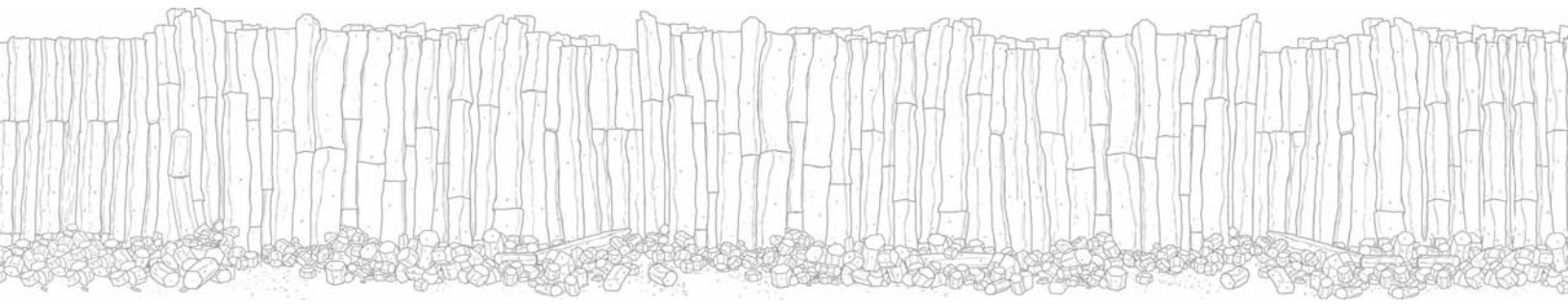


# Some Fundamental Concepts



# Some Fundamental Concepts

## Questions to be Considered in this Chapter:

1. What is igneous petrology and what techniques are used to study igneous rocks?
2. What is the Earth made of, and what are the major subdivisions of the Earth's interior?
3. How did the Earth form?
4. How do the conditions of pressure and temperature vary within the Earth?
5. If we can constrain what composes the Earth and the conditions at depth, what can we initially conclude about melting and the generation of igneous rocks?

**I**n this chapter, I will take the opportunity to generalize and set the stage for the more detailed chapters to come. It will be assumed that you, the student, are familiar with the most basic concepts. You may want to briefly review the chapter on igneous rocks and processes in your introductory geology text, as doing so will refresh your memory and provide an initial “big picture” as we proceed to refine the ideas. Reviewing a chapter on plate tectonics would also help in this regard.

## 1 INTRODUCTION TO IGNEOUS PETROLOGY

*Igneous petrology* is the study of melts (magma) and the rocks that crystallize from such melts, encompassing an understanding of the processes involved in melting and subsequent rise, evolution, crystallization, and eruption or emplacement of the eventual rocks. Origin by crystallization from a melt seems a simple enough criterion for considering a rock to be igneous. But we can only rarely observe the formation of igneous rocks directly, and then only for some surface lavas. The history of the study of igneous rocks is a tribute to the difficulties involved (see Young, 2003, for a comprehensive review). Many humans, including the Greeks, Romans/Italians, Japanese, Icelanders, and Indonesians, have lived with volcanoes. In the Western world, early biblical and Greek references to volcanic phenomena are generally attributed to angry gods. The Greek Hephaestus and later the Roman god Vulcan were gods of fire, and Romans considered volcanoes (particularly the island of Vulcano north of Sicily) as the chimneys of Vulcan's forge. The Greeks and Romans recognized that volcanoes emitted fiery-hot lava (or explosive ash) and that lava cooled to stone. Several ancient scholars even proposed theories of volcanism. Aristotle, apparently impressed by the explosive nature of some Greek eruptions, attributed volcanism to the movement and expulsion of subterranean winds, or “exhalations,” that also gave rise to earthquakes. Others, impressed with the heat, suggested that volcanic subterranean fires required fuel as well, such as sulfur, alum, or asphalt (later extended to include coal and pyrite in 17th and 18th century Europe). Recognizing ancient volcanic deposits and extending the concept of volcanism beyond the very recent past was a much more difficult endeavor. This was probably first done in the Auvergne in south-central France in the mid-18th century, where although volcanism

was no longer active, several conical mountains were still evident, and the summit craters, with flow-like and cindery deposits, were identical to those associated with active volcanoes. Similar extinct volcanic terranes were recognized in Germany and the United Kingdom.

Resistance to the idea of a volcanic origin of basalt was unusually strong in 18th and early 19th century Europe, led primarily by the influential German geologist Abraham G. Werner. He was impressed by numerous examples of flat-lying stratified layers of basalt with no associated volcanic mountain and the crystalline nature of those basalts, which resembled chemical precipitates such as limestone or salt. Thus was born the *Neptunist* school, that, while recognizing the existence of true volcanics, considered most basalt to be deposited by seawater. The word *magma*, from the Greek  $\mu\alpha\gamma\mu\alpha$  (“paste”), was actually introduced in a Neptunist context, not igneous, by Dolomieu in 1794, in the belief that the rocks originating from it were reduced to paste by evaporation. Another school of thought, the *Plutonists*, following James Hutton in the late 18th century, agreed that basalt was igneous, but intrusive, not volcanic. It was not until the early 19th century that sufficient field and experimental work was accumulated to lead to a general consensus on the volcanic origin of basalt. A similar controversy raged in the 19th and early 20th centuries over the origin of granite. The Neptunists believed it, too, was an aqueous precipitate, and later arguments involved those who thought granites originate from a dry melt, those who preferred a water-saturated solution/melt, and those who thought them to be metamorphosed sediments.

These historical controversies are now essentially resolved. Let’s begin, then, by considering some observational criteria for determining that a rock is indeed of igneous origin. Such criteria will be developed further later on, but, by way of introduction, they include:

**1. Field criteria** Intrusive igneous bodies commonly crosscut the “country rocks” into which they intrude, thereby truncating external structures, such as bedding or foliation. They may also exhibit some types of contact effects resulting from the sudden juxtaposition of hot magma and cooler country rocks. When developed, a narrow, fine-grained chilled margin (or “chill zone”) within the igneous body margin or localized baking of the country rocks are good indicators of an igneous origin for plutonic (intrusive) bodies. In addition, we have come to associate certain specific forms of rock bodies with an igneous origin. For example, a strato-volcano, a pahoehoe flow, a sill or laccolith, etc. have become associated with igneous processes, either by direct observation of an igneous event or by the application of some of the criteria mentioned above.

**2. Textural criteria** *Petrography* is the branch of petrology that deals with the description and systematic classification of rocks. By observing thin sections of igneous rocks under the petrographic microscope, we have come to associate a specific *interlocking* texture with slow crystallization from a melt (Figure 1). When crystals are forming



**FIGURE 1** Interlocking texture in a granodiorite. From Bard (1986). Copyright © by permission Kluwer Academic Publishers.

in a cooling melt, they usually develop a nearly perfect crystal form, as the melt provides no obstruction to the preferred crystal shape. As the melt continues to cool, and more crystals form, they eventually begin to interfere with one another and intergrow. The resulting interlocking texture shows interpenetrating crystals, much like a jigsaw puzzle. The relative development of crystal form, inclusions, and interpenetration can be used to infer the sequence in which different mineral species crystallized. Henry Sorby introduced the polarizing microscope to the geological world in the 1850s, and it quickly quelled many of the controversies mentioned above.

Because liquids cannot sustain substantial directional stresses, **foliations** rarely develop in igneous rocks. A common textural criterion for distinguishing an igneous from a high-grade metamorphic crystalline rock in hand specimen is thus based on the *isotropic* texture (random orientation of elongated crystals) of the former. One must use caution, however, when applying this criterion, as some igneous processes, such as crystal settling and magmatic flow, can produce mineral alignments and foliations in igneous rocks.

**Pyroclastic** deposits (those resulting from explosive eruptions) can perhaps be the most difficult to recognize as igneous. Usually the magmatic contribution to these deposits has solidified and cooled considerably before being deposited along with variable proportions of pulverized preexisting rocks caught in the explosion. The actual deposition of pyroclastic material is in large part a sedimentary process, and hence the difficulty in recognition. There is still some debate among geologists as to whether pyroclastics should be considered igneous or sedimentary. They are igneous in the sense that nearly all of their matter crystallized from a melt, even though a proportion may have been earlier volcanic deposits. This is the *pyro* part. They are sedimentary as well, in the sense that they represent solid particles deposited by a

fluid medium: air or sometimes water. This is the *clastic* part. Some geologists have wisely suggested that we avoid this hopeless debate by considering pyroclastics to be igneous going up and sedimentary coming down.

As we initially consider the study of igneous rocks and processes, perhaps we should consider what exactly it is that we want to know. The types of very “broad-brush” questions that we would expect to have answered might include the following: How are melts generated? What is melted, and where? What is produced by this melting? How do the melts so produced crystallize to igneous rocks, and what processes accompany this crystallization? In what way(s) do the liquid and solid portions evolve (change composition) during the process of melting or crystallization? Does the large variety of igneous rock compositions now found at the Earth’s surface result from different sources of melts, or can it be attributed to variations in the processes of melting and crystallization? Is there a relationship between igneous rock type and tectonic setting? If so, what controls this? Finally we might ask, What do we need to know to assess these? In other words, what background and approach does a good modern petrologist need? I would suggest the following as imperative background:

1. A petrologist needs experience looking at rocks and textures. One cannot begin to study rocks without knowing how to recognize, describe, organize, and analyze them. As H. H. Read (1957) quipped, the best geologist is the one “who has seen the most rocks.”
2. Experimental data (based on synthetic and natural samples) are necessary. We can best understand the generation and crystallization of melts by re-creating these processes in the laboratory, simulating the conditions found at depth, and analyzing the results. This also allows us to place some constraints on the physical conditions under which igneous processes may have taken place.
3. Some theory is required, so we can organize and understand the experimental results better and apply those results beyond the exact compositions and conditions of the experiments. A bit of chemistry is necessary, encompassing major elements, trace elements, and isotopes, as is some thermodynamics. As we shall see, these techniques also help us characterize rocks and evaluate source regions and evolutionary processes. A knowledge of physics is also helpful because it permits us to place reasonable constraints on magmatic processes. It is useful to know something

about the viscosity, density, heat capacity, thermal conductivity, and other properties of materials if we are to understand the processes of melting, cooling, crystallization, rise, and emplacement of magmatic systems.

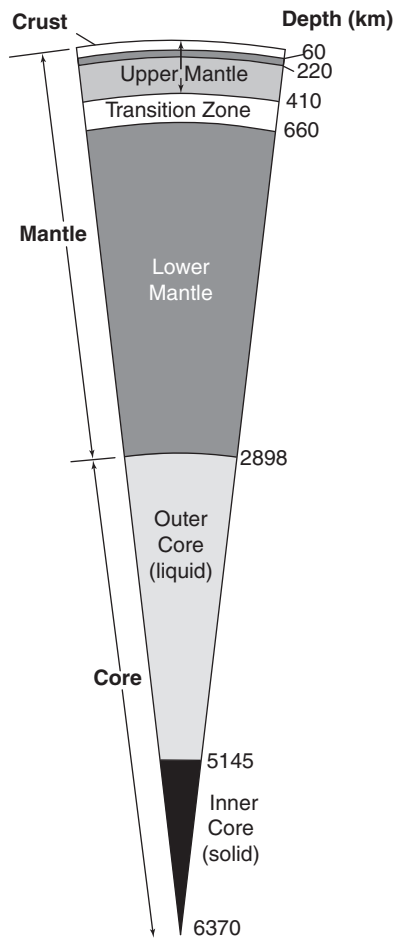
4. We need a knowledge of what comprises the Earth’s interior and the physical conditions that exist there. Melts are created deeper than we can directly observe. If we want to know what is melted and how, we must review what is known about the Earth’s interior and how the constituents and conditions vary with tectonic setting.
5. Finally, we need some practical experience with igneous activity. A literature-based survey of common igneous rocks and processes in nature can provide a framework for all of the above and give a more complete picture.

Although it would perhaps be preferable to develop each of the above skills sequentially, this is not practical, and may be impossible, as these skills are integrated in the scientific process.

On a grand scale, igneous, metamorphic, and sedimentary processes all contribute to a differentiated planet. Igneous processes are by far the most dominant in this regard, as they are largely responsible for the segregation of the crust from the mantle and for the origin of many natural resources. It will benefit us all to keep this large-scale differentiation in mind as we explore the more focused components in what follows.

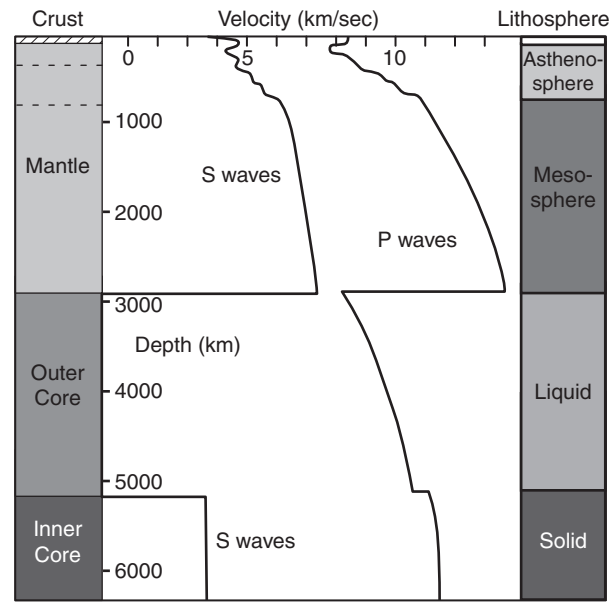
## 2 THE EARTH’S INTERIOR

As mentioned above, virtually all igneous rocks originate by melting of material at some depth within the Earth. All terrestrial rocks that we now find at the Earth’s surface were derived initially from the mantle, although some have since gone through one or more cycles of subsequent sedimentary, metamorphic, and/or igneous processes. If these rocks have an ultimate origin at depth, it follows that we need to know what makes up the Earth if we want to understand their origin more fully. From a *compositional* perspective, the Earth’s interior is subdivided into three major units: the *crust*, the *mantle*, and the *core* (Figure 2). These units were recognized decades ago, during the early days of seismology, because they were separated by major discontinuities in the velocities of *P* (compressional) and *S* (shear) waves as they propagate through those layers in the Earth (Figure 3).



**FIGURE 2** Major subdivisions of the Earth.

The **crust** comprises about 1% of the volume of the Earth. There are two basic types of crust—*oceanic* and *continental*—and both are too thin to represent accurately on Figure 2 (even the thickest continental crust would be thinner than the uppermost line). The thinner of the two, the **oceanic crust** (about 10 km thick), has an essentially basaltic composition. Because plate tectonics is creating oceanic crust at mid-ocean ridges and consuming it at subduction zones, the oceanic crust is continually being renewed and recycled. The oldest oceanic crust is in the southwest Pacific and is about 160 Ma old. The **continental crust** is thicker: typically 30 to 45 km beneath stable areas but generally 50 to 60 km thick in orogenic areas and extending up to 90 km at a few localities. It is also more heterogeneous, including all sorts of sedimentary, igneous, and metamorphic rocks. A *very crude* average composition of the continental crust would be represented by a granodiorite. Continental crust covers about 40% of the Earth's surface. Unlike the oceanic crust, it is too buoyant to subduct far. The amount of continental crust has been increasing over the past 4 Ga. Some continental crust is thus very old, whereas some is quite new. The stable continents (**cratons**) consist of more ancient crystalline **shields** and stable **platforms**. Platforms typically have basement crystalline rocks (typically younger than shields, but not necessarily so) overlain by a few kilometers of sedimentary rocks. Several



**FIGURE 3** Variation in *P* and *S* wave velocities with depth. Compositional subdivisions of the Earth are on the left, and rheological subdivisions are on the right. After Kearey and Vine (1990). Reprinted by permission of Blackwell Science, Inc.

marginal **orogenic belts** may also flank cratons, typically associated with subduction, which add to the continental crust over time. The lower crust is believed to be depleted in many of the more mobile elements and to have a more mafic character than the upper crust. A seismic discontinuity in the lower continental crust (the Conrad discontinuity) is recognized in some areas and may mark the transition between this deeper crust and the shallower sialic portion. It was once believed that the Conrad discontinuity was continuous and separated a more granitic shallow crust from a basaltic lower crust (similar to the oceanic crust) that formed the continental base, but we now realize that the upper and lower continental crust is much more heterogeneous. The base of the continents is not always sharply defined seismically and may locally be gradual and even have a layered transition into the sub-continental mantle. There will be lots more to say about the nature of the continental crust and the igneous and metamorphic processes associated with it throughout this text. For a good summary of continental structure, see Sleep (2005), and for the lower portion, see Fountain et al. (1992).

Immediately beneath the crust, and extending to nearly 3000 km, is the **mantle**, comprising about 83% of the Earth's volume. The boundary, or discontinuity, between the crust and mantle is called the Moho, or M discontinuity (shortened from Mohorovičić, the name of the Balkan seismologist who discovered it in 1909). At this discontinuity, most readily observed beneath oceanic crust, the velocity of *P* waves increases abruptly from about 7 to over 8 km/sec. This results in refraction, as well as reflection, of seismic waves as they encounter the discontinuity, making it relatively simple to determine the depth. The mantle is composed predominantly of Fe- and Mg-rich silicate minerals.

Within the mantle, several other seismic discontinuities separate layers that are distinguished more by physical than chemical differences. The shallowest such layer, between 80 and 220 km, is called the **low-velocity layer** because within it, seismic waves slow down slightly, as compared to the velocity both above and below the layer (Figure 3). The slowness of seismic waves is unusual because seismic velocities generally increase with depth because they propagate more readily through more compacted (hence more incompressible and rigid) materials. The reason seismic waves slow down in the low velocity layer is attributed to 1 to 5% partial melting of the mantle. The melt probably forms a thin discontinuous film along mineral grain boundaries, which retards the seismic waves. The melt also weakens the mantle in this layer, making it behave in a more ductile fashion. The low-velocity layer varies in thickness, depending on the local pressure, temperature, melting point, and availability of H<sub>2</sub>O.

Below the low-velocity layer we encounter two more seismic discontinuities within the mantle. The 410-km discontinuity is believed to result from a phase transformation in which olivine (the major mineral constituent of the upper mantle) changes from the well-known (“ $\alpha$ -phase”) structure to *wadsleyite* (“ $\beta$ -phase”) and then to *ringwoodite* (“ $\gamma$ -phase”) with an isometric spinel-type structure. At 660 km, the coordination of Si in mantle silicates changes from the familiar IV-fold to VI-fold, and the dominant silicate becomes an (Mg,Fe)SiO<sub>3</sub> magnesium silicate with a perovskite-like structure, and the excess Mg and Fe form an (Mg,Fe)O oxide called either *magnesiowüstite* or *ferropericlase*. This latter transition, of course, is not a simple a  $\rightarrow$  b phase transformation, but an a  $\rightarrow$  b + c reaction. Both the 410-km and 660-km transitions result in an abrupt increase in the density of the mantle, accompanied by a jump in seismic velocities.

Below the 660-km discontinuity, the velocities of seismic waves increase fairly uniformly with depth (Figure 3). At the very base of the mantle is a ~200 km thick heterogeneous layer of anomalously low seismic velocity called the **D''** layer. A thin (~40km), apparently discontinuous layer with even lower velocities has also recently been resolved at the mantle–core boundary, most clearly beneath the central Pacific (Garnero and Helmberger, 1995, 1996). The nature of the D'' and 40-km sublayer are not entirely clear, but their properties are sufficiently anomalous to require more than a thermal boundary perturbation, and they probably represent a layer of different composition (and hence greater density) than the overlying mantle. A popular proposition is that they represent an accumulation of dense “dregs” of subducted oceanic crust that has settled to the base of the mantle (Christensen and Hofmann, 1994).

Beneath the mantle is the **core**. The mantle–core boundary is a profound chemical discontinuity at which the silicates of the mantle give way to a much denser Fe-rich metallic alloy with minor amounts of Ni, S, Si, O, etc. The outer core is in the liquid/molten state, whereas the inner core is solid. The composition of the inner core and outer core is probably similar. The transition to a solid is a response to increased pressure with depth, which favors

the solid state. *S*-waves cannot propagate through a liquid because liquids cannot resist shear. Although *S*-waves are only slowed by the thin liquid films in the low-velocity layer, they disappear entirely as they reach the outer core (Figure 3). *P*-waves slow in the liquid core and refract downward, resulting in the seismic “shadow zone,” a ring-like zone in which earthquake *P*-waves don’t reach the surface of the Earth on the side away from which they originated.

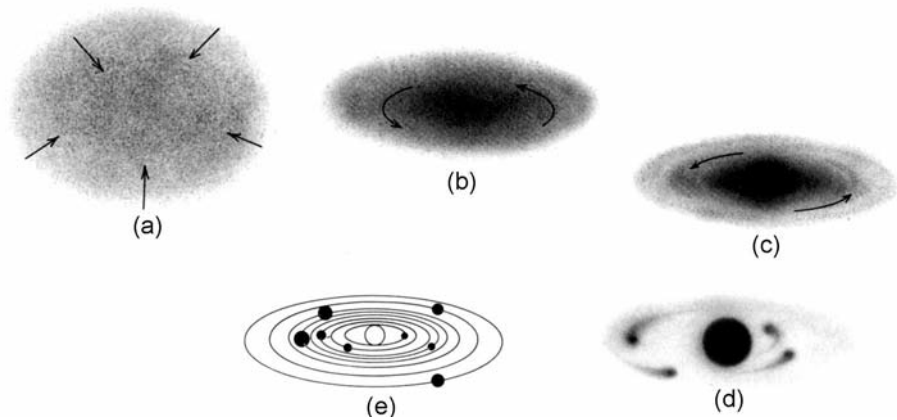
The two types of crust, the mantle, and the core are distinguished on the basis of composition. An alternative way to consider the subdivisions of the Earth is based on *rheological* properties (right side of Figure 3). Using these criteria (how materials respond to deformation), we can consider the crust plus the more rigid portion of the uppermost mantle above the low-velocity layer to behave as a strong, coherent unit, collectively called the **lithosphere**. Oceanic lithosphere is thin (~50 km) near warm mid-ocean ridges and thickens to about 110 km when cool and mature. The lithosphere is 200 to 250 km thick under the stable continental shields (McKenzie and Priestly, 2008). The more ductile mantle immediately below the lithosphere is called the **asthenosphere** (from the Greek *asthenes*, “without strength”). The asthenosphere is important to plate tectonics because the ductility is thought to provide the zone of dislocation upon which the rigid lithospheric plates may differentially move. From a rheological standpoint, the mantle below the asthenosphere is called the **mesosphere**. The lithosphere–asthenosphere–mesosphere boundaries are all within the mantle and correspond to the transition from rigid to ductile and back to less ductile material with depth. The transitions are somewhat gradual and difficult to resolve seismically, particularly the bottom of the ductile layer. The asthenosphere is probably about 150 km thick. The rheological nature of the mesosphere is not well known, but seismic waves are not greatly attenuated, suggesting that this layer is relatively strong. The liquid outer core and solid inner core are of course distinguishable on a mechanical basis.

### 3 ORIGIN OF THE SOLAR SYSTEM AND THE EARTH

Now that you have some idea of what comprises the Earth, it is interesting to speculate on how it got that way. The following scenario summarizes the most generally accepted theories on the origin of the solar system. It will be presented as fact only to avoid the constant use of disclaimers and indefinite phrases. Remember, however, that this is only a collection of internally consistent ideas by which we explain what we now observe, although the extent of this consistency does lend credence to the models.

The most popular model for the origin of the universe has the Big Bang occurring between 12 to 15 Ga before present (b.p.). According to radiometric dating of meteorites, the solar system began to form about 4.56 Ga b.p. as a huge cloud of matter called the **solar nebula** (Figure 4).

**FIGURE 4** Nebular theory of the formation of the solar system. (a) The solar nebula condenses from the interstellar medium and contracts. (b) As the nebula shrinks, its rotation causes it to flatten to a disk (c), with most of the matter concentrated toward the center as the primordial sun (d). Outer solid particles condense and accrete to form the planets (e). From Abell et al. (1988). Copyright © by permission Saunders.



The nebula consisted mostly of molecular  $H_2$  plus some He and minor Be and Li (the only products of the Big Bang). A further 2% comprised heavier elements, including some other gases and fine solid particles (“dust”), presumably created by nuclear synthesis reactions in earlier nearby stars and supernovae. The nebular cloud began to collapse slowly because of the gravitational interactions of its constituents. Because it was rotating, it flattened to a disk-like shape as a result of centrifugal forces, with 1 to 10% of the mass constituting the central disk. The balance between gravitational collapse, centrifugal force, and conservation of angular momentum resulted in the majority of the mass losing angular momentum and falling to the center of the disk, eventually to form the sun. In the area that eventually became the inner terrestrial planets, small particles collided and aggregated, gradually forming larger bodies that swept up more material, growing to form meter- to several kilometer-sized bodies, called **planetesimals**. The gravitational collapse of the mass and its compression generated considerable heat, eventually reaching the stage where nuclear synthesis (fusion) of hydrogen to helium became possible and the sun became a star.

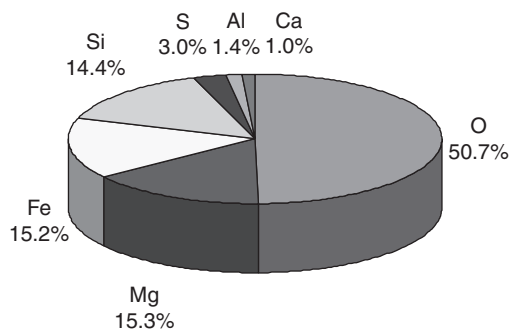
The first 100,000 years witnessed a very rapid evolution of the “proto-sun,” accompanied by high luminosity caused by the heat generated by the initial contraction. When the compression was nearly over, the sun entered the **T-Tauri stage**, characterized by less vigorous activity, lasting up to 10 Ma. The **solar wind**, a stream of charged particles, changed character during the T-Tauri stage and began to emanate radially outward from the sun rather than spirally from the poles. The nebula may have lost about half of its initial mass during this stage.

Of the remaining material, 99.9% of the mass collapsed to form the sun, and the other 0.1%, with the majority of the angular momentum, remained in the disk. The disk material had sufficient mass to contract to the median plane, where it eventually separated into localized accumulations that formed the planets. The process of planetary accretion took place within a strong temperature and pressure gradient generated by the early sun. As a result, the more volatile elements within the solid particles of the nebula vaporized in the inner, hotter portion of the solar system. The vapor particles were then stripped off by the intense T-Tauri solar wind and condensed directly to solids further outward, where the

temperature was lower. Only the larger planetesimals survived this intense activity in the inner solar system. The actual condensation temperatures (and hence the distance from the sun at which condensation took place) depended upon the particular elements or compounds involved. Only the most refractory elements survived or condensed in the innermost zone, whereas the more volatile constituents were moved further outward. As a result, then, primarily due to the temperature gradient and solar wind, the nebula experienced a chemical differentiation based on condensation temperatures. Refractory oxides such as  $Al_2O_3$ , CaO, and  $TiO_2$  either failed to volatilize at all or condensed quickly in the innermost portions of the solar system. Fe-Ni metal alloys, Fe-Mg-Ni silicates, alkali metals and silicates, sulfides, hydrous silicates,  $H_2O$ , and solids of ammonia, methane, etc., condensed and concentrated progressively outward. The distance beyond which the very volatile compounds such as water and methane condensed has been referred to as the **snow line**. Apparently, a gradient of decreasing pressure outward from the center of the nebula also had an effect, principally on the relative condensation temperatures of Fe metal versus silicates, and thus on the Fe/Si ratio (and oxygen content) of the planets.

The condensed solids continued to accrete as planetesimals. In the inner portion of the solar system, the more refractory planetesimals further accumulated and formed the **terrestrial** (Earth-like) planets (Mercury, Venus, Earth, and Mars) as well as the parent bodies that produced the present asteroids and meteorites. In the outer portions, beyond the snow line, the large **gaseous** planets formed. Pluto was considered a planet until August 2006, when it was demoted following the discovery of a similar and even larger object (subsequently named Eris) orbiting the sun. Pluto and Eris are two of many recently discovered objects with orbits beyond Neptune. They are now considered part of the *Kuiper belt* of icy objects (frozen methane, ammonia, and  $H_2O$ ). As astronomers considered alternatives to increasing the number of planets to accommodate these bodies, they wanted to call the larger of them “plutons,” but geologists objected loudly to the theft of our term. They are now called “minor solar system bodies,” and poor Pluto, no longer a planet, is reduced to “number 134340.”





**FIGURE 5** Relative atomic abundances (by mass) of the seven most common elements that comprise 97% of the Earth's mass.

From this very brief sketch, it seems clear that the composition of a planet is in large part the result of rather specific conditions that existed at a particular radial distance from the center of the solar nebula during the first 10 Ma of stellar evolution. The composition of the Earth is largely a result, then, of the nature of the ancient supernova that “seeded” the solar nebula with solid particles and the evaporation/condensation processes associated with the temperature at Earth's particular distance from the T-Tauri sun. Thus we would not expect the Earth's composition to be equal to that of other planets or to that of the solar nebula as a whole.

The differentiation process that produced the chemical variation across the solar system was not perfectly efficient. The composition of the Earth is complex, including some of every stable element, and not just those that could condense at our distance from the sun. Some of the varied constituents of the Earth, including the volatiles, were contained in the early planetesimals that were large enough to resist complete vaporization during the hot T-Tauri stage of solar evolution, whereas others may have been added later via impact of comet-like bodies from the outer solar system. Nonetheless, the process described above strongly favored the concentration of certain elements, and only seven elements now account for 97% of the mass of the Earth (Figure 5). These elements are consistent with the solar abundances and condensates that we expect to have formed at the pressures and temperatures at the Earth's position in the nebular gradients described above.

#### 4 DIFFERENTIATION OF THE EARTH

The planetesimals that now form the Earth probably accumulated in a sequential fashion caused by the gravitational accretion of denser ones first, concentrating Fe-Ni alloys and denser oxides toward the Earth's center. Whether or not the Earth got this “head start” toward further differentiation is hard to say, but it differentiated more extensively soon thereafter (probably about 50 Ma after the beginning of the solar system). This extensive differentiation resulted from heating, caused by gravitational collapse, impacts, and concentrated radioactive heat. Eventually the planet heated sufficiently to initiate melting at some shallow depth, probably beneath a solid crust that stayed cooler by radiating heat to

space. Because iron-nickel alloys melt at lower temperatures than silicates, a dense Fe-Ni-rich liquid probably separated first. Once melting began, mobility within the Earth increased. Denser portions of the melts moved downward, whereas lighter portions rose. The gravitational energy released by this process, plus late impacts, probably generated enough heat to melt the entire Earth, with the possible exception of the outermost chilled layer. This layer may also have melted if there was sufficient gaseous atmosphere to retard radiant cooling. The Moon also formed at this early time, probably due to the impact at a glancing angle of a body about one-tenth the size of the Earth (about the size of Mars) and traveling at a velocity of approximately 2 km/sec. Some of the debris from this collision coalesced to form the Moon, and the rest fell back to the Earth (Taylor and Esat, 1996; Cameron, 1997).

The result of the early differentiation process was the Earth separating into layers controlled by density and the chemical affinities of the elements that comprise it. In simplest terms, we can say for now that element behavior is controlled by the configuration of electrons in the outermost shells, which affects their bonding characteristics.

Goldschmidt (1937) proposed that the elements of the Earth tended to incorporate themselves into separate phases, analogous to the layers in ore smelting pots. Although his notion was simplistic by present standards, we have inherited his terms:

- **Lithophile** (“stone-loving”) elements form a light silicate phase.
- **Chalcophile** (“copper-loving”) elements form an intermediate sulfide phase.
- **Siderophile** (“iron-loving”) elements form a dense metallic phase.

A separate phase of **atmophile** elements may also have formed in early Earth as a very minor ocean and atmosphere, but most of these light gaseous elements were not held by the Earth during the earliest stages and escaped into space. Most of the oceans and atmosphere probably accumulated slowly later.

It is simple enough to determine the affinity of every element empirically and use the results to predict the size (thickness) of each layer/reservoir in the early differentiated Earth, but this approach doesn't work very well. For instance, Fe, which should be siderophile, occurs in all three phases. To explain this, we must remember that the atoms are typically ionized, so the requirement of *electric neutrality* must be satisfied as well. We usually concentrate on cations, but anions are equally important. For example, sulfur is obviously required to create a sulfide, so the amount of sulfur dictates the size of the chalcophile layer in smelting pots. Because there was not enough sulfur to satisfy all the chalcophile cations in the Earth, excess chalcophile cations had to go elsewhere. Oxygen is the principal anion in silicate minerals. It combined with silicon for the lithophile layer, but other cations were required before neutrality was achieved. The most common minerals



in the lithophile layer of the early Earth were probably olivine ( $(\text{Fe,Mg})_2\text{SiO}_4$ ), orthopyroxene ( $(\text{Fe,Mg})\text{SiO}_3$ ), and clinopyroxene ( $\text{Ca}(\text{Fe,Mg})\text{Si}_2\text{O}_6$ ). The relative abundance of oxygen thus determined the thickness of the lithophile layer. The inner siderophile layer was determined by the excess of siderophile cations (mostly Fe) left over after neutrality was achieved with O and S. All the other elements, accounting for the remaining 3% of the Earth's mass, went preferentially into one of these layers, in accordance with a particular atom's affinity.

As with the differentiation of the solar system, the Earth's differentiation was certainly not perfectly efficient: not all of the elements are restricted to the predicted layer. Otherwise, we would never find such elements as gold (siderophile), copper (chalcophile), etc., at the Earth's surface today. This may be caused, in part, by a lack of complete equilibrium during the differentiation process, but even if equilibrium is attained, elements typically partition themselves into different reservoirs in less than the most extreme proportions (not all into one reservoir).

After a few hundred million years, this molten, differentiated Earth cooled and mostly solidified to a condition similar to the planet we now inhabit, having a distinct temperature and pressure gradient with depth.

The lithophile, chalcophile, and siderophile layers are not to be confused with the present layers of the Earth: crust, mantle, and core. The core of the modern Earth is the siderophile layer, but the chalcophile component was probably dissolved in the siderophile core and never separated as a distinct phase. Although such a phase does form in smelters, it is much less likely to do so at the high pressures associated with the core. The Earth is not a smelting pot. If a separate chalcophile phase did form, it might be an outermost layer of the outer core.

The mantle certainly represents the early lithophile segregation, but it is unlikely that either the oceanic or the continental crust formed at this point by a large-scale differentiation event in the early Earth (although this probably did happen for the plagioclase-rich highlands of the Moon). If any crust formed from a primordial surface magma ocean, no samples have yet been found. The Earth's crust is believed to have formed later and more progressively. The processes by which the mantle differentiates to produce the crust are predominantly igneous in nature and are occurring to this day. As outlined above, the basaltic oceanic crust is created by partial melting of the mantle at divergent plate boundaries, and most is eventually consumed by subduction and recycled. The oldest non-subducted oceanic crust is only about 160 Ma old. Most of the heterogeneous continental crust was probably created during the Archean via partial melting of mafic source rocks in subduction zone and rift settings, typically with intermediate to silicic products. Archean island arcs and micro-continents are believed to have assembled to form larger stable continental masses by the early Proterozoic. Early continents have assembled to larger "supercontinents" and rifted apart several times. The most recent breakup is that of "Pangea," beginning in the

Jurassic. Continents grow by collision, arc accretion, and other orogenic processes at continental margins. Due to its thickness, buoyancy, and high viscosity, continental crust is not recycled at subduction zones, so the amount has been increasing over the past 4 Ga.

## 5 HOW DO WE KNOW ALL THIS?

If you are now asking yourself how we can possibly know what has just been presented, you're approaching petrology with the right attitude. Theories, such as those concerning the origin of the universe, the solar system, and the Earth, represent the best inferences we can make based on our interpretation of the data. The simplest explanation of all data, without violating physical "laws," is preferred. The more varied the nature of the phenomena a theory explains, the more confidence we place in it. The scenario described above is consistent with the physical "laws" of celestial mechanics, gravity, nuclear synthesis, and so on. It is also consistent with our observations of seismic waves and the nature and composition of the solar system. But rigorous evaluation of these criteria is well beyond our scope. The scenario is intended only as background information, however. The information on the composition and layering of the Earth's interior in Figures 2 and 3 is the final result of the process and is presented as fact. This information is very important to the material that we will address in the pages ahead. After all, if igneous rocks are the products of melting at depth, it might be nice to know with some confidence what is being melted. For petrologic concerns, let's focus our skepticism here for a moment. We have not yet drilled a hole to the mantle (and will never do so to the core) in order to directly sample these materials. And our hypothetical mantle and core are far different from the materials we find at the Earth's surface. What evidence do we have to support the alleged composition and structure of our planet?

First, from careful measurements, we can accurately determine the gravitational constant, and use that, plus the measured moment of inertia of the Earth, to calculate its mass and, from that, the average density. This places several constraints on the materials that make up the Earth. For instance, the average density of the Earth is approximately  $5.52 \text{ g/cm}^3$ . It is relatively easy to observe and inventory the chemical composition of the rocks exposed at the surface of the Earth. But the density of surface rocks is rarely greater than  $3.0 \text{ g/cm}^3$ . The Earth must thus contain a large proportion of material that is much denser than can be accomplished by compression of surface-type rocks due to the increased pressure at depth.

One could come up with a variety of recipes for the dense material at depth, by mixing proportions of atoms of various atomic weights. However, such a random approach would better be guided by having some idea of which elements are more naturally abundant. The Earth must have formed from the solar nebula, so the composition of the nebula must provide us with significant clues to the makeup of our planet. The material that makes up the solar system can

be analyzed from a distance by spectroscopic means. Atoms can be excited by heat or particle interactions and emit characteristic light spectra when they return to their lower-energy “ground state.” The wavelength of light that reaches the Earth can be determined and related to the type of element or compound that emits it. By comparison with spectra of elements measured in the laboratory, the emitting atoms or molecules can be identified. The intensity of the spectral lines is proportional to their concentrations at the source. We thus get a good idea of what elements constitute the sun, other stars, and even other planetary surfaces, and by analogy with these, our own planet.

Figure 6 illustrates the estimated concentrations of the elements in the solar nebula (estimated from certain meteorites, as discussed below). Note the logarithmic scale for concentration, which makes it easier to show the full range of abundances. Hydrogen is by far the most abundant element, as it made up most of the original nebula. Other elements (except He) were synthesized from H in the sun and other stars. The decrease in abundance with increasing atomic number ( $Z$ ) reflects the difficulty of synthesizing progressively larger atoms. Another interesting feature that is clear from Figure 6 is the “sawtooth” nature of the curve. This is in accordance with the Oddo-Harkins rule, which says that atoms with even atomic numbers are more stable, and hence more abundant, than their odd-numbered neighbors.

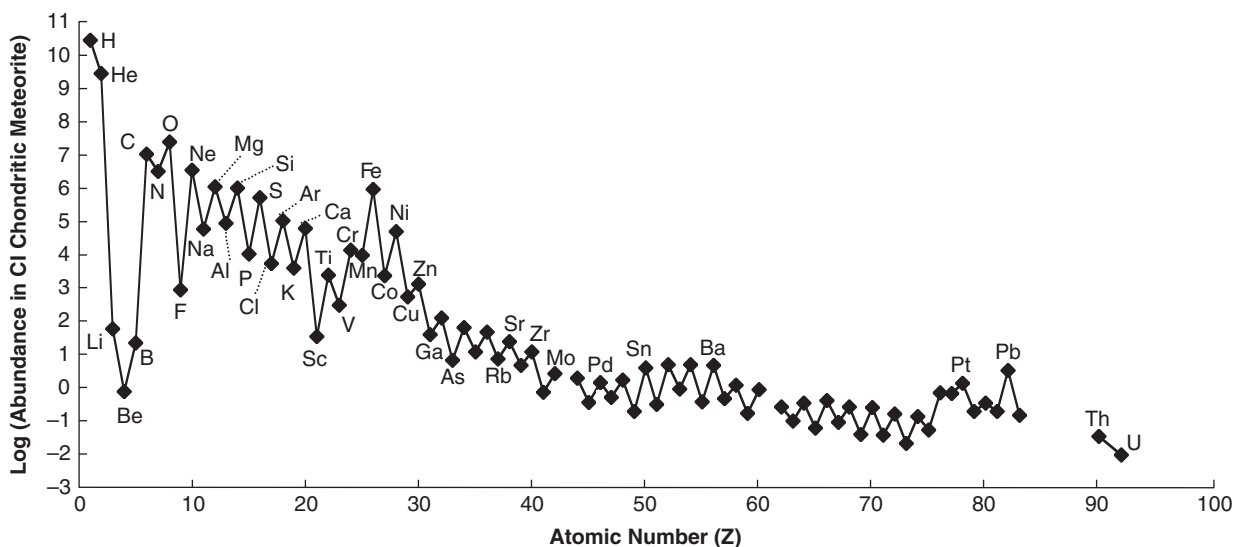
We must assume that the elements that compose the Earth are among the more common elements in Figure 6. For example, Fe, and to a lesser extent Mg and Ni, are much more abundant in the solar system than in the Earth’s crust, so we might infer that these elements are concentrated elsewhere in the Earth. Fe is also dense enough to satisfy the Earth’s high density requirement. In other words, using the data in Figure 6 as a starting point to model a planet with an average density of  $5.52 \text{ g/cm}^3$  should lead us in the direction of the concentrations in Figure 5. Of course, the process is complicated by such inhomogeneities as the radial differentiation of

the solar nebula and density variations and phase changes associated with increasing pressure in the Earth.

Seismic studies place further constraints on the materials that constitute the Earth. The velocities of  $P$  and  $S$  waves in various materials at elevated pressures and temperatures can be measured in the laboratory and compared to seismic velocities within the Earth, as determined from earthquakes or human-made explosions (Figure 3). In addition, reflection and refraction of seismic waves at discontinuities within the Earth provide direct evidence for the Earth’s internal structure and the depths of the discontinuities that subdivide it into crust, mantle, outer core, and inner core, as well as other more detailed features.

Finally, although we haven’t visited the mantle or core for samples, we have had samples delivered to us at the surface (or so we believe). There are a number of rocks found at the surface that we believe to be of mantle origin. In many active and fossil subduction zones, slivers of oceanic crust and underlying mantle are incorporated into the accretionary prism. Thickening of the prism, followed by uplift and erosion, exposes the mantle-type rocks. Xenoliths of presumed mantle material are occasionally carried to the surface in some basalts. Deeper mantle material is believed to come to the surface as xenoliths in diamond-bearing kimberlite pipes. The vast majority of samples found in all these situations are olivine- and pyroxene-rich ultramafic rocks. When many of these samples are partially melted in the laboratory, they produce melts similar to natural lavas that we believe to be mantle derived.

Because of the great density and depth of the Earth’s core, not a single sample of the core has reached the surface. However, we do believe that pieces of the core of other planetesimals have reached Earth in the form of some meteorites. We will briefly discuss meteorites in the next section, as they are quite varied and provide important information about the composition of the Earth and solar system.



**FIGURE 6** Estimated abundances of the elements in the solar nebula (atoms per  $10^6$  Si atoms). After Anders and Grevesse (1989). Copyright © with permission from Elsevier Science.

## 6 METEORITES

Meteorites are solid extraterrestrial objects that strike the surface of the Earth after surviving passage through the atmosphere. Most of them are believed to be fragments derived from collisions of larger bodies, principally from the asteroid belt between the orbits of Mars and Jupiter. They are very important because many are believed to represent arrested early to intermediate stages in the development of the solar nebula that have not undergone subsequent alteration or differentiation like the Earth. They thus provide valuable clues to the makeup and development of the solar system. Meteorites have been classified in a number of ways. Table 1 is a simplified classification, in which I have combined several subclasses to give a general indication of the more important types and the percentages of each from observed falls.

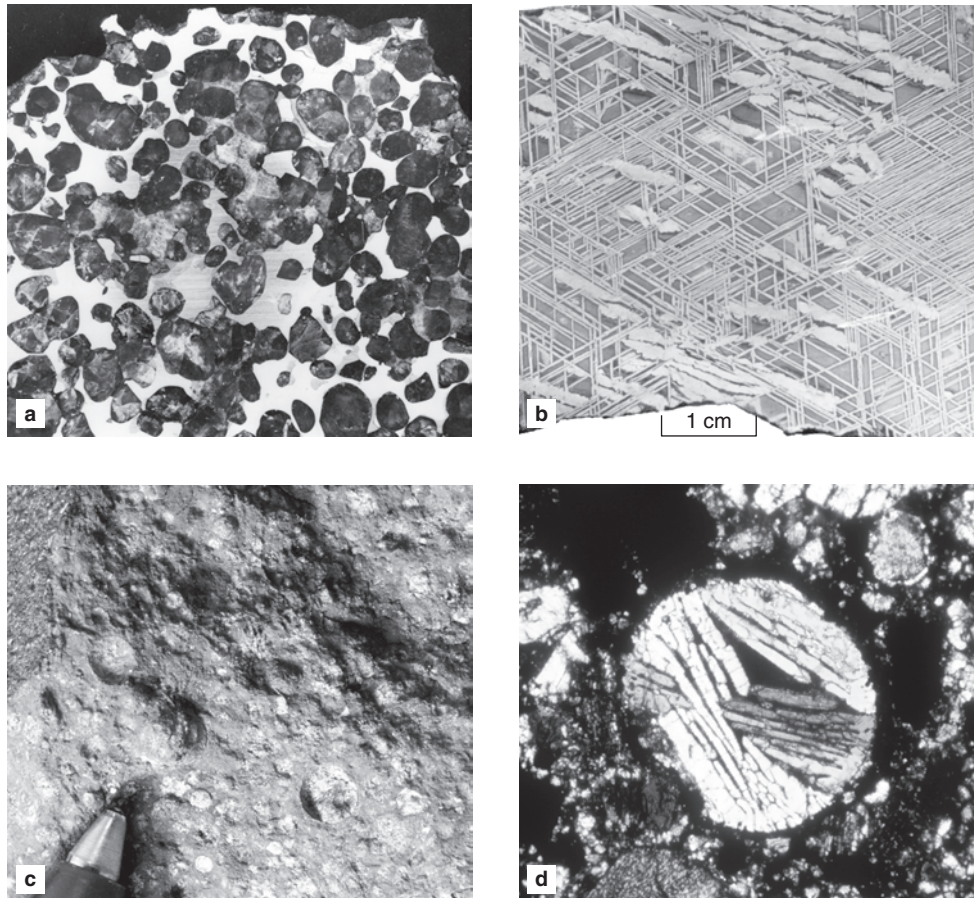
**Irons** (Figure 7b) are composed principally of a metallic Fe-Ni alloy, **stones** are composed of silicate minerals, and **stony-irons** (Figure 7a) contain subequal amounts of each. Because stones look much like terrestrial rocks,

**TABLE 1** Simplified Classification of Meteorites

Class	Subclass	# of Falls	% of Falls	
Irons	All	42	5	
Stony-irons	All	9	1	
Stones	Achondrites	SNC's	4	8
		Others	65	
	Chondrites	Carbonaceous	35	86
		Others	677	

After Sears and Dodd (1988).

they are seldom recognized as meteorites, so irons (quickly recognizable by their density) tend to dominate museum collections. When we consider only specimens collected after an observed fall, however, stones comprise 94% of meteoritic abundance. Iron meteorites are believed to be fragments of the core of some terrestrial planets that have undergone differentiation into concentrations of silicate,



**FIGURE 7** Meteorite textures. (a) Polished section of the stony-iron Springwater (Saskatchewan, Canada) meteorite (light is iron, dark is ~5mm olivines). © Courtesy Carleton Moore, Arizona State University. (b) Widmanstätten texture in the Edmonton (Kentucky) iron meteorite. © Courtesy John Wood and the Smithsonian Astrophysical Observatory. (c) Chondrules in the Allende chondrite meteorite (Mexico). Ball pen tip for scale. (d) Photomicrograph of a 0.5 mm diameter chondrule from the Dhajala (India) chondrite meteorite. The chondrule is composed of skeletal plates of olivine in a finer matrix, including crystal fragments and glass. Other chondrules may be dominated by orthopyroxene, or mixtures of several minerals, including sulfides. Chondrules may be very fine grained or even totally glassy in meteorites that have suffered little or no reheating. Chondrite photographs © courtesy of the Geological Museum, University of Copenhagen.

sulfide, and metallic liquids, in the manner discussed above for the Earth. These meteorites contain portions of siderophile (Fe-Ni alloy) and chalcophile (segregations of troilite: FeS) phases. The Fe-Ni alloy is composed of two phases, kamacite and taenite, which exsolved from a single, homogenous phase as it cooled. The two phases are commonly intergrown in a crosshatched pattern of exsolution lamellae called Widmanstätten texture (Figure 7b). Stony-irons are similar to irons but include a significant proportion of the silicate (lithophile) segregation mixed in. All irons and stony-irons are considered “differentiated” meteorites because they come from larger bodies that underwent some degree of chemical differentiation. Meteorites, however, display a large variation in the degree to which they represent differentiated portions of a planet. The parent bodies, most believed to be present in the asteroid belt, were of various sizes and thus capable of different degrees of differentiation. In addition, the collisions that disrupted the parent bodies into meteoritic fragments have remixed and even brecciated the material that we find in many meteorites.

Stones are further subdivided on the basis of whether they contain **chondrules** (Figure 7c and d), nearly spherical silicate inclusions between 0.1 and 3.0 mm in diameter. At least some chondrules appear to have formed as droplets of glass that have subsequently crystallized to silicate minerals. Stones with chondrules are called **chondrites**, whereas those without are called **achondrites**. As with irons and stony-irons, achondrites are differentiated meteorites. Chondrites, on the other hand, are considered “undifferentiated” meteorites because the heat required to initiate melting and differentiation of a planet would certainly have destroyed the glassy chondrules. The small size of the chondrules indicates rapid cooling (< 1 hr), requiring a cooler nebula at the time of their formation. The chondrules probably formed after condensation but before formation of the planetesimals. Chondrites are thus considered to be the most “primitive” type of meteorites, in the sense that they are thought to have compositions closest to the original solar nebula (hence their use in estimating solar abundances in Figure 6). It has been suggested that all of the inner terrestrial planets formed from a material of average chondritic composition. This has led to the Chondritic Earth Model (CEM), which provides a close fit to the composition of the Earth for most elements, but with a few important differences. For example, the Earth is much denser and must have a higher Fe/Si ratio than chondrites. Models such as the one presented above, based on condensation temperatures as a function of distance from the sun, are much better for explaining the chemical composition of the planets (particularly their variations) than is assuming that some meteorite represents them all.

Further subdivision of meteorites is based on their textures and/or mineral content. There is considerable variety in the overall (“bulk”) composition, as well as in the mineralogy. More than 90 minerals have been found among the stony meteorites, some of which are not found elsewhere on Earth. Some meteorites appear to come from the moon and neighboring planets. The SNC meteorites, for example, appear to be from Mars. Given this variety, the study of meteorites can

provide us with valuable information on the chemical composition of the solar system and its constituents.

## 7 PRESSURE AND TEMPERATURE VARIATIONS WITH DEPTH

We now have a good idea of what comprises the Earth and how it got there. If we are to proceed to an understanding of melts (and later of metamorphism), we should next attempt to understand the physical conditions (pressure and temperature) that occur at depth, so that we can appreciate how these materials respond and behave. As depth within the Earth increases, both pressure and temperature increase as well. Pressure increases as a result of the weight of the overlying material, whereas temperature increases as a result of the slow transfer of heat from the Earth’s interior to the surface.

### 7.1 The Pressure Gradient

The pressure exerted in a ductile or fluid medium results from the weight of the overlying column of the material. For example, the pressure that a submarine experiences at depth is equal to the weight of the water above it, which is approximated by the equation:

$$P = \rho gh \quad (1)$$

where:  $P$  = pressure

$\rho$  = the density (in this case, that of water)

$g$  = the acceleration caused by gravity at the depth considered

$h$  = the height of the column of water above the submarine (the depth)

Because water is capable of flow, the pressure is equalized so that it is the same in all directions. The horizontal pressure is thus equal to the vertical pressure (the axis along which the imaginary column of water would exert itself). This equalized pressure is called **hydrostatic** pressure. Near the surface, rocks behave in a more brittle fashion, so they can support unequal pressures. If the horizontal pressures exceed the vertical ones (or vice versa), rocks may respond by faulting or folding. At depth, however, the rocks also become ductile and are capable of flow. Just as in water, the pressure then becomes equal in all directions, and is termed **lithostatic** pressure. Equation (1) will apply then, too, with  $\rho$  being the density of the overlying rock.

The relationship between pressure and depth is complicated because density increases with depth as the rock is compressed. Also,  $g$  decreases as the distance to the center of the Earth decreases. A more accurate approach would be to use a differential form of the  $P$ -depth relationship, complete with estimates of the variation in  $g$  and  $\rho$ , and integrate it over the depth range. However, the changes in  $g$  and the density of a given rock type are relatively minor in the crust and upper mantle, and they also tend to offset each other, so Equation (1) should suffice for our needs. Only when the rock type



changes, as at the Moho, would a different value of  $\rho$  be required. One need only calculate the pressure to the base of the crust, using an appropriate average crustal density, and continue with depth using a density representative of the mantle.

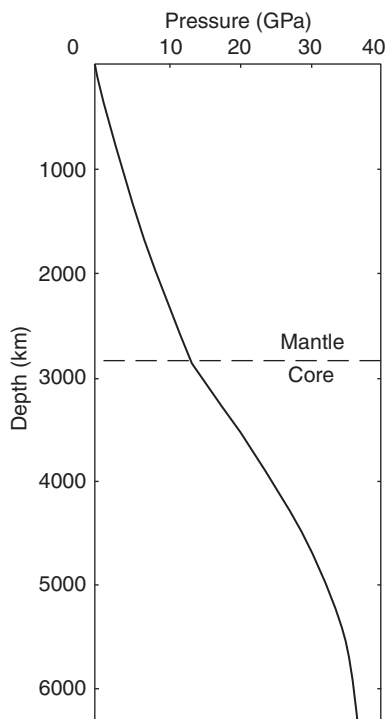
For example, a reasonable estimate of the average density of the continental crust is  $2.8 \text{ g/cm}^3$ . To calculate the pressure at the base of 35 km of continental crust, we need only substitute these data into Equation (1), being careful to keep the units uniform (refer to the front inside cover for units and constants):

$$\begin{aligned} P &= \frac{2800 \text{ kg}}{\text{m}^3} \cdot \frac{9.8 \text{ m}}{\text{s}^2} \cdot 35,000 \text{ m} \\ &= 9.6 \times 10^8 \text{ kg}/(\text{m s}^2) \\ &= 9.6 \times 10^8 \text{ Pa} \approx 1 \text{ GPa} \end{aligned}$$

This result is a good average pressure gradient in the continental crust of 1 GPa/35 km, or about 0.03 GPa/km, or 30 MPa/km. Because of (upward) round-off, this gradient is also suitable for oceanic crust. A representative density for the upper mantle is  $3.35 \text{ g/cm}^3$ , resulting in a mantle pressure gradient of about 35 MPa/km. These are numbers worth remembering, as they provide a good way to interpret pressures in the phase diagrams to come. Figure 8 shows the variation in pressure with depth using the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981).

## 7.2 Heat Transfer and the Temperature Gradient

Determining the **geothermal gradient**, the temperature variation with depth, is much more difficult than doing so for pressure, as there is no simple physical model analogous



**FIGURE 8** Pressure variation with depth. From Dziewonski and Anderson (1981).

to Equation (1). There are models, however, based on methods of heat transfer, that must be constrained to conform to measured heat flow at the surface. There are two primary sources of heat in the Earth.

- 1. Secular cooling** Primordial heat developed early in the history of the Earth from the processes of accretion and gravitational differentiation described in Section 4 has been gradually escaping since that time. This set up an initial temperature gradient once the planet solidified and began to cool. Some continued gravitational partitioning of iron in the inner core may contribute some heat as well. Estimates of the primordial contribution to the total surface heat flux is 10 to 25%.
- 2. Heat generated by the decay of radioactive isotopes** Most of the more radioactive elements are concentrated in the continental crust. Radioactive decay produces 75 to 90% of the heat that reaches the surface.

Once generated, heat is transferred from hotter to colder regions by any of four processes, depending on the nature of the material involved in the transfer:

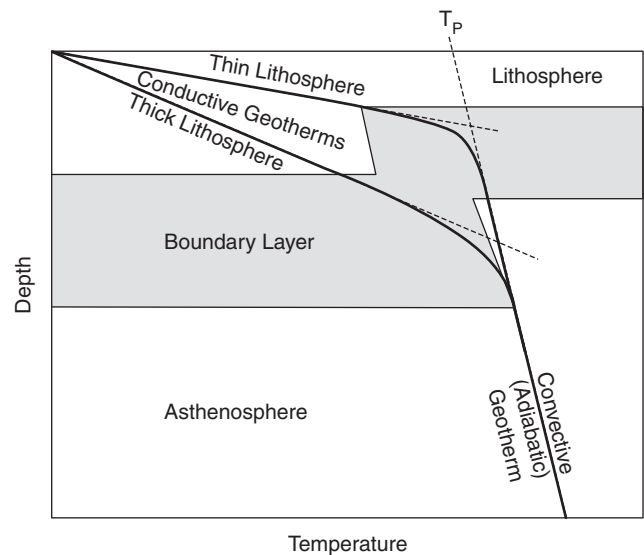
- 1.** If a material is sufficiently transparent or translucent, heat can be transferred by **radiation**. Radiation is the movement of particles/waves, such as light or the infrared part of the spectrum, through another medium. This is the principal way a lamp loses heat, or how the Earth loses heat from its surface into space. It is also the way we receive heat energy from the sun. Heat transfer by radiation is not possible within the solid Earth except possibly at great depth, where silicate minerals may become hot enough to lose some of their opacity to infrared radiation.
- 2.** If the material is opaque and rigid, heat must be transferred through **conduction**. This involves the transfer of kinetic energy (mostly vibrational) from hotter atoms to adjacent cooler ones. Heat conduction is fairly efficient for metals, in which electrons are free to migrate. This is why you can burn yourself if you handle an iron bar that has one end in a fire. Conduction is poor for silicate minerals but relatively efficient in the core.
- 3.** If the material is more ductile, and can be moved, heat may be transferred much more efficiently by **convection**. In the broadest sense, convection is the movement of material due to density differences caused by thermal or compositional variations. For the present purposes, we shall consider the type of convection that involves the expansion of a material as it heats, followed by the rise of that material due to its gain in buoyancy. This convection explains why it is hotter directly above a candle flame than beside it. The air is heated, it expands, and it rises because it is now lighter than the air around it. The same thing can happen to ductile rocks or liquids. Convection may involve flow in a single direction, in which case the moved hot material will accumulate at the top of the

ductile portion of the system (or a cooled density current will accumulate at the bottom). Convection may also occur in a cyclic motion, typically in a closed cell above a localized heat source. In such a *convection cell*, the heated material rises and moves laterally as it cools and is pushed aside by later convective matter. Material pushed to the margin cools, contracts, and sinks toward the heat source, where it becomes heated and the cycle continues.

4. **Advection** is similar to convection but involves the transfer of heat with rocks that are otherwise in motion. For example, if a hot region at depth is uplifted by either tectonism, induced flow, or erosion and isostatic rebound, heat rises physically (although passively) with the rocks.

Convection works well in the liquid core and in the somewhat fluid sub-lithospheric mantle. Convection is also a primary method of heat transfer in hydrothermal systems above magma bodies or within the upper oceanic crust, where water is free to circulate above hot rock material. Beyond these areas, however, conduction and advection are the only available methods of heat transfer. The transfer of heat is a very important concept in petrology, as it controls the processes of metamorphism, melting, and crystallization, as well as the motion, mixing, and mechanical properties of Earth materials. Several petrologic processes, from explosive volcanism to metamorphism to lava flows and pluton emplacement, are critically dependent upon maintaining a heat budget. We will consider heat transfer again later, in association with the cooling of a body of magma. The rate at which a body cools depends upon a number of variables, including the size, shape, and orientation of the body; the existence of a fluid in the surroundings to aid in convective heat transfer; and the type, initial temperature, and permeability of the country rocks.

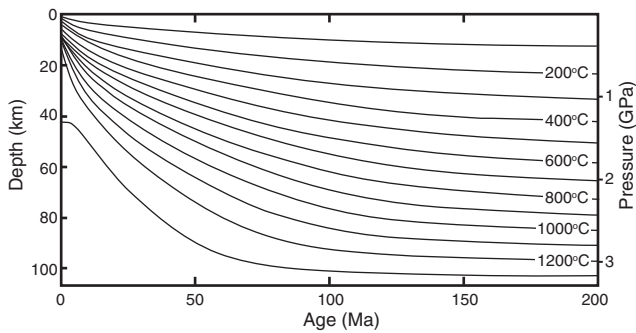
Figure 9 schematically illustrates heat flow in the upper few hundred kilometers of the Earth's interior. The passage of seismic waves through the mantle demonstrates that it reacts as a solid to abrupt stress changes. The mantle is therefore solid, but the sub-lithospheric mantle is capable of flow when subject to slower stress changes, such as those associated with heating. Heat is thus transported more effectively by convection at depth (steep gradient in Figure 9) but only by conduction across the stagnant rigid lithosphere (shallow gradients in Figure 9). Both the conductive and convective geothermal gradients are nearly constant (linear) but of different magnitudes. Deep convection is sufficiently rapid that no heat is lost by conduction to the surroundings, so heat is retained within the rising material. Processes in which heat is neither lost nor gained are called **adiabatic** processes. Between the convecting and conducting layers is a "boundary layer" a few tens of kilometers thick across which occurs the transition in rheology and heat transfer mechanism. Thinner lithosphere (such as in the ocean basins and illustrated on the right side of Figure 9) allows convective heat rise to shallower depths than does thicker



**FIGURE 9** Diagrammatic cross-section through the upper 200 to 300 km of the Earth, showing geothermal gradients reflecting more efficient adiabatic (constant heat content) convection of heat in the mobile asthenosphere (steeper gradient) and less efficient conductive heat transfer through the more rigid lithosphere (shallower gradients). The boundary layer is a zone across which the transition in rheology and heat transfer mechanism occurs. The thickness of the boundary layer is exaggerated here for clarity: it is probably less than half the thickness of the lithosphere. Notice that thinner lithosphere (on right) allows convective heat transfer to shallower depths, resulting in a higher geothermal gradient (greater change in temperature for a given pressure increment) across the boundary layer and lithosphere.  $T_p$  is the *potential temperature*: the temperature that the deep solid mantle would attain at atmospheric pressure as extrapolated adiabatically to the surface (McKenzie and Bickle, 1988).

lithosphere (continental, on the left in Figure 9). This results in a higher geothermal gradient (greater temperature change for a given increment in pressure) across the oceanic boundary layer and lithosphere, across which it must drop to ambient surface conditions. As we shall see, most melting occurs within the upper few hundred kilometers of the Earth's interior, so the temperature distribution at these levels is of great interest to petrologists.

Geothermal gradients are typically calculated using models based on measured heat flow at the surface or in drill holes or mines. Estimates of the steady-state heat flow from the mantle range from 25 to 38 mW/m<sup>2</sup> beneath the oceans to 21 to 34 mW/m<sup>2</sup> beneath the continents. Heat flow is commonly expressed in heat flow units (HFU), where 1 HFU = 41.84 mW/m<sup>2</sup>. Continental geothermal gradients are fairly well constrained by equilibration temperatures and pressures determined for mantle xenoliths from various depths (e.g., Rudnick and Nyblade, 1999). Because xenoliths are rarely found in the open ocean, oceanic geothermal gradients rely strictly on mathematical models. Oceanic geothermal gradients are further complicated by divergent plate boundaries, which result in mantle upwelling and partial melting, which in turn creates oceanic crust and lithospheric plates at ocean ridges. The geothermal gradient in the ocean basins thus depends on the age of the lithosphere (and hence



**FIGURE 10** Temperature contours calculated for an oceanic plate generated at a mid-ocean ridge (age 0) and thickening as it cools. The 1300°C isotherm is a reasonable approximation for the base of the oceanic lithosphere. The plate thus thickens rapidly from zero to 50 Ma and is essentially constant beyond 100 Ma. From McKenzie et al. (2005).

distance from the ridge and spreading rate). Figure 10 shows the temperature contours calculated by McKenzie et al. (2005) for depth versus age of oceanic lithosphere. The transition from rigid (lithospheric) to ductile (asthenospheric) behavior is in the 1300 to 1400°C range (1375°C, according to McKenzie), so the 1300°C isotherm can be used as an approximation for the base of the oceanic lithospheric plate. The plate thus thickens from about 50 km at the ridge to ~110 km at “maturity” of ~100 Ma, beyond which it remains essentially constant in thickness and thermal structure. In the continental lithosphere heat flow is high in rifts and orogenic belts but settles down to a steady state after about 800 Ma in the platforms and shields (Sclater et al., 1980).

Several estimates of geothermal gradients for continental shields and mature oceanic intraplate settings are shown in Figure 11. Because the thermal conductivity of shallow crustal rocks is very low, heat transfer is slow, and the shallow geothermal gradient is correspondingly high. Simply extrapolating the shallow gradient on the basis of heat flow measurements leads to an impossibly high geothermal gradient (as shown). As expected, the geothermal gradient is higher in the ocean setting than in the continental cratons because the oceanic lithosphere is thinner. Shallow gradients in old oceanic lithosphere in Figure 11 range from about 17 to 20°C/km, gradually steepening to about 7°C/km with depth. Shallow shield gradients range from about 12 to 18°C/km, transitioning to about 4 to 6°C/km below about 30 km. The oceanic versus shield temperature differences are essentially restricted to the lithosphere and boundary layer, so the oceanic and continental curves converge by about 250 to 300 km where convection has a homogenizing influence (see Figure 9). Nearly adiabatic (constant heat content) convective heat flow deeper than about 300 km results in a linear geothermal gradient of approximately 0.3°C/km (~10°C/GPa), as shown in Figure 12. The gradient shallows across the D” layer and then becomes steep in the metallic core, where the thermal conductivity and convection (in the liquid portion at least) is very high. The density contrast across the core–mantle boundary prohibits core material from rising into the mantle (convection), so heat can only be

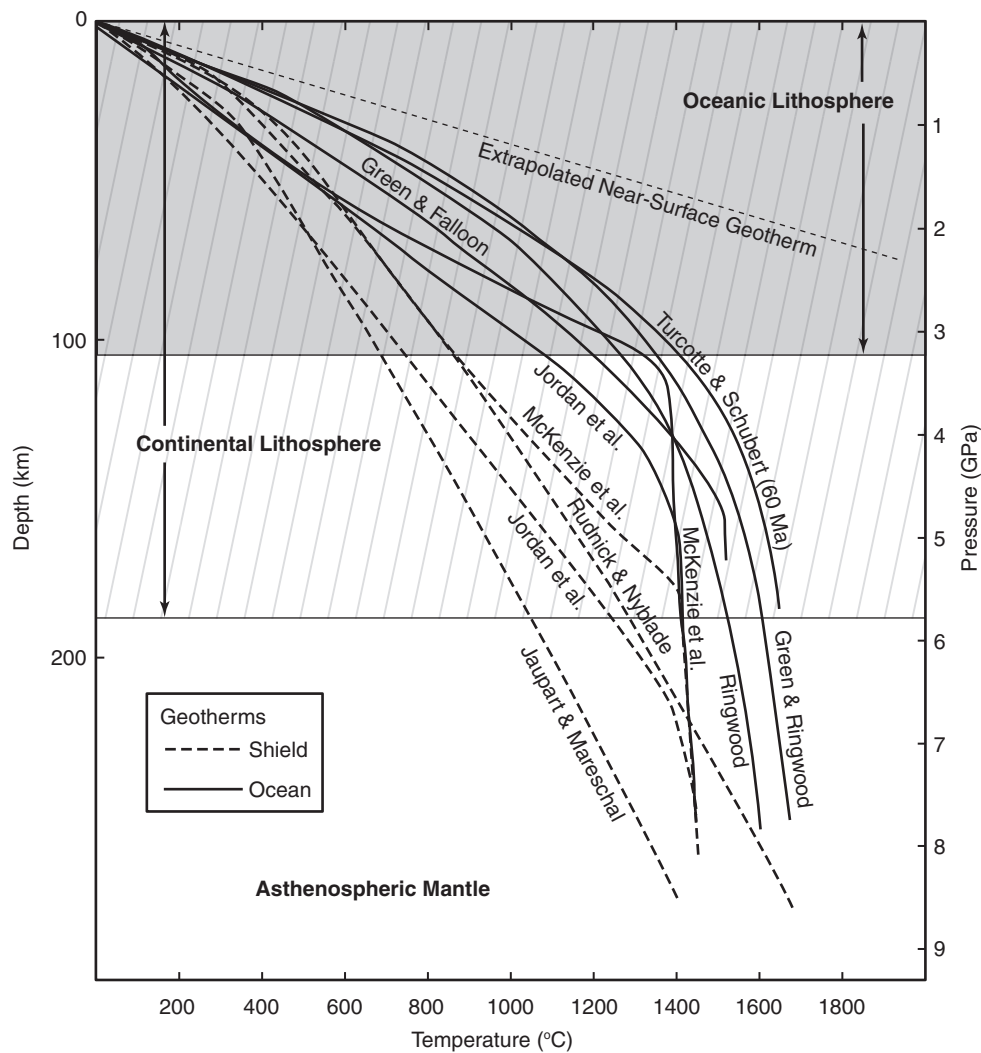
transferred upward (across the boundary) by conduction, resulting in a thermal boundary layer across which the thermal gradient is estimated to be 300 to 1000°C.

### 7.3 Dynamic Cooling of the Earth: Geodynamics and Plate Tectonics

Figure 11 is intended to illustrate the thermal structure at shallow levels of the Earth’s interior in stabilized areas, similar to a static, chemically and physically stratified Earth with continents and oceans. The Earth is far more dynamic, however, behaving like a viscous solid in a gravitational field, heated from within and below, cooled from above, and expanding when heated. Convection in such a fluid occurs when the thermal **Rayleigh number** (a dimensionless ratio representing the potential vigor of convection) exceeds 1000. With a Rayleigh number about 10,000 times this, the lower mantle not only convects but probably does so vigorously. Laboratory and model studies demonstrate that the temperature variations that drive convective flow are concentrated in thin boundary layers that are much smaller than the overall circulation pattern. Given the temperature-dependent viscosity of mantle material, the upper boundary of this system is a cool thermal boundary layer with low viscosity: the lithosphere. Because of its high density, the lithosphere is gravitationally unstable (negatively buoyant). In laboratory models of convecting fluids, cool dense upper thermal boundary layers descend as either cylindrical downwellings or as larger networks of partially connected tabular-shaped downwelling slabs, depending on the rheology of the material and the ratio of internal to basal heating. The mantle is perhaps 80 to 90% internally heated, largely by radioactive decay, with roughly 10% of its heat coming from the base/core. This ratio and the stiff rheology of the lithosphere clearly lead to slab-like downwelling, resulting in plate tectonics. Early debates as to whether mantle convection drives plate tectonics or plate instabilities lead to mantle convection now appear misguided. Mantle convection and plate tectonics are inseparable manifestations of the heat-driven dynamic cooling process of the Earth, given its present thermal state and rheological properties. For the Earth, plate tectonics *is* mantle convection.

Several forces are at work, however, and their relative contribution has been debated since plate tectonics was first recognized. The negative buoyancy of the descending plates has been called **slab pull**, with the regrettable implication that the dense subducting slab pulls the rest of the plate after it. The motion of the lithospheric plates down and away from the elevated mid-ocean ridges has also received an unfortunate name: **ridge push**. Both, however, are actually *body* forces, affecting the entire plate (not just a pulled or pushed end). Slab pull may be considered a horizontal pressure gradient associated with the sinking of a slab. As it descends in a subduction zone, the slab can’t pull significantly on the rest of the plate, which would simply fault and break. Instead, as it sinks, the slab sets up circulation patterns in the mantle that exert a sort of suction force. As the slab descends from the surface, the pressure behind it is lowered, which is immediately compensated by feeding more plate



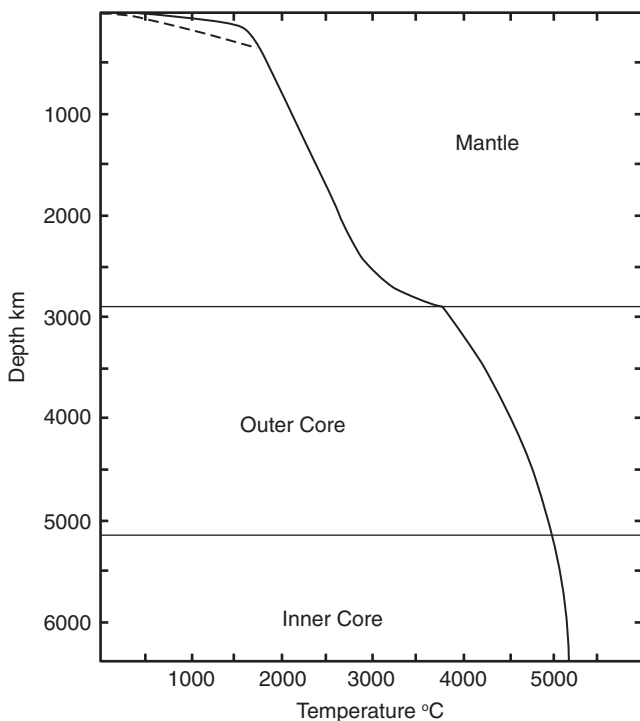


**FIGURE 11** Estimates of oceanic (solid curves) and continental shield (dashed curves) geothermal gradients to a depth of 300 km. The thickness of mature (> 100Ma) oceanic lithosphere is shaded, and that of continental shield lithosphere is hatched. Data from Green and Falloon (1998), Green and Ringwood (1963), Jaupart and Mareschal (1999), McKenzie et al. (2005 and personal communication), Ringwood (1966), Rudnick and Nyblade (1999), and Turcotte and Schubert (2002).

into the subduction zone. Ridge push is the gravitational sliding of a plate off the elevated ridge, not an active push by ascending mantle flow, which would buckle the plate. Forsyth and Uyeda (1975) found that plate motion velocities are roughly proportional to the percentage of a plate's perimeter that corresponds to subduction zones and not to ridges, strongly implying that the negative buoyancy associated with slab pull is the principal force. Lithgow-Bertelloni and Richards (1998) estimate that ridge push constitutes only 5 to 10% of the driving force of slab subduction. The predominance of slab pull and the lack of any significant gravity anomaly at mid-ocean ridges suggest that mantle upwelling at divergent plate boundaries, which moves heat upward, is essentially a passive response to plate separation and descent. Upward heat transfer is thus essentially *advective* in this situation (carried upward in material that is rising for reasons other than thermal expansion). From this brief description, we can conclude that the scale of the plates effectively con-

trols the scale of mantle convection (i.e., the location of major upwelling and downwelling circulation), and not the reverse.

But what of the other boundary layer at the base of the mantle? The density more than doubles from the lower mantle to the core, thus preventing convection across the boundary, regardless of thermal differences. The liquid outer core must be internally convecting in order to create the Earth's magnetic field, and this convection delivers heat to the base of the mantle. The D'' layer may be the resulting mantle-side thermal boundary layer, across which the temperature gradient is estimated to be 300 to 1000°C. Convection within the mantle results from thermal instabilities as the heated D'' layer becomes positively buoyant. The viscosity is much lower than that of the plates at these elevated temperatures, and the instabilities take the alternative form of rising cylindrical plumes rather than slabs. These **plumes** are more of an



**FIGURE 12** Estimate of the geothermal gradient to the center of the Earth (after Stacey, 1992). The shallow solid portion is very close to the Green and Ringwood (1963) oceanic geotherm in Figure 11, and the dashed geotherm is the Jaupart and Mareschal (1999) continental geotherm.

active upwelling, in comparison to the more passive shallow mantle rise in response to plate separation. The plumes appear to be largely independent of plate tectonics, but in many cases are sufficiently vigorous to penetrate the lithosphere and reach the Earth's surface, where they result in **hotspot** volcanism and large elevated lithospheric swells. Calculations of the excess heat required to create such swells have yielded estimates of the total heat transported by plumes, which agree with estimates of the heat flux from the core to the mantle (approximately 10% of the total heat flux to the surface). Plate tectonics thus appears to be the method by which a largely internally heated and viscous mantle cools by convection, whereas plumes essentially cool the core and have an increasing effect on the lower mantle.

In addition to plate tectonics and plumes, Earth rotation, melt migration, frictional drag, compositional heterogeneity, phase transformations, "torroidal" (strike-slip) plate motion, and the insulating effects of continents, among other factors, further complicate the heat flow and dynamics of the mantle. All of these factors can modify the simple static Earth model depicted in Figures 9 and 11. Figure 13 illustrates the variation in surface heat flux based on a smoothing of data from surface measurements from more than 20,000 measuring stations. The darker patterns clearly indicate the role of mid-ocean ridges as heat "leaks" with high geothermal gradients. Pollack et al. (1993) also note that fully half of the Earth's heat loss is associated with the cooling of relatively young Cenozoic oceanic lithosphere. Cratons are cooler, but not uniformly so. The mathematical

smoothing process used hides small-scale features such as hotspot plumes. A technique called *seismic tomography* uses powerful computers to model three-dimensional seismic wave velocity distributions in the Earth. Velocity anomalies are usually associated with temperature variations, so the technique has great promise for yielding a clearer picture of detailed mantle dynamics. For a general discussion of seismic tomography, see Lowrie (1997). Schubert et al. (2001) also provide a comprehensive review of thermal and seismic models of the Earth. Comparison of mantle temperatures from one locality to another is complicated by the natural variation in temperature with depth. The problem can be alleviated by using a pressure reference frame. McKenzie and Bickle (1988) suggested a 1-atmosphere reference they called the *potential temperature* ( $T_p$ ): the temperature that the solid mantle would attain if it could reach the surface adiabatically without melting (see Figure 9).

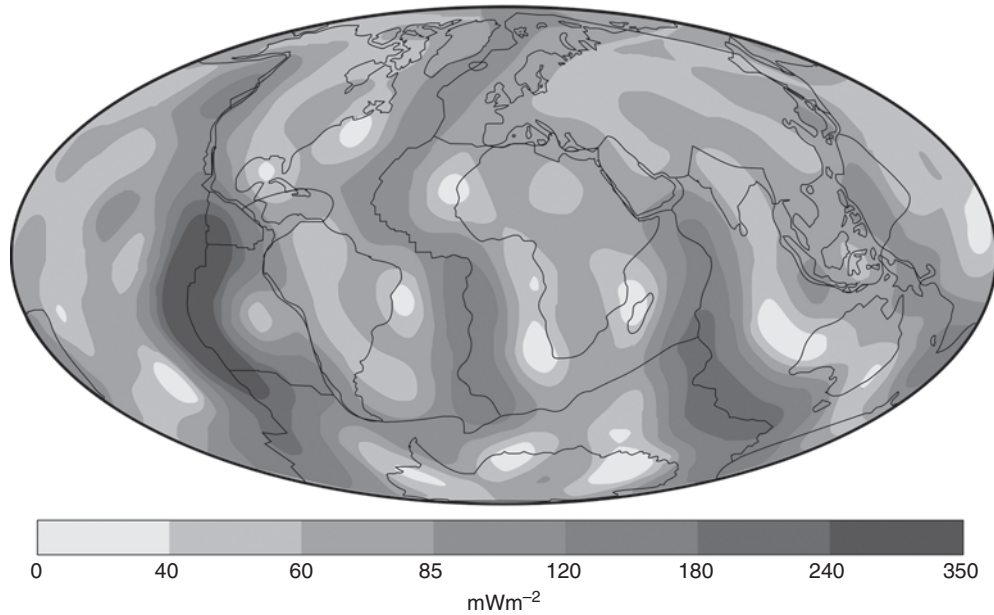
Even a general model of mantle convection, plumes, etc. depends on the influence of the 660-km phase transition on mantle dynamics. Some investigators consider the density contrast across the boundary insufficient to impede convection, so that mantle convection cells span the full vertical extent of the mantle. Such models are called *whole-mantle* models. Others believe the 660-km transition can impede convection, so that warm, buoyant mantle material in rising portions of deep convection cells would be less buoyant than the material above the transition, causing the rising material to spread laterally instead. If so, then heat can transfer across the boundary only by conduction, perhaps inducing convection in the upper mantle. Flow in the upper layer can also be induced by plate motion, rising at divergent boundaries and sinking with subduction. Models with a 660-km barrier are called *two-layer* mantle models. Figure 14 combines aspects of both model types. In it, the 660-km boundary impedes most lower mantle convection from rising further, but subducted oceanic lithosphere (although impeded somewhat) is dense enough to sink through the boundary layer and accumulate as the D" layer. According to this model, when heated at the core-mantle boundary, this material regains positive buoyancy, resulting in ascending plumes that rise vigorously enough to penetrate the boundary again and reach the surface (e.g., Hawaii). The 660-km boundary layer in such models allows some material transfer but impedes wholesale mantle convection from homogenizing the full vertical extent of the mantle and thus permits the composition of the shallow and deep mantle to evolve independently. Mantle flow dynamics and heat transfer play critical roles in magma genesis and evolution, so we will be exploring many of these processes (and the mantle layering controversy) more fully in subsequent chapters.

## 8 MAGMA GENERATION IN THE EARTH

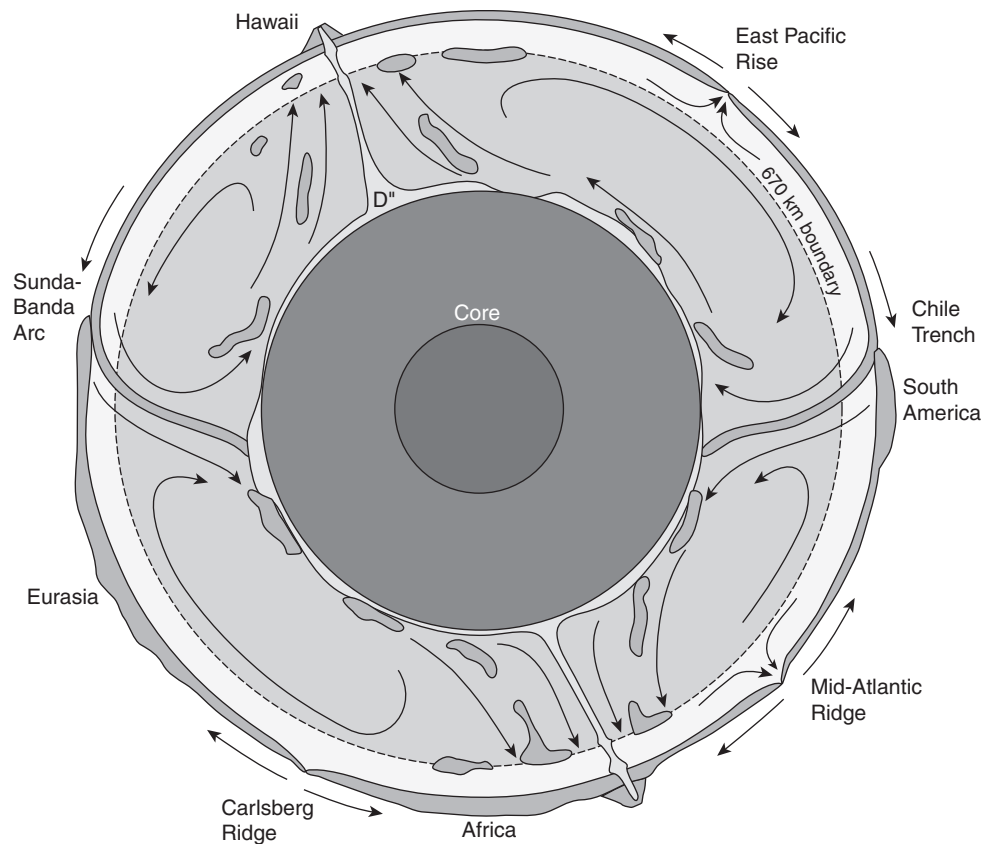
We next address the problem of magma generation. **Petrogenesis** is a good general term in igneous petrology for the generation of magma and the various methods of diversification of such magmas to produce igneous rocks. Most magmas originate by melting in the Earth's mantle, but some show evidence of at least a partial crustal component.

Some Fundamental Concepts

Heat Flow



**FIGURE 13** Pattern of global heat flux variations compiled from observations at more than 20,000 sites and modeled on a spherical harmonic expansion to degree 12. From Pollack et al. (1993). © AGU with permission.



**FIGURE 14** Schematic diagram of a two-layer dynamic mantle model in which the 660-km transition is a sufficient density barrier to separate lower mantle convection (arrows represent flow patterns) from upper mantle flow, largely a response to plate separation. The only significant things that can penetrate this barrier are vigorous rising hotspot plumes and subducted lithosphere (which sink to become incorporated in the D" layer, where they may be heated by the core and return as plumes). After Silver et al. (1988).

Plate tectonics plays a major role in the generation of several magma types, but other types seem to result from processes at depths in the mantle greater than are influenced directly by plate tectonics. Figure 15 is a very generalized summary of the principal types of magmas and their geologic setting.

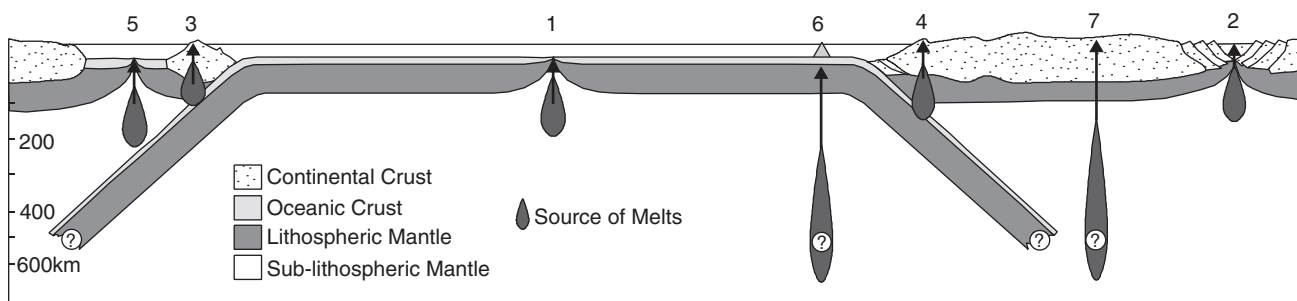
The most voluminous igneous activity occurs at divergent plate boundaries. Of these, mid-ocean ridges (location 1 in Figure 15) are the most common. As introduced above, the shallow mantle beneath the ridge undergoes partial melting, and the resulting basaltic magma rises and crystallizes to produce the oceanic crust. If a divergent boundary is initiated beneath a continent (location 2), a similar process takes place. The resulting magmatism, particularly at the early stages of continental rifting, is commonly alkaline and typically shows evidence of contamination by the thick continental crust. If the rift continues to develop, oceanic crust will eventually be created in the gap that forms between the separating continental fragments. The result will be a new ocean basin and igneous activity similar to that in location 1.

An oceanic plate created at mid-ocean ridges moves laterally and eventually is subducted beneath a continental or another oceanic plate. Melting also takes place at these subduction zones. The number of possible sources of magma in subduction zones is far more numerous than at ridges and may include various components of mantle, subducted crust, or subducted sediments. The types of magma produced are correspondingly more variable than for divergent boundaries, but andesites are the most common. If oceanic crust is subducted beneath oceanic crust (location 3), a volcanic island arc forms. If oceanic crust is subducted beneath a continental edge (location 4), a continental arc forms along the “active” continental margin. A continental arc is generally more silica rich than is an oceanic arc. Plutons are also more common in continental arcs, either because the melts rise to the surface less efficiently through the lighter continental crust or because uplift and erosion is greater in the continents and exposes deeper material.

A different, and slower, type of plate divergence typically takes place behind the volcanic arc associated with subduction (location 5). Most geologists believe some sort of “back-arc” extension is a natural consequence of subduction, probably created by frictional drag associated with the

subducting plate. Such drag pulls down part of the overlying mantle, requiring replenishment from behind and below. Back-arc magmatism is similar to mid-ocean ridge volcanism. Indeed, a ridge also forms here, and oceanic crust is created and spreads laterally from it. Back-arc spreading, however, is slower, volcanism is more irregular and less voluminous, and the crust created is commonly thinner than in the oceans. At times, rifting occurs behind a continental arc, and the volcanic portion separates from the continent as a marginal sea forms through back-arc spreading. Such a process is believed to have separated Japan from the Asian mainland. At other times, such a process seems to initiate, and then it mysteriously ceases. The result may just be a graben structure, or plateau-type basalts may form prior to cessation of activity.

Although magmatism is certainly concentrated at plate boundaries, some igneous activity also occurs within the plates, both oceanic (location 6) and continental (location 7). Ocean islands such as Hawaii, the Galapagos, and the Azores all form via volcanism within the oceanic plates. The products are usually basaltic but are commonly more alkaline than ridge basalts. The reason for this type of igneous activity is much less obvious than it is for plate margins because our plate tectonic paradigm is of little use in these mid-plate regimes. The source of the melts is also less clear but appears to be deep, certainly well into the asthenosphere. Several of these occurrences exhibit a pattern of igneous activity that gets progressively younger in one direction. The direction correlates well with plate motion in a manner which suggests that the plate is moving over a stationary “hotspot,” or mantle, “plume” (as described above) with the most recent activity occurring directly over the plume. Intraplate activity within continental plates is much more variable than that within the oceans. It is compositionally variable but usually alkaline—and occasionally extremely so. This reflects the more complex and heterogeneous continental crust and subcontinental mantle as well. Some of the most unusual igneous rocks, such as kimberlites and carbonitites, occur within continental provinces. The term **igneous–tectonic association** refers to these broad types of igneous occurrence, such as mid-ocean ridge, island arc, and intra-continental alkalic systems.



**FIGURE 15** Generalized cross section illustrating magma generation associated with various plate tectonic settings.

---

## Summary

Petrology is the study of the nature and origin of rocks. Igneous petrology addresses the processes that produce melts (magma), how those melts then rise, the chemical and mineralogical evolution as they cool and crystallize, and the eventual eruption or emplacement of the resulting rock bodies. Field, chemical, and textural criteria (the latter best observed using the polarized-light microscope) are the principal ones used to evaluate these processes. The Earth comprises the ~10 km thick Mg-Fe-rich (mafic) oceanic crust and the ~40–60 km thick Si-Al-rich (sialic) continental crust, both underlain by ultramafic mantle (~3000 km thick), as well as the central metallic Fe-rich core. From a rheological perspective, the Earth comprises the rigid lithospheric plates, underlain by the slightly molten ductile asthenosphere, the more rigid mesosphere, and the outer liquid and inner solid core. The Earth originated by accretion of matter in the solar nebula to form small planetesimals, which further collapsed under the influence of gravity to form the Earth itself. Concentration of heavy elements (principally Fe and lesser Ni) toward the center of the Earth released sufficient potential energy to melt at least most of the Earth and differentiate it into the metallic core and silicate mantle (and perhaps a thin ocean and atmosphere). The crust evolved later and more gradually by plate tectonic processes: oceanic crust at divergent plate boundaries and continental crust as small island-arc terranes at convergent (subduction) boundaries, which accreted by collisions to form larger continental masses.

If we are to evaluate melting and crystallization in the Earth, we must understand the pressure and temperature gradients. Because melts that reach the surface are generated in the crust and upper mantle, we are most interested in the gradients across the upper few hundred kilometers. Pressure increases with depth due to the weight of the progressively greater overlying material. A reasonable shallow pressure gradient is ~35 km/GPa. Because the Earth is still cooling following its early accretion and due to radioactive heat, it gets hotter with increasing depth. The increase in temperature with depth is called the geothermal gradient. Heat is transferred upward by conduction across the rigid lithosphere and more effectively by convection below that. Because the oceanic lithosphere is thinner, convection transfers heat to shallower levels, and the geothermal gradient is thus higher in the upper 200 to 300 km in oceanic areas than in continental shields.

Magma generation is largely controlled by plate tectonic processes. Magmas are generated in six principal settings:

- Mid-ocean ridges (oceanic divergent plate boundaries)
- Continental rifts (incipient continental plate boundaries)
- Island arcs (intra-oceanic subduction zones)
- Continental arcs (ocean-continent subduction zones)
- Back-arc basins (divergence behind the leading edge of the overriding plate at subduction zones)
- Hotspots (rising plumes that can penetrate either oceanic or continental lithospheric plates)

---

## Key Terms

Magma	Lithosphere	Lithostatic pressure
Lava	Asthenosphere	Conduction of heat
Volcanic	Low-velocity layer	Convection of heat
Plutonic	Lithophile	Radiation of heat
Oceanic crust	Chalcophile	Advection of heat
Continental crust	Siderophile	Geothermal gradient
Shield	Iron meteorites	Slab-pull/ridge-push
Platform	Stone meteorites	Petrogenesis
Mantle	Stony-iron meteorites	Igneous-tectonic association
Core	Chondrite meteorites	

---

## Review Questions and Problems

Review Questions and Problems are located on the author's web page at the following address: <http://www.prenhall.com/winter>

---

## Important "First Principle" Concepts

- Igneous rocks crystallize from a melt. They can be recognized either by field setting or by textural criteria.
- The oceanic crust, continental crust, mantle, and core are the four principle *compositional* subdivisions of the Earth.
- The lithosphere, asthenosphere, mesosphere, outer core, and inner core are the five principal *mechanical* subdivisions of the Earth.

- The lithosphere contains both the crust and some upper mantle, so the moho (which separates the crust from the mantle) lies *within* the lithosphere. Lithospheric plates are about 110 km thick, and their base is called the low-velocity layer, where the mantle becomes more ductile (probably due to small proportions of partial melt).
- The Earth probably accreted from a gravitationally collapsing cloud of planetesimals and quickly separated into an Fe-rich metallic core and a silicate mantle. The oceanic and continental crust formed over long time periods (and continue to do so today).
- Basaltic oceanic crust is generated at mid-ocean ridges and is consumed at subduction zones. Therefore, the crust in the ocean basins is relatively young (< 160 Ma) and recycles.
- Heterogeneous continental crust is more silicic, thicker, and more buoyant than oceanic crust. It is thus not recycled at subduction zones but continues to increase, typically as a result of those very subduction zone processes.
- The composition of chondrite meteorites is considered a close approximation of the primordial Earth (before it differentiated).
- The most common elements in the Earth are, in decreasing order of abundance, Fe, O, Si, Mg, S, Ca, and Al.
- A good average pressure gradient in the crust is about 30 MPa/km (yielding 1 GPa at the base of typical 35-km thick continental shield crust). In the denser upper mantle the pressure gradient is about 35 MPa/km.
- Plate tectonics is the process that convectively cools the Earth's mantle. Convection involving descending slabs (as compared to cylindrical plumes) is the natural method, given the physical properties of the mantle at present (positive thermal expansion, temperature-dependent viscosity, etc.). Plume-style convection is more appropriate for the deeper and hotter boundary layer at the core–mantle boundary, however, and such rising plumes probably cool the core and rise through the mantle, largely independently of plate tectonics.
- Plate tectonics plays a pivotal role in most igneous processes, particularly those processes at divergent and consuming plate margins.

---

## Suggested Further Readings

### History of Igneous Petrology

Young, D. A. (2003). *Mind Over Magma. The Story of Igneous Petrology*. Princeton University Press. Princeton, NJ.

### The Interior of the Earth and Global Dynamics

Anderson, D. L. (1992). The Earth's interior. In: *Understanding the Earth* (eds. G. C. Brown, C. J. Hawkesworth, and P. C. L. Wilson). Cambridge University Press. Cambridge, UK.

Brown, G. C., and A. E. Mussett. (1993). *The Inaccessible Earth*. Chapman & Hall. London.

Davies, G. F. (2001). *Dynamic Earth: Plates, Plumes, and Mantle Convection*. Cambridge University Press. Cambridge, UK.

Gurnis, M., M. E. Wysession, E. Knittle, and B. A. Buffett (eds.). (1998). *The Core–Mantle Boundary Region*. Geodynamics Series **28**. American Geophysical Union. Washington, DC.

Jackson, I. (1998). *The Earth's Mantle. Composition, Structure, and Evolution*. Cambridge University Press. Cambridge, UK.

Jacobs, J. A. (1975). *The Earth's Core*. Academic Press. London.

Jones, J. H., and M. J. Drake. (1986). Geochemical constraints on core formation in the Earth. *Nature*, **322**, 221–228.

Karato, S.-I., A. M. Forte, R. C. Liebermann, G. Masters, and L. Stixrude. (eds.). (2000). *Earth's Deep Interior: Mineral Physics and Tomography from the Atomic to the Global Scale*. Geophysical Monograph **117**. American Geophysical Union. Washington, DC.

Montagner, J.-P., and D. L. Anderson. (1989). Constrained reference mantle model. *Phys. Earth Planet. Sci. Lett.*, **44**, 205–207.

Palme, H., and H. St. C. O'Neill. (2003). Cosmochemical estimates of mantle composition. In: *The Mantle and Core* (ed. R. W. Carlson), Vol. 2 *Treatise on Geochemistry* (eds. H. D. Holland and K. K. Turekian). Elsevier-Pergamon. Oxford, UK. pp. 1–38.

Richards, M. A., R. G. Gordon, and R. D. van der Hilst. (eds.). (2000). *The History and Dynamics of Global Plate Motions*. Geophysical Monograph **121**. American Geophysical Union. Washington, DC.

Schubert, G., D. L. Turcotte, and P. Olson. (2001). *Mantle Convection in the Earth and Planets*. Cambridge University Press. Cambridge, UK.

Silver, P. G., R. W. Carlson, and P. Olson. (1988). Deep slabs, geochemical heterogeneity, and the large-scale structure of mantle convection: Investigation of an enduring paradox. *Ann. Rev. Earth Planet. Sci.*, **16**, 477–541.

Turcotte, D. L., and G. Schubert. (2002). *Geodynamics*. Cambridge University Press. Cambridge, UK.

Wyllie, P. J. (1971). *The Dynamic Earth: Textbook in Geosciences*. John Wiley and Sons. New York.

### The Origin of the Earth and Solar System

Brown, G. C., and A. E. Mussett. (1993). *The Inaccessible Earth*. Chapman & Hall. London.

Cameron, A. G. W. (2001). From interstellar gas to the Earth–Moon system. *Meteorit. Planet. Sci.*, **36**, 9–22.

Ernst, W. G. (2007). Speculations on evolution of the terrestrial lithosphere–asthenosphere system—Plumes and plates. *Gondwana Research*, **11**, 38–49.

Lauretta, D., L. Leshin, and H. Y. McSween, Jr. (eds.). (2006). *Meteorites in the Early Solar System II*. University of Arizona Press. Tucson, AZ.

Newsom, H. W., and J. H. Jones. (1990). *Origin of the Earth*. Oxford University Press. New York.

Taylor, S. R. (1992). *Solar System Evolution: A New Perspective*. Cambridge University Press. Cambridge, UK.

Taylor, S. R. (1992). The origin of the Earth. In: *Understanding the Earth* (eds. G. C. Brown, C. J. Hawkesworth, and M. S. Matthews). Cambridge University Press. Cambridge, UK.

Walther, M. J., and R. G. Trønnes. (2004). Early Earth differentiation. *Earth Planet. Sci. Lett.*, **225**, 253–269.

Weatherill, G. W. (1989). The formation of the solar system: Consensus, alternatives, and missing factors. In: *The Formation and Evolution of Planetary Systems* (eds. H. A. Weaver and L. Danly). Cambridge University Press. Cambridge, UK.

Weaver, H. A., and L. Danly (eds.). (1989). *The Formation and Evolution of Planetary Systems*. Cambridge University Press. Cambridge, UK.

Wood, J. A. (1979). *The Solar System*. Prentice Hall. Englewood Cliffs, NJ.

### **Meteorites**

- Goldstein, J. I., and J. M. Short. (1967). The iron meteorites, their thermal history and parent bodies. *Geochim. Cosmochim. Acta*, **31**, 1733–1770.
- Kerrige, J. F., and M. S. Matthews (eds.). (1988). *Meteorites and the Early Solar System*. University of Arizona Press. Tucson, AZ.
- MacSween, H. Y. (1987). *Meteorites and Their Parent Planets*. Cambridge University Press. Cambridge, UK.
- Wasson, J. T. (1985). *Meteorites*. W.H. Freeman. New York.

### **Pressure and Temperature Gradients in the Earth**

- McKenzie, D., J. Jackson, and K. Priestly. (2005). Thermal structure of oceanic and continental lithosphere. *Earth Planet. Sci. Lett.*, **233**, 337–349.
- Sclater, J. G., C. Jaupart, and D. Galson. (1980). The heat flow through oceanic and continental crust and the heat loss of the Earth. *Rev. Geophys. Space Sci.*, **18**, 269–311.
- Williamson, E. D., and L. H. Adams. (1923). Density distribution in the Earth. *J. Wash. Acad. Sci.*, **13**, 413–428.