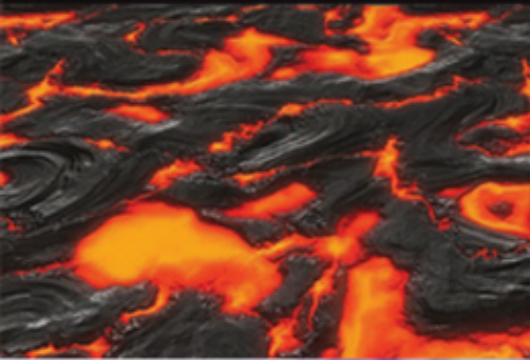


# EARTH MATERIALS



Kevin Hefferan  
John O'Brien

 WILEY-BLACKWELL

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# Earth Materials

*Kevin Hefferan was born and raised in Jersey City, NJ to parents originating from Kiltimagh, County Mayo, Ireland.*

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## **Companion website**

A companion website for this book, with resource materials for students and instructors is available at:

[www.wiley.com/go/hefferan/earthmaterials](http://www.wiley.com/go/hefferan/earthmaterials)

# Earth Materials

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Kevin Hefferan and John O'Brien

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# ***Preface***

Particularly since the 1980s, Earth science at the undergraduate level has experienced fundamental changes with respect to curricula and student goals. Many traditional geology and Earth science programs are being revamped in response to evolving employment and research opportunities for Earth science graduates.

As a result, many colleges and universities have compressed separate mineralogy, optical mineralogy, petrology and sedimentology courses into a one- or two-semester Earth materials course or sequence. This in part reflects the increasing demand on departments to serve students in environmental sciences, remote imaging and geographical information systems and science education. This change has occurred at an accelerating pace over the last decade as departments have adjusted their course offerings to the new realities of the job market. At present, a glaring need exists for a textbook that reflects these critical changes in the Earth science realm.

No book currently on the market is truly suitable for a one- or two-semester Earth materials course. Currently available texts are restricted to specific topics in mineralogy, sedimentology or petrology; too detailed because they are intended for use in traditional mineralogy, sedimentology or petrology course sequences; or not appropriately balanced in their coverage of the major topic areas. This book is intended to provide balanced coverage of all the major Earth materials subject areas and is appropriate for either a one-semester or two-semester mineralogy/petrology or Earth materials course.

The chapters that follow illuminate the key topics involving Earth materials, including:

- Their properties, origin and classification.

- Their associations and relationships in the context of Earth's major tectonic, petrological, hydrological and biogeochemical systems.
- Their uses as resources and their fundamental role in our lives and the global economy.
- Their relation to natural and human-induced hazards.
- Their impact on health and on the environment.

This *Earth Materials* text provides:

- A comprehensive descriptive analysis of Earth materials.
- Graphics and text in a logical and integrated format.
- Both field examples and regional relationships with graphics that illustrate the concepts discussed.
- Examples of how the concepts discussed can be used to answer significant questions and solve real-world problems.
- Up-to-date references from current scientific journals and review articles related to new developments in Earth materials research.
- A summative discussion of how an Earth materials course impacts both science and non-science curricula.

Chapter 1 contains a brief introduction to Earth materials and an overview of system Earth, including a discussion of Earth's interior and global tectonics. This introductory chapter provides a global framework for the discussions that follow.

A minerals section begins with Chapter 2, which addresses necessary background chemistry and mineral classification. Chapter 3 examines the fundamentals of crystal chemistry, phase diagrams and stable and unstable isotopes. Chapter 4 reviews the basic principles of crystallography. Chapter 5 examines mineral formation, macroscopic mineral properties and the major rock-forming minerals. Chapter 6 focuses on the microscopic optical properties of minerals and petrographic microscope techniques.

The igneous rocks section begins with Chapter 7, which discusses the composition, texture and classification of igneous rocks. Chapter 8 addresses the origin and evolution of magmas and plutonic structures. Chapter 9 focuses on volcanic structures and processes. In Chapter 10, the major igneous rock associations are presented in relation to plate tectonics.

The sedimentary rock section begins with Chapter 11, which is concerned with the sedimentary cycle and sedimentary environments. This chapter also focuses on sediment entrainment, transport and deposition agents and the sedimentary structures produced by each. Chapter 12 addresses weathering and soils and the production of sedimentary materials. Chapter 13 examines the composition, textures, classification and origin of detrital sedimentary rocks. Chapter 14 focuses on the composition, texture, classification and origin of carbonate sedimentary rocks, while providing coverage of evaporites, siliceous, iron-rich and phosphatic sedimentary rocks. It ends with a brief synopsis of carbon-rich sedimentary materials, including coal, petroleum and natural gas.

The metamorphic rock section begins with Chapter 15, which introduces metamorphic agents, processes, protoliths and types of metamorphism. Chapter 16 addresses metamorphic structures in relationship to stress and strain. Chapter 17 investigates rock textures and the classification of metamorphic rocks. Chapter 18 concentrates on metamorphic zones, metamorphic facies and metamorphic trajectories in relationship to global tectonics. Lastly, Chapter 19 addresses ore minerals, industrial minerals, gems and environmental and health issues related to minerals.

In addition to information contained in the book, graphics, links and resources for instructors and students are



available on the website that supports the text:  
[www.wiley.com/go/hefferan/earthmaterials](http://www.wiley.com/go/hefferan/earthmaterials).

Our overall goal was to produce an innovative, visually appealing, informative textbook that will meet changing needs in the Earth sciences. *Earth Materials* provides equal treatment to minerals, igneous rocks, sedimentary rocks and metamorphic rocks and demonstrates their impact on our personal lives as well as on the global environment.

# ***Acknowledgments***

We are indebted to Wiley-Blackwell publishers for working with us on this project. We are especially indebted to Ian Francis, who accepted our proposal for the text in 2005 and worked with us closely over the last 4 years, offering both guidance and support. Kelvin Matthews, Jane Andrew, Rosie Hayden, Delia Sandford, Camille Poire and Catherine Flack all made significant contributions to this project.

We gained much useful input from our mineralogy and petrology students at the University of Wisconsin-Stevens Point (UWSP) and New Jersey City University (NJCU). UWSP and NJCU provided sabbatical leave support for the authors that proved essential to the completion of the text, given our heavy teaching loads. We are also particularly thankful to the excellent library staffs at these two institutions.

We are truly appreciative of the many individuals and publishers who generously permitted reproduction of their figures and images from published work or from educational websites such as those created by Stephen Nelson, Patrice Rey and Steve Dutch.

Several reviewers provided critical feedback that greatly improved this book. Reviews by Malcolm Hill, Stephen Nelson, Lucian Platt, Steve Dutch, Duncan Heron, Jeremy Inglis, Maria Luisa Crawford, Barbara Cooper, Alec Winters, David H. Egger, Cin-Ty Lee, Samantha Kaplan and Penelope Morton were particularly helpful.

Lastly we would like to thank our families, to whom we dedicate this text. Kevin's family includes his wife Sherri and children Kaeli, Patrick, Sierra, Keegan and Quintin. John's family includes his wife Anita and sons Tyler and Owen.

# ***Chapter 1***

## ***Earth materials and the geosphere***

- 1.1** Earth materials 1
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- 1.3** Detailed model of the geosphere 3
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- 1.5** Hotspots and mantle convection 17

### **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 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 ( $\text{CaCO}_3$ ) forms by inorganic processes (stalactites and other cave stones) 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 ( $\text{NaCl}$ ). 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  $(\text{Mg,Fe})_2\text{SiO}_4$ . 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.

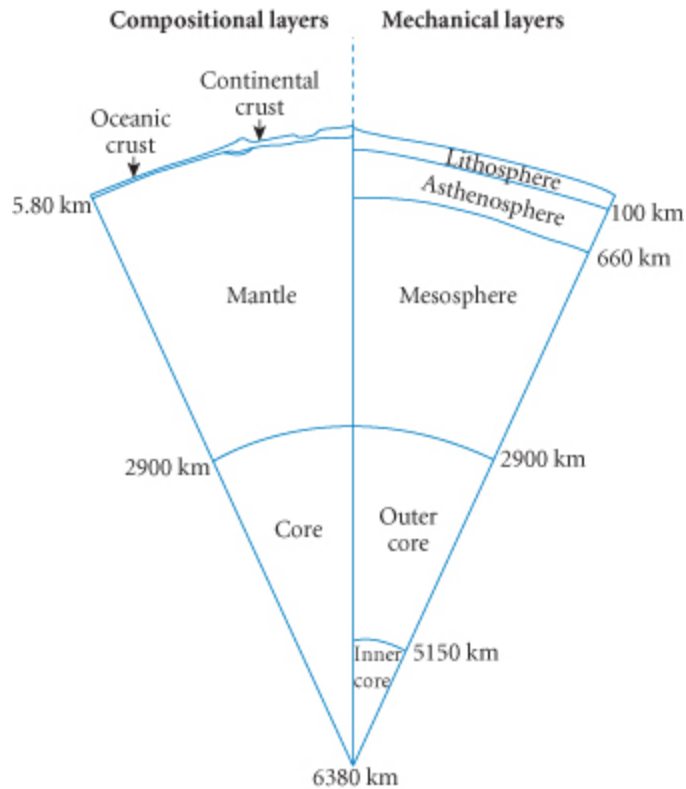
Most of the solid Earth is composed of various types of rock. A rock is an **aggregate of mineral crystals and/or mineraloids**. A **monomineralic** rock consists of multiple crystals of a single mineral. Examples include the sedimentary rock quartz sandstone, which may consist of nothing but grains of quartz held together by quartz cement, and the igneous rock dunite, which can consist entirely of olivine crystals. Most rocks are **polymineralic**; 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.

**[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.



## 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 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 DETAILED MODEL OF THE GEOSPHERE**

### **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.

**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 <sub>2</sub>	Complex; many lighter colored felsic rocks Enriched in K <sub>2</sub> O, Na <sub>2</sub> O and SiO <sub>2</sub> Averages ~60% SiO <sub>2</sub>
Density	Higher; less buoyant Average 2.9–3.1 g/cm <sup>3</sup>	Lower; more buoyant Average 2.6–2.9 g/cm <sup>3</sup>
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

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.

## ***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.

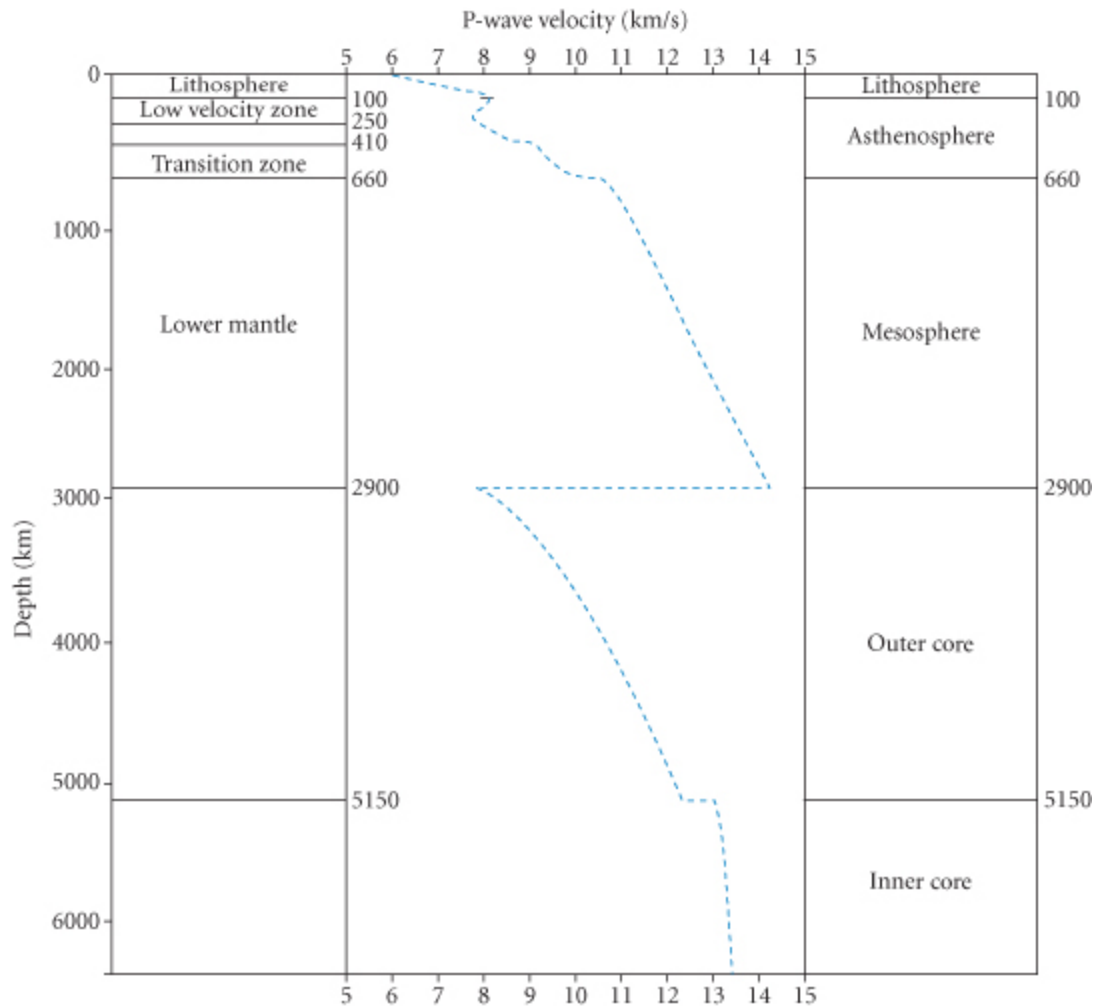
Whereas modern oceans are underlain by oceanic crust younger than 180 Ma, the oldest well-documented continental crust includes 4.03 Ga rocks from the Northwest Territories of Canada (Stern & Bleeker, 1998). Approximately 4 Ga rocks also occur in Greenland and Australia. Greenstone belts (Chapter 18) 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<sub>2</sub> 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 by changes in seismic wave velocity, which reveal that the mantle contains a number of sublayers (Figure [1.2](#)) as discussed below.

**[Figure 1.2](#)** Major layers and seismic (P-wave) velocity changes within Earth, with details of the upper mantle layers.



## ***Upper mantle and transition zone***

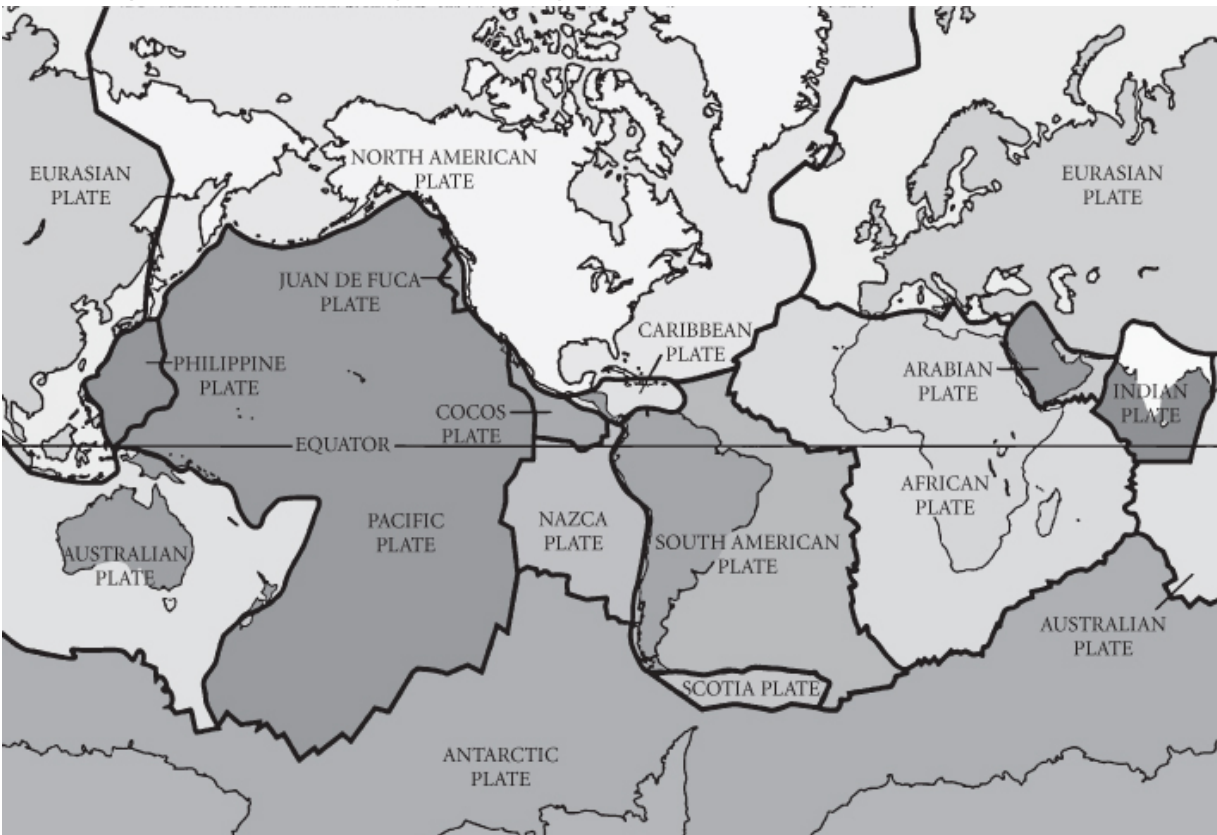
The uppermost part of the mantle and the crust together constitute the relatively rigid **lithosphere**, which is strong enough to rupture in response to stress. Because the lithosphere can rupture in response to stress, it is the site of most earthquakes and is broken into large fragments called plates,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

**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.)*



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 ( $\text{Mg}_2\text{SiO}_4$ ) is transformed to more rigid, incompressible beta spinel ( $\beta$ -spinel), also known as wadleysite ( $\text{Mg}_2\text{SiO}_4$ ). Within the transition zone, wadleysite is transformed into the higher pressure mineral ringwoodite ( $\text{Mg}_2\text{SiO}_4$ ). At ~660 km depth (~24 GPa), ringwoodite and garnet are converted to very rigid, incompressible perovskite  $[(\text{Mg},\text{Fe},\text{Al})\text{SiO}_3]$  and oxide phases such as periclase ( $\text{MgO}$ ). 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  $[(\text{Mg},\text{Fe})\text{O}]$ , magnesiowustite  $[(\text{Mg},\text{Fe})\text{O}]$ , stishovite ( $\text{SiO}_2$ ), ilmenite  $[(\text{Fe},\text{Mg})\text{TiO}_2]$  and ferrite  $[(\text{Ca},\text{Na},\text{Al})\text{Fe}_2\text{O}_4]$  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