Kevin Hefferan
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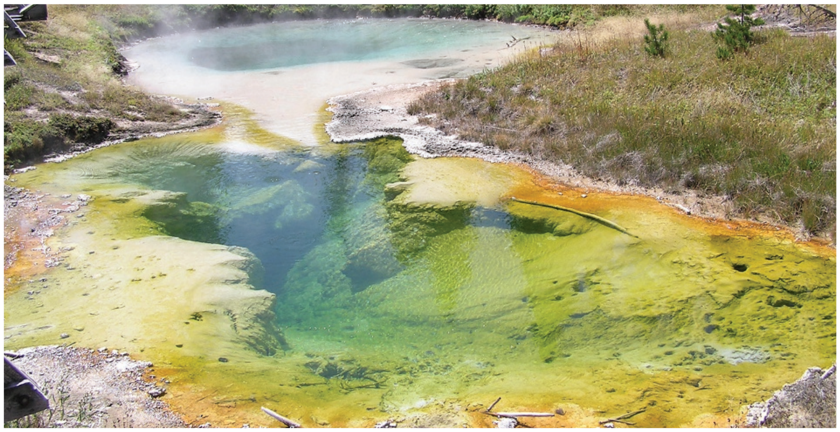
Earth Materials



Second Edition

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Contents

Preface

Earth Materials encompass the minerals, rocks, soil, and water that constitute our planet and the physical, chemical, and biological processes that produce them. Since the expansion of computer technology in the last two decades of the twentieth century, many universities have compressed or eliminated individual course offerings such as mineralogy, optical mineralogy, igneous petrology, sedimentology, and metamorphic petrology and replaced them with Earth materials courses. Earth materials courses have become an essential curricular component in the fields of geology, geoscience, Earth science, engineering geology, environmental geology, and many related areas of study. This textbook is designed to address the needs of a one or two semester Earth materials course. It can also serve as a text for those individuals who want or need an expanded background in minerals, rocks, soils, and water resources.

The chapters featured in this textbook illuminate key topics in Earth materials including their:

1. Properties, origin, and classification
2. Associations and relationships in the context of Earth's major tectonic, petrologic, hydrologic, and biogeochemical systems
3. Uses as resources and their fundamental role in our lives and the global economy
4. Relation to natural and human induced hazards
5. Impact on health and the environment.

This textbook provides:

1. Comprehensive descriptive analysis of Earth materials
2. Color graphics and insightful texthe in a logical integrated format
3. Field examples and regional relationships with graphics that illustrate concepts discussed
4. Examples of how concepts discussed can be used to address real world issues
5. Contemporary references from current scientific journals related to developments in Earth materials research
6. Summative discussions of how Earth materials are interrelated with other science and non-science fields of study.

Chapter 1 contains an introduction to Earth materials and an overview that includes a discussion of Earth's interior and plate tectonics. This introductory chapter provides a global framework for following chapters. Chapters 2–6 constitute the minerals portion of the textbook. Chapter 2 addresses mineral chemistry, bonding mechanisms, and classification. Chapter 3 explores fundamentals of crystal chemistry, including ionic substitution, phase diagrams, and isotope geochemistry. Chapter 4 highlights the principles of crystallography including symmetry operations, crystal lattices, and crystal systems. Chapter 5 examines mineral formation, macroscopic mineral properties, and the major groups of rock-forming minerals. Chapter 6 focuses on the microscopic optical properties of minerals and petrographic microscopic techniques.

Chapters 7–10 encompass the igneous portion of the textbook. Chapter 7 discusses the composition, textures, and classification of igneous rocks. Chapter 8 describes the origin and evolution of magmas and plutonic structures. Chapter 9 vividly presents volcanic processes and landforms in all their wonder. Chapter 10 explores igneous rock associations and their relations to plate tectonics.

Chapters 11–15 focus on sedimentary rocks and processes. Chapter 11 addresses weathering, sediment production, and soils. Chapter 12 discusses the sedimentary cycle, the agents, and processes by which sediments are eroded, transported, and deposited and how they imprint sediments deposited in different environments. Chapters 13 and 14 examine the composition, textures, classification, and origin of detrital sedimentary rocks and biochemical sedimentary rocks that include carbonates, evaporites, siliceous rocks, iron formations, phosphates, and carbon-rich sedimentary materials that include coal, petroleum, and natural gas.

Chapters 15–18 address metamorphic rocks and processes. Chapter 15 introduces metamorphic agents, processes, protoliths, and types of metamorphism. Chapter 16 discusses metamorphic structures, stresses, and deformation processes. Chapter 17 investigates metamorphic rock textures and classification. Chapter 18 concentrates on metamorphic zones, facies, facies series, and metamorphic trajectories in relationship to plate tectonics.

Lastly, Chapter 19 explores ores minerals, industrial minerals, and gems as well as environmental and health issues related to Earth materials.

In addition to information presented in this textbook, additional resources, including detailed descriptions of major rock-forming minerals and keys for identifying minerals using macroscopic and/or optical methods are available on the website that supports this text at:

www.wiley.com/go/hefferan/earthmaterials.

Our overall goal is to produce an innovative, visually appealing, informative, and readable textbook that addresses the full spectrum of Earth materials. We present equal treatment to minerals as well as igneous, sedimentary, and metamorphic rocks and demonstrate their impact on our personal lives and the global environment on this planet. We hope you enjoy this text and use it to further your knowledge of Earth materials.

Acknowledgments

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John O'Brien would like to thank his wife Anita, his sons Tyler and Owen and granddaughter Scarlett for their love and support. John worked for 41 years in the Department of Geoscience-Geography (now Earth and Environmental Sciences) at New Jersey City State University. He is now retired and living in the San Francisco Bay area. Sabbaticals from the University in 2005 and again in 2011, gave John the time required to complete this project. John wishes to acknowledge his indebtedness to Ansel Gooding, Charles Martin, Wayne Martin, John Pope, Robert Webb, John Crowell, Don Runnels, Joe Clark and Barry Perlmutter for encouraging, supporting and challenging him at critical times in his career.

The authors believe that an understanding of Earth materials is more important than ever in this twenty-first century in order to appropriately utilize essential resources, for mitigating geohazards and dealing with environmental change. Universities and government agencies would be wise to revitalize the Earth science and geology curricula in the United States and other countries around the world. We must all renew efforts to use our resources wisely and to engage the green economy.

About the Companion Website

This book is accompanied by a companion website.

www.wiley.com/go/hefferan/earthmaterials2

This website includes:

- Figures and Tables from the book
- Appendices and additional resources, including detailed descriptions of major rock-forming minerals and keys for identifying minerals using macroscopic and/or optical methods

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. These include their chemical, mineral, and rock composition. In simple terms, 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, their properties, and the processes that produce them.

1.2 MINERALS AND MINERALOIDS

The term **mineral** is used in a number of ways. For example, the chemical elements, such as calcium, iron, and potassium, listed on your breakfast cereal box, your bottle of vitamin supplements or your bag of fertilizer are called minerals. Coal, oil, and gas are referred to as mineral resources. All of these fall under a broad use of the term mineral. In a stricter sense used by many, but not all geologists, minerals are defined by the following properties:

1. Minerals are **solid**, so do not include liquids and gases. Minerals are solid because the atoms in them are held together in fixed positions by forces called chemical bonds (Chapter 2).
2. Minerals are **naturally occurring**, meaning that they occur naturally within the Earth. This definition excludes synthetic solids that are produced only by technologies in laboratories or factories. It does include solid Earth materials that are produced by both natural

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and synthetic processes, such as natural and synthetic diamonds and 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 Each mineral species has a **specific chemical composition** which may vary only within well-defined limits; that is to say that each mineral possesses a chemical composition that 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).

4 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**. They are minerals, in the narrow sense, if they are naturally occurring, inorganic solids with a well-defined chemical composition. Solid materials that lack a long-range crystal structure are **amorphous** materials, where amorphous means without form and without a long-range geometric order.

Many would add a fifth property that requires minerals to sometimes be **formed by inorganic processes**. It is certainly true that the vast majority of minerals conform to this property and that the vast majority of organically formed crystalline solids are not considered to be minerals. However, many solid Earth materials that form by both inorganic and organic processes are considered minerals, especially if they are important constituents of naturally formed rocks. For example, the mineral calcite is also precipitated as shell material by organisms such as clams, snails, and corals and is the major constituent of the rock limestone (Chapter 14).

Over 5500 minerals have been discovered to date (www.mindat.com) and each is distinguished by a unique combination of chemical composition and crystal structure. 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 in which the atoms lack long-range order and amber or ivory which are formed only by organic processes.

1.2.1 Rocks

Earth is largely composed of various types of **rock**. A rock is an aggregate of mineral crystals and/or mineraloids. Scarce **monominerallic** rocks consist of multiple crystals of a single mineral. Examples include the sedimentary rock quartz sandstone which may consist entirely of quartz grains held together by quartz cement and the igneous rock dunite which can consist entirely of olivine crystals. The vast majority of rocks are **polymineralic**; they are composed of many types of mineral crystals. For example, granite commonly contains quartz, potassium feldspar, plagioclase, hornblende and/or biotite, and various other minerals in small amounts.

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 minerals and rocks is the major emphasis of this text.

1.3 THE GEOSPHERE

Earth materials can occur anywhere on or within the **geosphere**, the portion the Earth from its surface to its center, whose radius is approximately 6370 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.

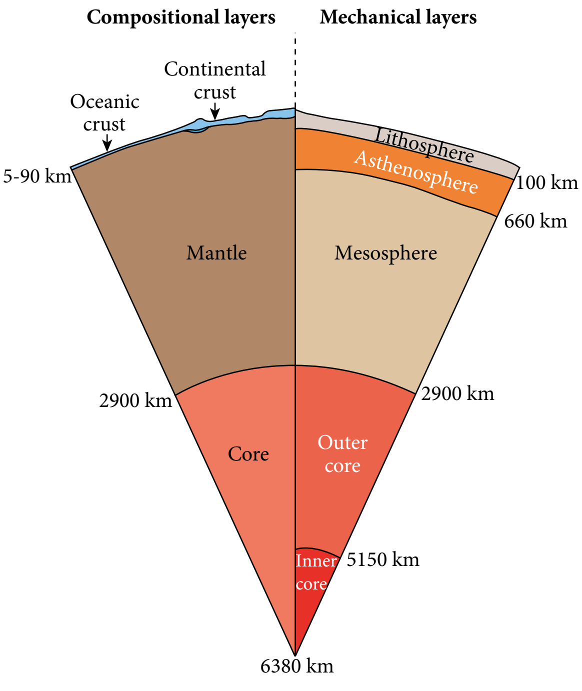

Figure 1.1 Standard cross-section model of the geosphere. Major compositional layers are shown on the left: core (red shade), mantle (brown shade), and continental and oceanic crust (blue shades). Major mechanical layers are shown on the right: inner core and outer core (red shades), lithosphere (light brown), mesosphere (dark brown), and asthenosphere (burnt orange).

1.3.1 Compositional layers

The layers within Earth that are defined largely on the basis of chemical composition (Figure 1.1; left side) include the: (1) **crust**, which is subdivided into **continental** and **oceanic** crust, (2) **mantle**, and (3) **core**. Each of these layers has a distinctive combination of chemical, mineral, and rock compositions that distinguishes it from the others, as described in the next section. The thin crust typically ranges from 5 to 85 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 ~3470 km and comprises ~16% of Earth's volume.

1.3.2 Mechanical layers

The layers within Earth defined principally on the basis of mechanical properties (Figure 1.1; right side) include: (1) a relatively strong **lithosphere** of variable thickness to an average depth of ~100 km that includes all of the crust and the upper part of the mantle, (2) a weaker **asthenosphere** that extends to depths between ~100 and 660 km and includes a **transition** *zone* from ~400 to 660 km, and (3) a **mesosphere** or **lower mantle** from ~660 to 2900 km. The underlying 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.

1.4 DETAILED MODEL OF THE GEOSPHERE

1.4.1 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 a boundary called the **Mohorovičić (Moho) discontinuity.** Two major types of crust occur.

Oceanic crust

Oceanic crust is composed largely of dark colored, basic (45–52% SiO_2) rocks

(Chapter 7) enriched in oxides of magnesium, iron, and calcium (MgO, FeO, and CaO) relative to average crust. The elevated iron (Fe) content is responsible for the both the dark color and 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 below sea level and covered by sea water to a depth of several thousand meters. Oceanic crust consists principally of basic igneous rocks such as basalt and gabbro composed largely of the minerals pyroxene and calcic plagioclase. These dark-colored, mafic igneous rocks comprise layers 2 and 3 of oceanic crust and are commonly topped with sediments that comprise layer 1 (Table 1.1). An idealized profile of typical ocean crust consists of these three main layers, each of which can be subdivided into sublayers which are briefly discussed later in this chapter.

Oceanic crust is young relative to the age of the Earth (~4.55 Ga = 4550 Ma). The oldest ocean crust in the major ocean basins, less than 190 million years old (190 Ma), occurs along the western and eastern borders of the Atlantic Ocean and in the Western Pacific Ocean. Recently, still older oceanic crust that may be 340 Ma has been discovered in the eastern Mediterranean Sea (Granot 2016). Still older oceanic crust has largely been destroyed by subduction, but fragments of such crust are preserved on land in the form of ophiolites. Ophiolites contain slices of ocean crust thrust onto continental margins and provide evidence for the existence of Precambrian oceanic crust. The age of the oldest true ophiolites of Precambrian age remains controversial (Chapter 18).

Continental crust

Continental crust has a much more variable composition than oceanic crust. Continental crust can be generalized as "granitic" in composition, and is enriched in K_2O, Na_2O, and SiO_2 relative to average crust. Although igneous and metamorphic rocks of granitic composition are fairly common in the upper portion of continental crust, lower portions contain more rocks of intermediate dioritic and even basic 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; depth to the Moho averages 30–40 km. Under areas of very high elevation, such as the Himalayas, its thickness approaches 85 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 continents with normal crustal thicknesses are above sea level. The distribution of Earth's land and sea 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 in rifts, have surfaces below sea level.

Table 1.1 Characteristics of oceanic and continental crust: a comparison.

Properties	Oceanic crust	Continental crust
Composition	Dark colored, mafic rocks enriched in MgO, FeO, and CaO	Complex; many lighter colored felsic rocks
		Enriched in K_2O, Na_2O, and SiO_2
	Averages ~50% SiO2	Averages ~60% SiO2
Density	Higher; less buoyant	Lower; more buoyant
	Average 2.9–3.1 g/cm^3	Average 2.6–2.9 g/cm^3
Thickness	Thinner; average 5–7 km thickness	Thicker; average 30 km thickness
	Up to 15 km under islands	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 190 Ma for in-place crust	Up to more than 4000 Ma
	~3.5% of Earth history	85–90% of Earth history

Whereas modern oceans, with the exception of a small area in the Mediterranean Sea, are underlain by oceanic crust younger than 190 Ma, the oldest well-documented continental crust includes 4.03 Ga rocks from the Northwest Territories of Canada (Stern and 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) which suggests that continental crust began forming within 300 million years of Earth's birth. Individual detrital zircon grains, derived from the erosion of older continental crust, occur in metamorphosed sedimentary rocks in Australia. These zircons have been dated at 4.4 Ga (Wilde et al. 2001) an age recently confirmed by Valley et al. (2014). These data suggest that continental crust may have existed no more than 150 Ma after Earth formed. 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 generally too low for it to be subducted on as large a scale. Table 1.1 summarizes the major differences between oceanic and continental crust.

1.4.2 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 Mg-rich silicate minerals olivine and pyroxene dominate; spinel, plagioclase and garnet are locally common. These minerals combine to produce generally dark colored **ultramafic rocks** such as peridotite, the dominate 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 crystal structures. These high-pressure minerals are produced by transformations that are largely indicated by changes high pressure 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 Earth stresses. 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 most areas of the upper mantle at depths of ~100–250 km below the surface. The top of low velocity zone marks the contact between the strong lithosphere and the underlying, weak asthenosphere (Figure 1.3). The **asthenosphere** is more plastic than the lithosphere and flows slowly, rather than rupturing, when subjected to stress. The anomalously low rigidity of the LVZ has been explained by small amounts of partial melting (Anderson et al. 1971). This is supported by laboratory studies that suggest peridotite should be very near its melting temperature at these depths due to the high temperature. This is especially likely if it contains small amounts of water or water-bearing minerals. Below the base of the low velocity zone (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). This interval (~410–660 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 (pressures of ~14 GPa), olivine (Mg_2SiO_4) is transformed into 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 approximately 660 km depth (~24 GPa), ringwoodite and garnet are converted to very rigid, incompressible perovskite [(Mg,Fe,Al) SiO_3], also known as bridgmanite (Tschauner

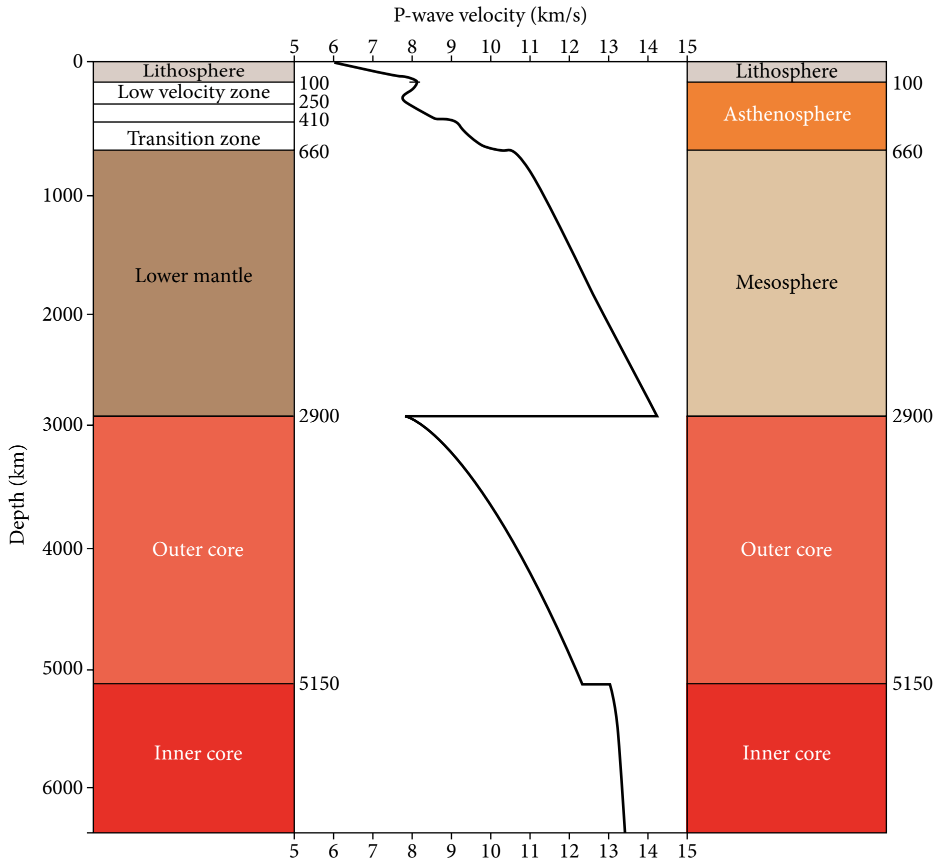

Figure 1.2 Major layers and seismic (p-wave) velocity changes within Earth; showing details of upper mantle layers. Colors are as for Figure 1.1.

Figure 1.3 World map showing the distribution of major plates separated by boundary segments that end in triple junctions. *Source*: From USGS.

et al. 2014) 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 respectively (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 (Figure 1.2).

The lower mantle (mesosphere)

The **lower mantle**, also called the **mesosphere**, extends from depths of 660 km to the core–mantle boundary at approximately 2900 km. Based on high pressure, high temperature laboratory studies, bridgmanite [(Mg,Fe,Al) SiO_3], ferropericlase [(Mg,Fe)O], magnesiowustite [(Mg,Fe)O], stishovite (SiO_2), and calcium-rich ferrite (Ca,Na,Al)Fe_2O_4 are thought to be the major minerals in the lower mantle. Our knowledge of the deep mantle continues to expand, largely based on high temperature, high pressure laboratory studies and on anomalous seismic signals deep within the Earth. A deeper layer has been proposed at about 1600 km depth where the rigidity of the mantle may increase considerably (Miyagi and Marquardt 2015). Anomalous seismic velocities are particularly common in a complex zone, of variable thickness, 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 ultra-low velocity zones 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 which 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 some type of whole mantle convection and may play a significant role in the evolution of a highly heterogeneous mantle, but these concepts are still controversial (Foulger et al. 2005).

1.4.3 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 which is **Gutenberg discontinuity** or **core–mantle boundary (CMB)**. Because S-waves are not transmitted by nonrigid substances such as fluids, the **outer core** is inferred to be a fluid. Geophysical studies suggest that the Earth's outer core is a highly compressed liquid with a density of ~10–12 g/cm^3. Slowly circulating molten, iron-rich, very viscous liquids in the outer core are believed to be 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 reemergence of low velocity S-waves. This suggests that the **inner core** is rigid. The inner core is solid and has a density of ~13 g/cm^3. Density and magnetic studies suggest that the Earth's inner core also consists of largely of iron, with nickel and less oxygen, sulfur, and/or hydrogen than 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. Recent discoveries suggest that the inner core is divided into two layers with the inner layer more rigid than the outer one and with a different orientation of its fast seismic wave direction (Ishii and Dziewonski 2002; Wang et al. 2015).

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-scale tectonics is one of the most fascinating tales of scientific discovery in the last century and new discoveries continue to be made in this one.

1.5 GLOBAL TECTONICS

1.5.1 Introduction

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 rocks (Chapters 7–18). It also helps to explain the distribution of diverse phenomena that include faults, earthquakes, volcanoes, mountain belts, mineral deposits, and even the evolution of life and the evolving composition of the atmosphere.

The fundamental tenet of **plate tectonics** (Le Pichon 1968; Isacks et al. 1968) is that the lithosphere is broken along major fault systems into large, relatively rigid pieces called plates that move relative to one another. The existence of the strong, breakable lithosphere permits plates to form. Most plates are huge, with areas of 10^5–10^8 km^2 and thicknesses that average ~10^3 km; some plates are smaller and microplates are smaller still. The fact that they overlie a weaker, slowly flowing asthenosphere permits them to move very slowly. Each plate is separated from adjacent plates by **plate boundary segments** that end 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 (2) convergent, (3) transform, (4) divergent-transform hybrid, and (5) convergent-transform hybrid.

Each type of plate boundary produces a characteristic suite of features and Earth materials. This relationship between the kinds of Earth materials formed and the plate tectonic settings in which they are produced provides a major theme of the chapters that follow.

1.5.2 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 rifts form where large-scale horizontal extension occurs in continental lithosphere (Figure 1.5). In such regions, the lithosphere is progressively stretched and thinned. A candy bar being very slowly stretched in two is a crude metaphor. This stretching occurs by brittle, normal, and detachment faulting near the cooler surface and by ductile flow at deeper, warmer levels.

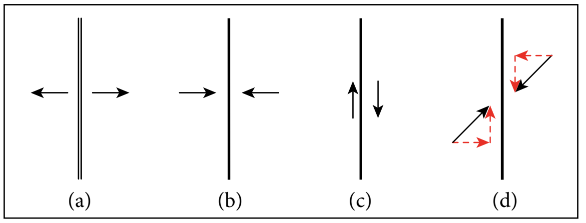

Figure 1.4 Principal types of plate boundaries: (a) divergent; (b) convergent; (c), transform; thick lines represent plate boundaries and black arrows indicate relative motion between plates; (d) hybrid convergent-transform boundary; red arrows show components of convergent and transform relative motion.

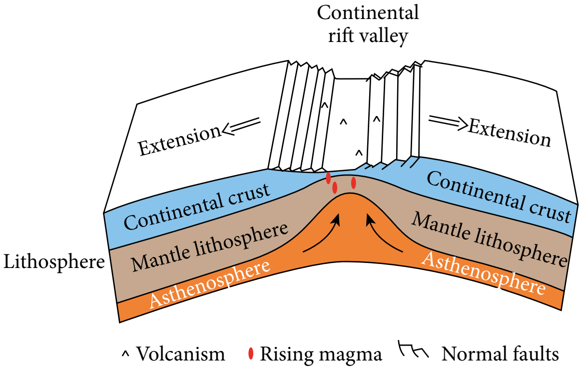

Figure 1.5 Major features of continental rifts include rift valleys, thinned continental crust (blue) and lithosphere (brown) with volcanic-magmatic activity from melts generated in rising asthenosphere (burnt orange).

Long-term extension is accompanied by uplift of the surface as the hot asthenosphere rises under the thinned lithosphere. Rocks near the surface of the lithosphere that rupture along normal and detachment faults 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 diachronous 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 continental fragments by the process of **sea floor spreading** (Figure 1.6). The most recent example of this occurred when the Saudi Arabian Peninsula separated from Africa to produce the Red Sea basin some 5 Ma. Older examples include the separation of India from Africa to produce the northwest Indian Ocean basin (c. 115 Ma) and the separation of the Americas from Africa to produce the Atlantic Ocean basin (beginning c. 190 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 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, the average slope is 3 km/750 km which is about 0.004; the average slope is less than half a degree. We often exaggerate the vertical dimension on profiles

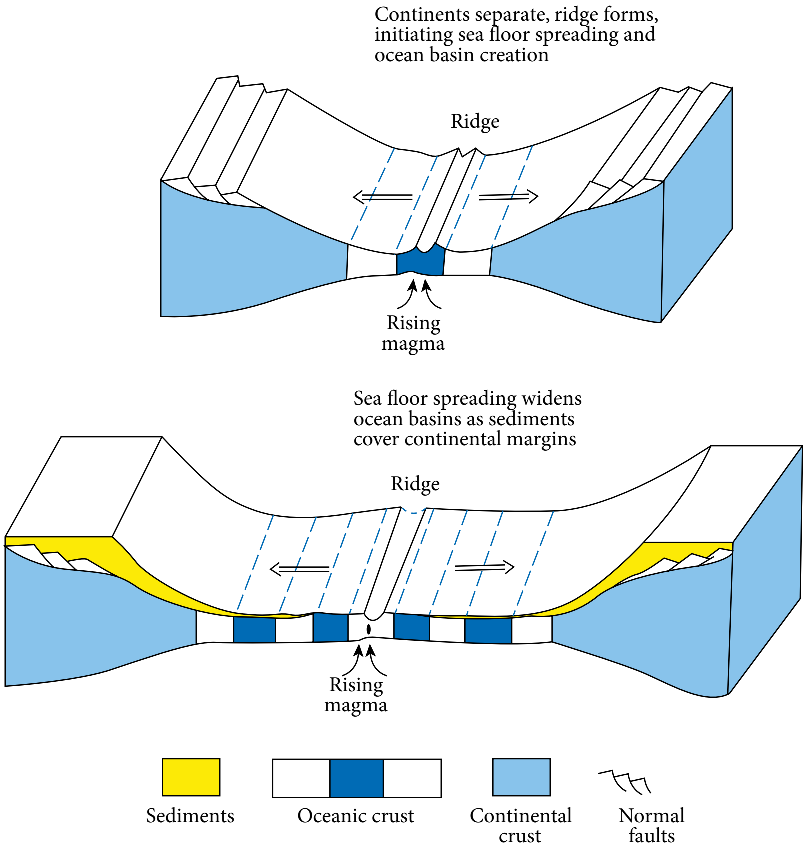

Figure 1.6 Model showing the growth of ocean basins by sea floor spreading from the ridge system following the complete rifting of continental lithosphere along a divergent plate boundary. Separated continental crust (light blue, topped with yellow sediments) and recently formed oceanic crust (dark blue striped pattern) are produced.

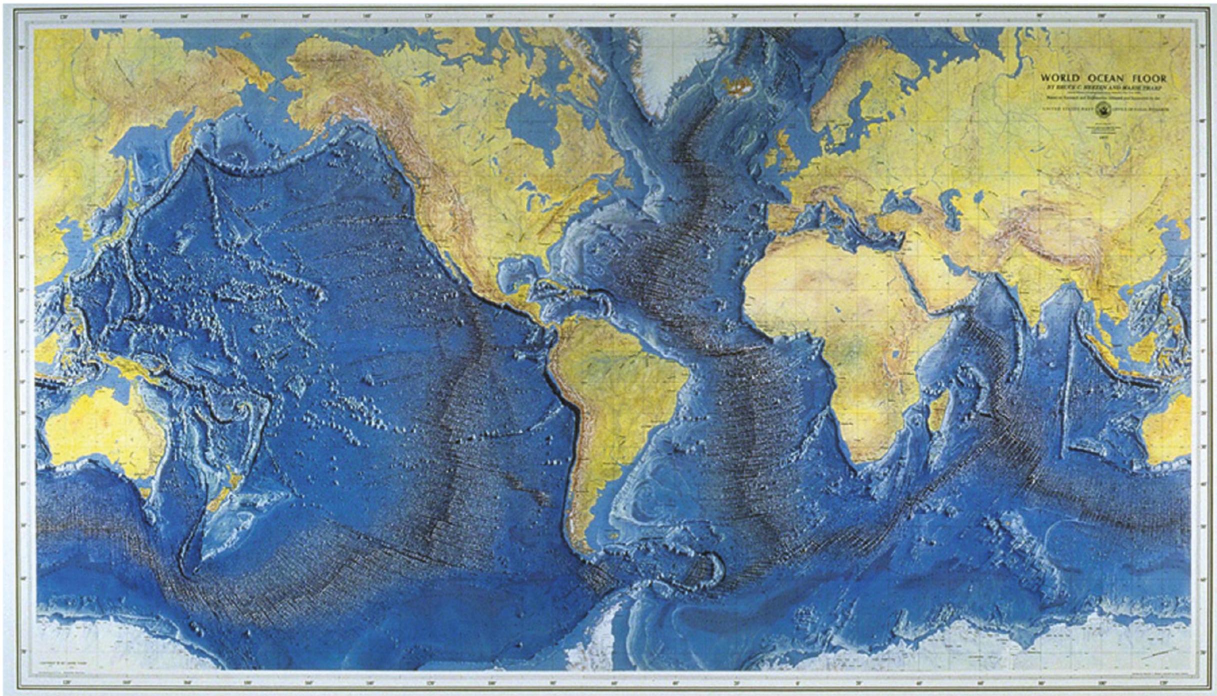

Figure 1.7 Map of the ocean floor showing the distribution of the oceanic ridge system. *Source*: World Ocean Floor Manuscript Map; drawn by Berann, H.C., US Library of Congress, public domain after Heezan, Bruce C. and Tharpe, Marie.

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 commonly marked by a **rift valley**, especially along slower spreading segments. This marks the position of a divergent plate boundary in oceanic lithosphere (Figure 1.7).

One of the most significant discoveries of the twentieth century (Dietz 1961; Hess 1962) was that oceanic crust forms 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 basic plutonic rocks such as gabbro that become layer 3 in oceanic crust. Melts intruded into near vertical fractures above the chamber form the

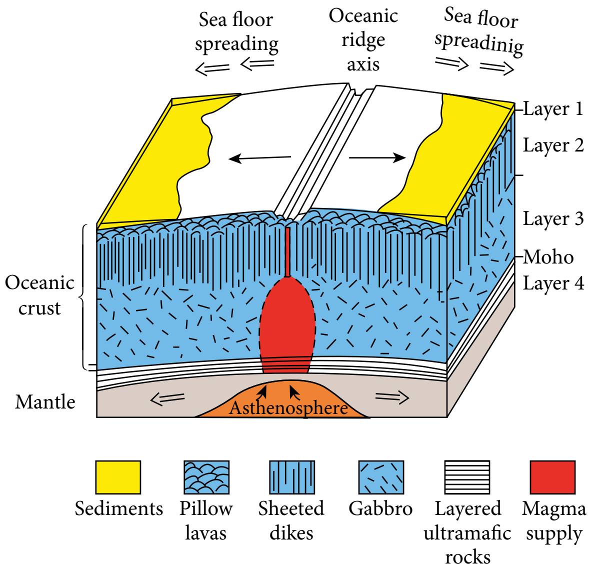

Figure 1.8 The formation of oceanic crust along the ridge axis generates layer 2 pillow basalts and dikes and layer three gabbros of the oceanic crust (blue) and layer 4 mantle peridotites (gray). Sediment deposition atop these rocks produces layer 1 (yellow) of the crust. Sea floor spreading (black arrows) carries these laterally away from the ridge axis in both directions.

gabbroic-basaltic parallel "sheeted" dikes that become layer 2b. Lavas that flow onto the ocean floor commonly form basaltic pillow and sheet lavas that become layer 2a. The marine sediments of layer 1 are deposited atop the basalts as they spread away from the ridge axis. 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 bodies.

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 asthenosphere and the process is repeated, sometimes over >100 Ma. In this way ocean basins grow by sea floor spreading as though new sea floor was being added to two slowly moving 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, these boundaries are also called **constructive** plate boundaries.

As sea floor spreads away from the ridge axis, the crust thickens from above by the accumulation of marine sediments and the lithosphere thickens from below by a process called **underplating** that 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 lithosphere 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

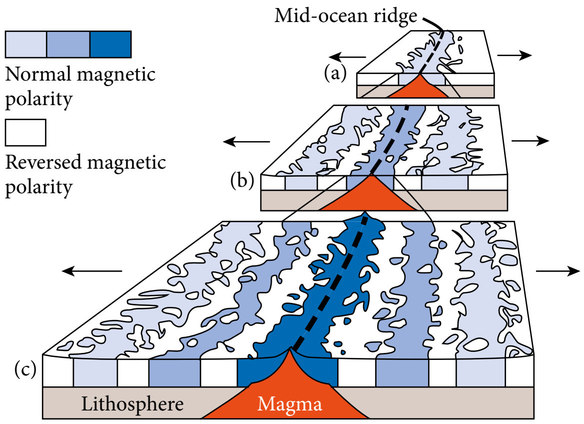

Figure 1.9 Model depicts the production of alternating normal (colored) and reversed (white) magnetic bands in oceanic crust by progressive sea floor spreading and alternating normal and reversed periods of geomagnetic polarity (a–c). The age of such bands should increase away from the ridge axis. *Source*: Courtesy of USGS.

in two and spread away from the ridge axis. New crust formed during the subsequent period of reversed magnetic polarity will form between the two areas of normally polarized crust and the reversely magnetized crust will also split in two. 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, as initially explained by Vine and Matthews (1963).

Sea floor spreading was convincingly demonstrated in the middle to late 1960s by paleomagnetic studies and radiometric dating which 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 and being extended horizontally at divergent plate boundaries over long periods of time, then there must be places where it is undergoing long-term horizontal shortening of similar magnitude. As ocean lithosphere ages and

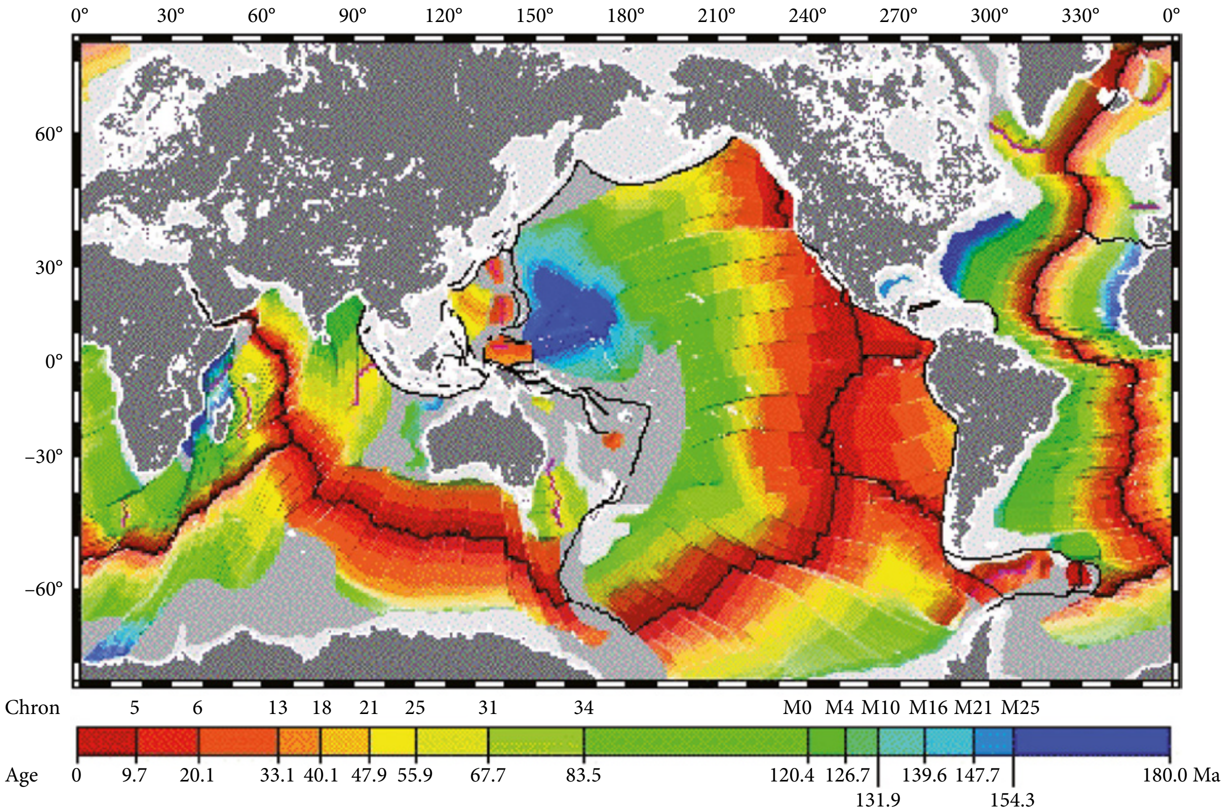

Figure 1.10 World map showing the age of oceanic crust; such maps confirmed the origin of oceanic crust by sea floor spreading. *Source*: From Lamont Doherty Earth Observatory.

continues to move away from oceanic spreading centers, it cools, subsides, and becomes denser over time. The increased density eventually causes the strong ocean lithosphere to become denser than the underlying, weak asthenosphere. As a result, a plate carrying old, cold, dense ocean lithosphere begins to sink downward into the asthenosphere under a more buoyant plate edge, creating a convergent plate boundary.

1.5.3 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

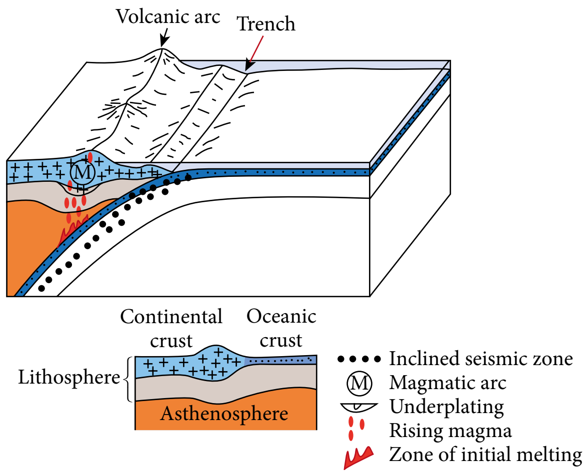

Figure 1.11 Convergent plate boundary, showing trench-arc system, inclined seismic zone and subduction of oceanic lithosphere.

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** or slab. The area where this process occurs is a **subduction zone**. The subducted plate, whose thickness averages 100 km, is generally composed of dense oceanic lithosphere. Subduction is the major process by which oceanic lithosphere is destroyed and recycled into the asthenosphere and deeper Earth at rates similar to its creation 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 large **trench-arc systems** of the kind that encircle most of the shrinking Pacific Ocean (Isacks et al. 1968). **Trenches** are deep, elongate troughs in the ocean floors marked by water depths that can exceed 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 subducted plate to varying degrees. 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 four largest magnitude earthquakes in the past 120 years occurred along inclined seismic zones beneath Chile (1909), Alaska (1964), Sumatra (2004), and Japan (2011). The latter two events produced the devastating 2004 Banda Aceh tsunami which killed some 300 000 people around the Indian Ocean region and the Fukushima tsunami which killed tens of thousands in eastern Japan and destroyed an atomic power plant.

What is the ultimate fate of subducted slabs? Earthquakes occur in subducted slabs to a depth of 660 km, so we know they 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 into the lower mantle. Seismic tomography, which images three-dimensional variations in seismic wave velocity within the mantle, has shed some light on this question, while raising many others. A consensus has emerged (Grand 2002; 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 slab remnants may ultimately be involved in the formation of mantle plumes, as proposed 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 volatile fluids as it undergoes metamorphism which lowers melting temperatures and 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. Granitic–rhyolitic melts may also be generated by partial melting of older continental crust heated by rising magma and volatiles. 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 the many volcanoes in Central America and the Andes of South America are all examples of volcanic arc composite volcanoes that form over Pacific Ocean subduction zones. These volcanic arcs add to the volume of continental crust.

When magmas intrude the crust and solidify below the surface, they produce plutonic igneous rocks that add new continental crust to 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 (Chapter 19) 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. Once produced, the low density of continental crust prevents most of it from being subducted. This helps to explain its preservation potential and the great age that continental crust can achieve (>4.0 Ga).

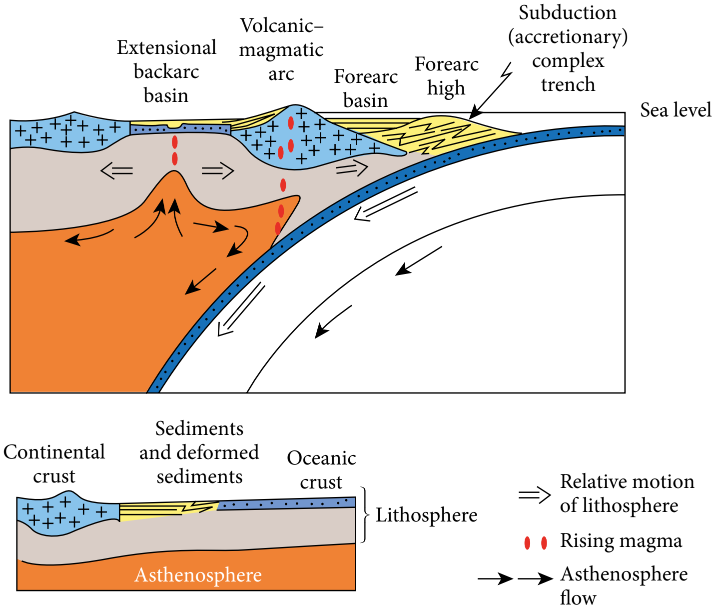

Figure 1.12 Subduction zone depicting details of sediment distribution, sedimentary basins and volcanism in trench-arc system forearc and backarc regions.

Areas of significant relief, such as trench-arc systems are ideal sites for the production and accumulation of detrital (epiclastic) sedimentary rocks (Chapter 13). 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 commonly metamorphosed (Chapters 15 and 18).

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, and the continents, microcontinents or arcs on either side are brought closer together as subduction proceeds. Eventually they converge to produce a continental collision.

When a **continental collision** (Dewey and Bird 1970) occurs, subduction eventually ceases. This occurs because most continental lithosphere is too buoyant to be subducted to great depths for prolonged periods of time. Small amounts may be subducted in this tectonic setting, likely because the density contrasts between the leading edges of the two plates are small. The continental lithosphere involved in the collision may be part of a continent, a microcontinent or a volcanic-magmatic arc complex. Typically the collision occurs over millions to tens of millions of years as irregularly shaped ocean basins close at different times. As convergence continues, the margins of both continental plates are compressed and shortened horizontally and thickened vertically in a manner roughly analogous to what happens to two vehicles in a head-on collision. However, in the case of continents colliding at a convergent plate boundary, the convergence occurs over millions of years. The result is 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** adjacent to the orogenic belt. 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 rocks within it. If the temperatures become high enough, partial melting may occur to produce melts in the deepest

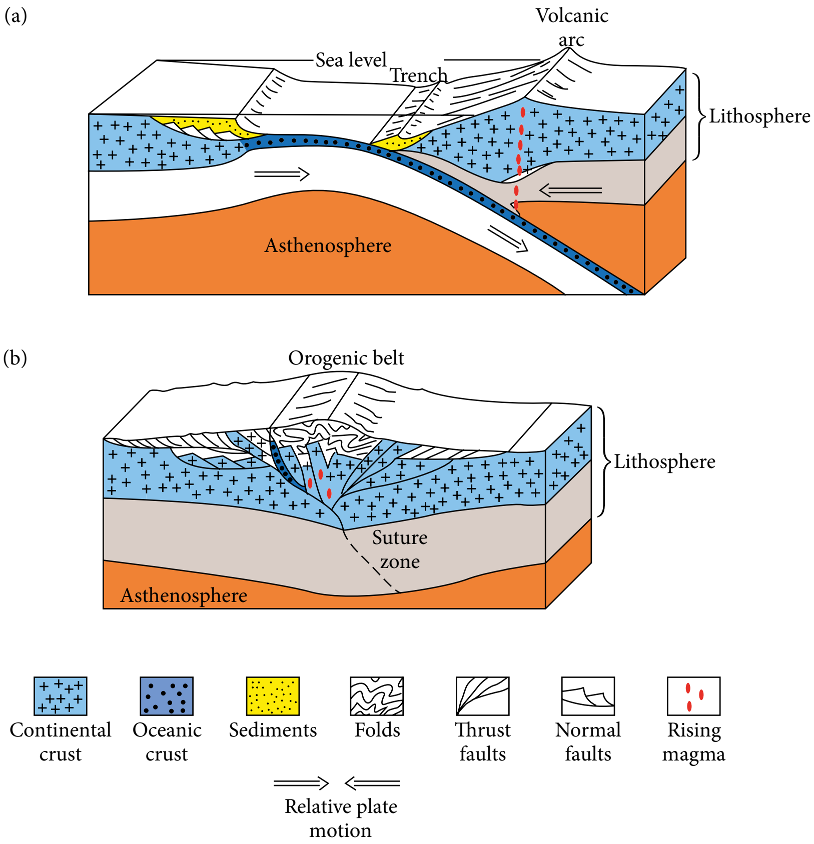

Figure 1.13 (a) Ocean basin shrinks by subduction, as continents on two plates converge. (b) Continental collision produces a larger continent from two continents joined by a suture zone. Horizontal shortening and vertical thickening are accommodated by folds and thrust faults in the resulting orogenic belt.

parts of orogenic belts which rise to produce a variety of felsic igneous rocks.

A 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 micro-continent with Asia has resulted in shortening and regional uplift of the Himalayan Mountain Belt along a series of major thrust faults and produced the Tibetan Plateau. Limestones near the summit of Mt. Everest, Earth's highest mountain, were formed on the floor of the Tethys Ocean that once separated India and Asia. They 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.

1.5.4 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 relative motion

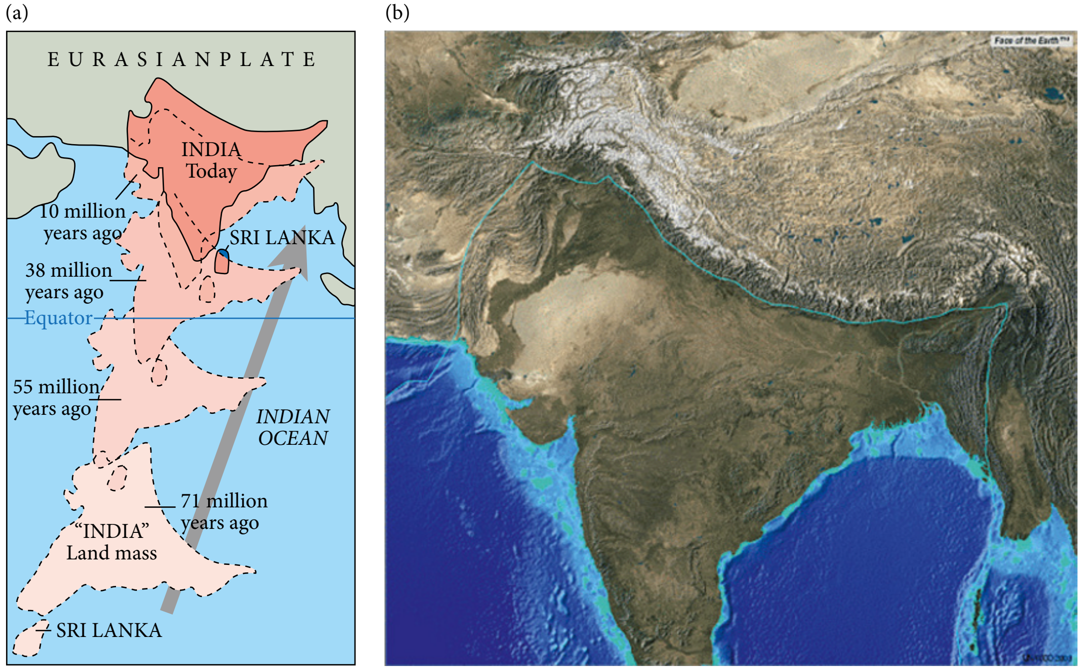

Figure 1.14 (a) Diagram depicting the convergence of India and Asia which closed the Tethys Sea. *Source*: Courtesy of NASA; (b) Satellite image of southern Asia showing indentation of Eurasia by India, the uplift of Himalayas and Tibetan Plateau and the mountains that "wrap around" India. *Source*: From UNAVCO.

along fault systems that is parallel to the plate boundary segment that separates two plates (Figure 1.4). Because the rocks 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 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 parallel relative motion in opposite directions. This would produce shear stresses that result in strike-slip faulting of the lithosphere, frequent earthquakes and the development of a transform fault plate boundary. The exterior portion of fracture zones outside the ridge segments represents 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 exterior portions of fracture zones are appropriately called healed transforms or **transform scars**. They are no longer plate boundaries; instead, they are aseismic (lacking significant earthquakes) intraplate features because the seafloor on either side of them is spreading in the same direction (Figure 1.15). However, they do record the relative motion between two plates at the time they were actively forming.

Continental transform plate boundaries 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 of these are

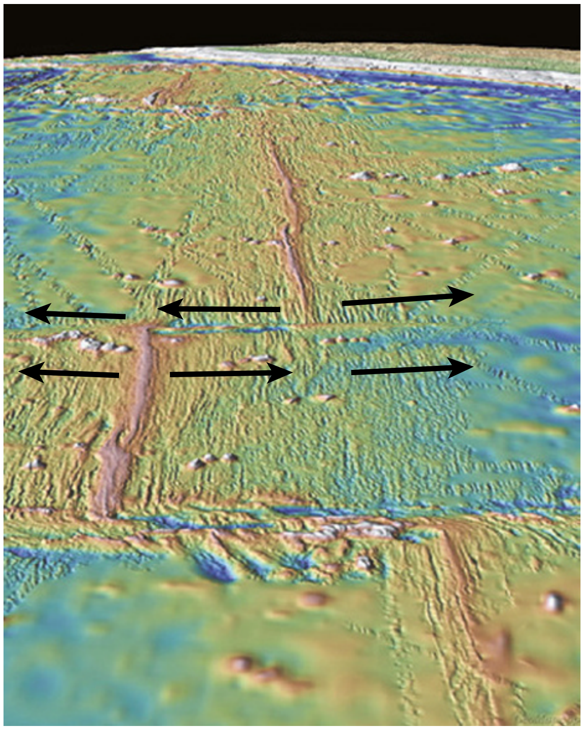

Figure 1.15 Transform faults offsetting ridge segments on the eastern Pacific Ocean floor off Central America. Arrows show directions of sea floor spreading away from the ridge. Oppositely directed arrows (black) indicate transform plate boundaries. Similarly directed arrows (red) indicate intraplate transform scars. *Source*: William Haxby, with the permission of Columbia University Earth Institute; Copyright Marine Geoscience Data System.

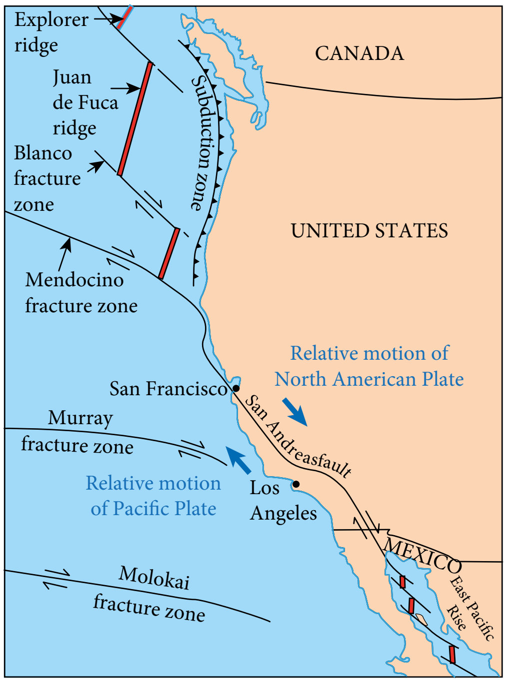

Figure 1.16 Fracture zones, transform faults and ridge segments in the eastern Pacific Ocean and western North America. The San Andreas Fault system is a continental transform fault plate boundary. *Source*: Courtesy of USGS.

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 hybrid boundaries 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. Because they neither create nor destroy large volumes of crust/lithosphere, these boundaries are sometimes referred to as **conservative plate boundaries**.

1.6 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** of volcanoes, such as the Hawaiian Islands (Figure 1.17), as surface expressions of hot spots, which he believed were fixed in one position for long periods. At any one

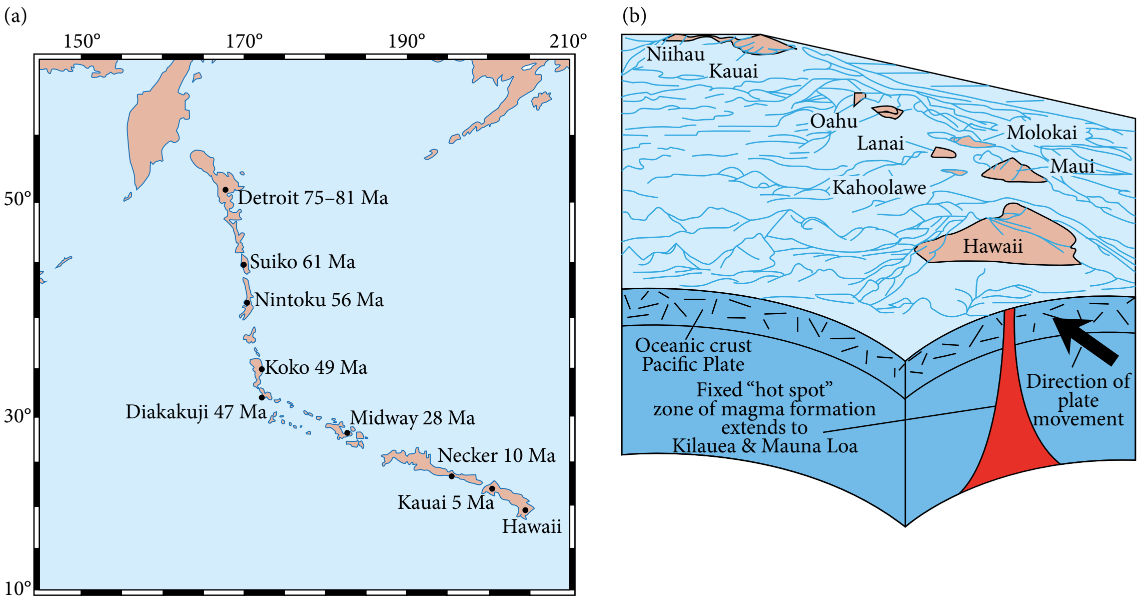

Figure 1.17 (a) Linear seamount chain formed by plate movement over the Hawaiian hotspot and/or hot spot motion. *Source*: Tarduno et al. (2009). © The American Association for the Advancement of Science; (b) Mantle plume feeding surface volcanoes of Hawaiian Chain. *Source*: From USGS.

time, volcanism is largely 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 hotspot and new volcanoes are formed above it. As a result, the age of these seamount chains increases systematically away from the hotspot in the direction of plate motion. For the Hawaiian chain, the data suggested a west-northwest direction of plate motion for the last 47 Ma. However, a change in orientation of the seamount chain to just a few degrees west of north for older volcanoes suggested that sea floor was spreading over the hotspot in a more northerly direction prior to 47 Ma. A similar trend of hotspot volcanism of increasing age over the past 15 Ma extends southwestward from the Yellowstone caldera. Some recent work suggests that the Hawaiian hotspot is not precisely fixed and some southward migration has been documented (Torsvik et al. 2017). Other work suggests that the amount of hotspot drift has been small (Wang et al. 2017). Stay tuned!

In the early 1970s, Morgan (1971) and others suggested that hotspots were the surface expression of fixed, long-lived mantle plumes. **Mantle plumes** were hypothesized to be columns of warm material which rose from near the core–mantle boundary. Some plumes appear to develop plume heads, as they spread outward near the base of the lithosphere (Griffiths and Campbell 1990). These evolving plume heads may be the cause of the apparent drift of hot spots, depending on how they spread out beneath the lithosphere. Later workers hypothesized that deep mantle plumes originate in the ultra-low velocity zone (LVZ) 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, and location (Arndt and Davaille 2013; Condie 2015) 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 in the twenty-first century. Many Earth scientists have offered significant evidence that mantle plumes do not exist (Foulger et al. 2005). 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 others penetrate it and apparently sink to the core–mantle boundary. Recent advances in new imaging methods that use powerful supercomputers have suggested that plumes originating near the base of the mantle do exist beneath many hotspots (French and Romanowicz 2015; Nelson and Grand 2018; Sanni et al. 2019) including Yellowstone, Hawaii, and Iceland, even though they are not always vertical. Wang et al. (2017) demonstrated that most groups of hotspots migrate very slowly, if at all, over time. It is very likely that hot spots are generated by a variety of processes related to mantle convection patterns, but these are still not well understood. Deep Earth tomography will continue to be an exciting area of Earth research over the coming decade.

In this chapter, we have attempted to provide a spatial and tectonic context for the processes which form Earth materials. 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.

CONTENT ASSESSMENT

1. What properties distinguish the following zones of Earth's interior? Elaborate.
   - **a.** continental crust, oceanic crust, and mantle
   - **b.** lithosphere, asthenosphere, and mesosphere
   - **c.** low velocity zone (LVZ), transition zone, and D″ layer
   - **d.** core, outer core, and inner core
2. Detail the processes by which oceanic crust is created and grows through time and contrast these with the processes by which it shrinks and is "destroyed."
3. Explain why the age of oceanic crust generally increases systematically away from the ridge system axis in both directions and the major reasons why there are so many local exceptions to this rule.
4. Describe the three major types of plate boundaries and the features that are associated with and produced by each.
5. Explain how transform faults between two ridge segments form and how, over time, they can generate long fracture zones in oceanic crust. In addition, contrast the earthquake activity on transforms with that on (the external portions of) fracture zones and explain the major reason for this contrast.
6. What is the major process are involved in "collisional tectonics"? Detail the features are produced by and that record such collisional events.

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