regions.
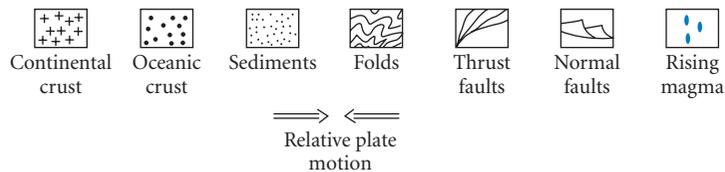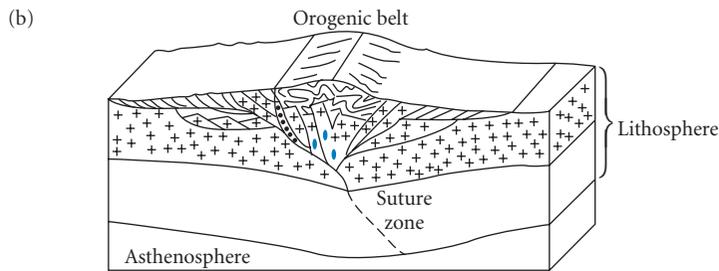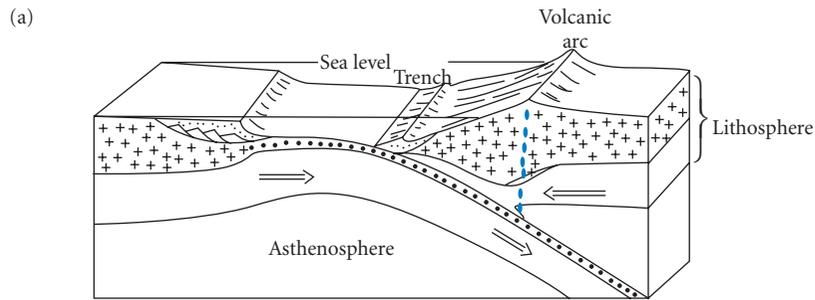
Figure 1.13 (a) Ocean basins shrink by subduction, as continents on two plates converge. (b) Continental collision produces a larger continent from two continents joined by a suture zone. Horizontal shortening and vertical thickening are accommodated by folds and thrust faults in the resulting orogenic belt.

(a)

EURASIANPLATE

INDIA
Today

10 million
years ago

SRI LANKA

38 million
years ago

Equator

55 million
years ago

INDIAN
OCEAN

71 million
years ago

"INDIA"
Land mass

SRI LANKA

(b)

Feet of the Earth™
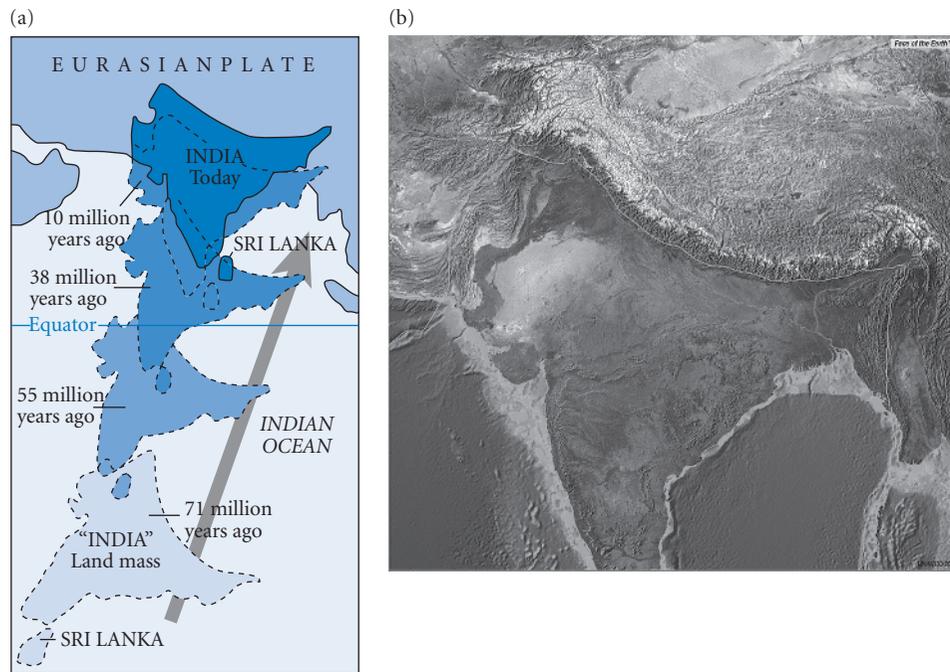
Figure 1.14   (a) Diagram depicting the convergence of India and Asia which closed the Tethys Ocean. (Courtesy of NASA.) (b) Satellite image of southern Asia showing the indentation of Eurasia by India, the uplift of Himalayas and Tibetan Plateau and the mountains that "wrap around" India. (Courtesy of UNAVCO.)

rocks within it. If the temperatures become high enough, partial melting may occur to produce melts in the deepest parts of orogenic belts that rise to produce a variety of igneous rocks.

The most striking example of a modern orogenic belt is the Himalayan Mountain range formed by the collision of India with Eurasia over the past 40 Ma. The continued convergence of the Indian microcontinent with Asia has resulted in shortening and regional uplift of the Himalayan mountain belt along a series of major thrust faults and has produced the Tibetan Plateau. Limestones near the summit of Mt Everest (Chomolungma) were formed on the floor of the Tethys Ocean that once separated India and Asia, and were then thrust to an elevation of nearly 9 km as that ocean was closed and the Himalayan Mountain Belt formed by continental collision. The collision has produced tectonic indentation of Asia, resulting in mountain ranges that wrap around India (Figure 1.14). The Ganges River in northern India flows approximately west–east in a trough that represents a modern foreland basin.

Continental collision inevitably produces a larger continent. It is now recognized that supercontinents such as Pangea and Rodinia were formed as the result of collisional tectonics. **Collisional tectonics** only requires converging plates whose leading edges are composed of lithosphere that is too buoyant to be easily subducted. In fact all the major continents display evidence of being composed of a collage of terranes that were accreted by collisional events at various times in their histories.

1.4.3   Transform plate boundaries

In order for plates to be able to move relative to one another, a third type of plate boundary is required. **Transform plate boundaries** are characterized by horizontal motion, along transform fault systems, which is parallel to the plate boundary segment that separates two plates (see Figure 1.4c). Because the rocks

Figure 1.15   Transform faults offsetting ridge segments on the eastern Pacific Ocean floor off Central America. Arrows show the directions of sea floor spreading away from the ridge. Portions of the fracture zones between the ridge segments are transform plate boundaries; portions beyond the ridge segments on both sides are intraplate transform scars. (Courtesy of William Haxby, LDEO, Columbia University.) (For color version, see Plate 1.15, between pp. 248 and 249.)

Figure 1.16   Fracture zones, transform faults and ridge segments in the eastern Pacific Ocean and western North America. The San Andreas Fault system is a continental transform fault plate boundary. (Courtesy of the US Geological Survey.)

on either side slide horizontally past each other, transform fault systems are a type of strike-slip fault system.

   Transform faults were first envisioned by J. T. Wilson (1965) to explain the seismic activity along fracture zones in the ocean floor. **Fracture zones** are curvilinear zones of intensely faulted, fractured oceanic crust that are generally oriented nearly perpendicular to the ridge axis (Figure 1.15). Despite these zones having been fractured by faulting along their entire length, earthquake activity is largely restricted to the transform portion of fracture zones that lies between offset ridge segments. Wilson (1965) reasoned that if sea

floor was spreading away from two adjacent ridge segments in opposite directions, the portion of the fracture zone between the two ridge segments would be characterized by relative motion in opposite directions. This would produce shear stresses resulting in strike-slip faulting of the lithosphere, frequent earthquakes and the development of a transform fault plate boundary. The exterior portions of fracture zones outside the ridge segments represent oceanic crust that was faulted and fractured when it was between ridge segments, then carried beyond the adjacent ridge segment by additional sea floor spreading. These portions of fracture zones are appropriately called healed transforms or **transform scars**. They are no longer plate boundaries; they are intraplate features because the sea floor on either side is spreading in the same direction (Figure 1.15).

   Transform plate boundaries also occur in continental lithosphere. The best known modern examples of continental transforms include the San Andreas Fault system in California (Figure 1.16), the Alpine Fault system

in New Zealand and the Anatolian Fault systems in Turkey and Iran. All these are characterized by active strike-slip fault systems of the type that characterize transform plate boundaries. In places where such faults bend or where their tips overlap, deep **pull-apart basins** may develop in which thick accumulations of sedimentary rocks accumulate rapidly.

Plates cannot simply diverge and converge; they must be able to slide past each other in opposite directions in order to move at all. Transform plate boundaries serve to accommodate this required sense of motion. Small amounts of igneous rocks form along transform plate boundaries, especially those hybrids that have a component of divergence or convergence as well. They produce much smaller volumes of igneous and metamorphic rocks than are formed along divergent and convergent plate boundaries.

1.5   HOTSPOTS AND MANTLE CONVECTION

Hotspots (Wilson, 1963) are long-lived areas in the mantle where anomalously large volumes of magma are generated. They occur beneath both oceanic lithosphere (e.g., Hawaii) and continental lithosphere (e.g., Yellowstone National Park, Wyoming) as well as along divergent plate boundaries (e.g., Iceland). Wilson pointed to linear seamount chains, such as the Hawaiian Islands (Figure 1.17), as surface expressions of hotspots. At any one time, volcanism is restricted to that portion of the plate that lies above the hotspot. As the plate continues to move, older volcanoes are carried away from the fixed hotspot and new volcanoes are formed above it. The age of these seamount chains increases systematically away from the hotspot in the direction of plate motion. For the Hawaiian chain, the data suggest a west–northwest direction of plate motion for the last 45 Ma. However, a change in orientation of the seamount chain to just west of north for older volcanoes suggests that the seafloor may have spread over the hotspot in a more northerly direction prior to 45 Ma. A similar trend of volcanism of increasing age extends southwestward from the Yellowstone Caldera.

In the early 1970s, Morgan (1971) and others suggested that hotspots were the surface expression of fixed, long-lived mantle
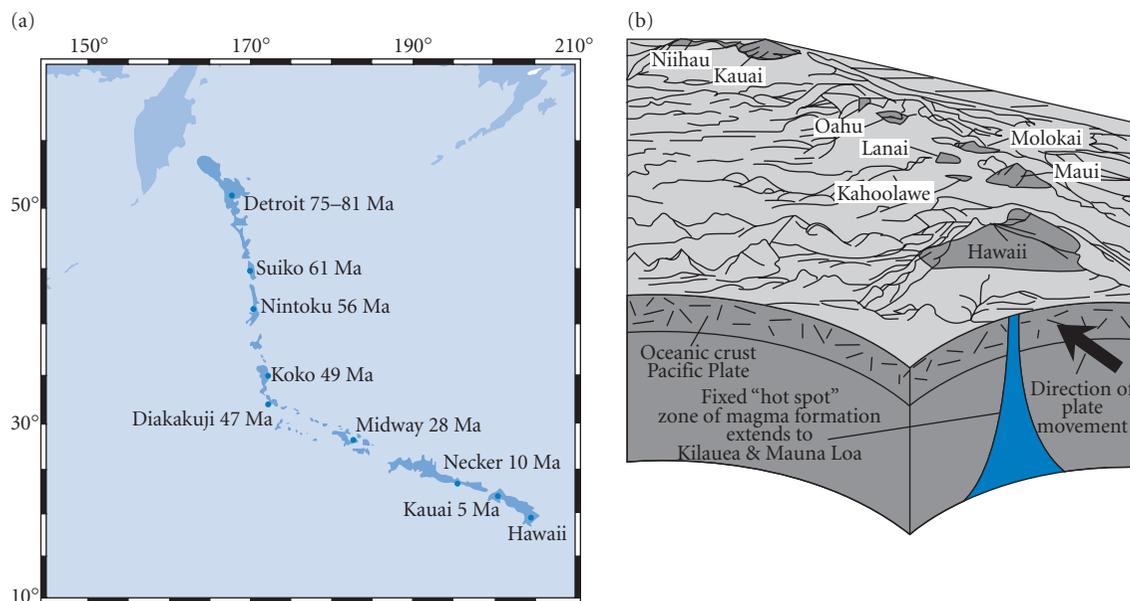
Figure 1.17   (A) Linear seamount chain formed by plate movement over the Hawaiian hotspot and/or hotspot motion. (After Tarduno et al., 2009; with permission of *Science Magazine*.) (B) "Fixed" mantle plume feeding the surface volcanoes of the Hawaiian chain. (Courtesy of the US Geological Survey.)

plumes. **Mantle plumes** were hypothesized to be columns of warm material that rose from near the core–mantle boundary. Later workers hypothesized that deep mantle plumes originate in the ULVZ of the D″ layer at the base of the mantle and may represent the dregs of subducted slabs warmed sufficiently by contact with the outer core to become buoyant enough to rise. Huge **superplumes** (Larson, 1991) were hypothesized to be significant players in extinction events, the initiation of continental rifting, and in the **supercontinent cycle** (Sheridan, 1987) of rifting and collision that has caused supercontinents to form and rift apart numerous times during Earth's history. Eventually most intraplate volcanism and magmatism was linked to hotspots and mantle plumes.

The picture has become considerably muddled over the past decade. Many Earth scientists have offered significant evidence that mantle plumes do not exist (Foulger et al., 2005). For example, there is no seismic velocity evidence for a deep plume source beneath the Yellowstone hotspot. Others have suggested that mantle plumes exist, but are not fixed (Nataf, 2000; Koppers et al., 2001; Tarduno et al., 2009). Still others (Nolet et al., 2006) suggest on the basis of fine-scale thermal tomography that some of these plumes originate near the core–mantle boundary, others at the base of the transition zone (660 km) and others at around 1400 km in the mesosphere. They suggest that the rise of some plumes from the deep mantle is interrupted by the 660 km discontinuity, whereas other plumes seem to cross this discontinuity. This is reminiscent of the behavior of subducted slabs, some of which spread out above the 660 km discontinuity, whereas other penetrate it and apparently sink to the core–mantle boundary. It is likely that hotspots are generated by a variety of processes related to mantle convection patterns that are still not well understood. Stay tuned; this will be an exciting area of Earth research over the coming decade.

We have attempted to provide a spatial and tectonic context for the processes that determine which Earth materials will form where. One part of this context involves the location of compositional and mechanical layers within the geosphere where Earth materials form. Ultimately, however, the geosphere cannot be viewed as a group of static layers. Plate tectonics implies significant horizontal and vertical movement of the lithosphere with compensating motion of the underlying asthenosphere and deeper mantle. Global tectonics suggests significant lateral heterogeneity within layers and significant vertical exchange of material between layers caused by processes such as convection, subduction and mantle plumes.

Helping students to understand how variations in composition, position within the geosphere and tectonic processes interact on many scales to generate distinctive Earth materials is the fundamental task of this book. We hope you will find what follows is both exciting and meaningful.