3 Geology and Heat Architecture of the Earth’s Interior

KEY CHAPTER OBJECTIVES

- Distinguish between the Earth’s compositional and physical mechanical layers.
- Identify the sources of Earth’s internal heat.
- Compare and contrast conductive and convective heat flow.
- Recognize conductive and convective zones of heat transfer from drill-hole temperature profiles.
- Explain the significance of heat flow maps and temperature-at-depth maps.

To more completely understand geothermal resources and their distribution on the planet, a review of the Earth’s compositional and physical make-up is necessary. The Earth is compositionally inhomogeneous, consisting of an iron–nickel core, a dense rocky mantle, and a thin, comparatively low-density rocky crust. This compositional diversification developed shortly after our planet formed when more dense material sank to the center and low-density material rose toward the surface. Furthermore, because of this compositional diversity, differences in physical or mechanical properties exist (liquid or molten vs. solid; brittle vs. ductile deformation). Brittle behavior means breaking or fracturing after a threshold level of stress is applied, such as what happens when a glass vase is dropped on a hard surface. Ductile deformation, on the other hand, reflects bending without breaking after a material’s yield strength is exceeded, such as bending a metal wire or molding clay. Understanding both the compositional and physical characteristics of the Earth’s interior lays the groundwork for the discussion about plate tectonics in Chapter 4; plate tectonics exerts a fundamental control on the distribution of Earth’s mineral, fossil fuel, and geothermal resources.

EARTH’S COMPOSITIONAL AND RHEOLOGICAL LAYERS

The Earth’s radius is just under 6400 km. Extending outward from Earth’s center, systematic changes occur in both composition and rheological behavior (physical or mechanical properties of a material, such as changes from solid to liquid or brittle to ductile). We will begin with compositional changes.
Earth’s Compositional Layers

The area extending from the center of the Earth to a depth of about 2900 km is known as the Earth’s core. The core consists of both solid and molten iron and nickel, and its temperature is comparable to the surface of the sun, or about 6000°C. Overlying the core is the mantle, which extends from a depth of about 2900 km to less than 100 km. Volumetrically, the mantle makes up the largest part of Earth’s interior. The mantle consists of dense iron- and magnesium-rich rock, whose temperature decreases progressively upward from about 5000°C to less than 1500°C.

The third and last layer is the Earth’s crust, which consists of a thin shell, varying from 70 to 80 km thick under parts of continents to less than a few kilometers thick under parts of the ocean floor. A useful analogy of the compositional layers of the Earth is a peach. The size of the pit would be proportional to the Earth’s core, the pulp (the edible part) proportionally represents the mantle, and the fuzzy skin would have the proportional thickness of the crust. Earth’s compositional layers are illustrated in Figure 3.1.

Unlike the more compositionally homogeneous core and mantle, the crust consists of two types: oceanic and continental (Figure 3.1). Oceanic crust underlies the ocean basins and consists of a dark-colored, moderately dense rock called basalt. It is relatively thin, reaching a maximum of 7 km and a minimum of less than a

![Figure 3.1](http://www.visionlearning.com/img/library/large_images/image_4859.gif)
kilometer below mid-ocean ridges. Continental crust is comprised mainly of lower
density, lighter colored igneous and metamorphic rocks, such as granite and gneiss
(discussed in more detail in Chapter 4). These igneous and metamorphic rocks of
continental crusts are capped in places by a veneer of sedimentary rocks, including
sandstone and limestone. Because continental crust is less dense than oceanic crust,
it sits higher compared to oceanic crust, explaining why continents for the most part
lie above sea level.

**Earth’s Rheological (Physical) Layers**

In response to changes in pressure and temperature, a material’s physical nature
(known as rheology) can change, such as from solid to liquid with rising temperature
or the reverse with falling temperature or rising pressure. The composition of the
material, however, remains essentially unchanged despite changes in the physical
state. Another change in rheology would be the change from brittle breaking, form-
ing fractures under low temperature and pressure, to ductile bending under high tem-
perature and pressure prior to actual melting. In other words, a ductile substance is a
solid that has the ability to flow, and within the Earth ductile materials flow at rates
of a few centimeters per year in response to pressure differences and convection.
The Earth is comprised of five main rheological layers, moving from the surface
downward: the lithosphere, asthenosphere, mesosphere, outer core, and inner core.
The relationship between Earth’s compositional and physical or mechanical layers is
illustrated in **Figure 3.2**.

![Cross-sectional view of Earth's compositional and mechanical layers.](http://www.visionlearning.com/img/library/large_images/image_4859.gif)

**FIGURE 3.2** Cross-sectional view of Earth’s compositional and mechanical layers. For
details, see text. (Adapted from Visionlearning®, http://www.visionlearning.com/img/library/
large_images/image_4859.gif.)
Lithosphere
The lithosphere represents the strong, relatively brittle outermost layer and averages about 100 km thick. It is compositionally diverse as it embraces both the crust and uppermost mantle because both compositional layers behave similarly from a rheological standpoint—relatively strong and brittle. The lithosphere will be discussed more in Chapter 4 because it makes up the Earth’s tectonic plates, great chunks of rock that are continually moving with respect to each other.

Asthenosphere
Underlying the lithosphere, between 100 km and about 300 km, is a weak zone of rock called the asthenosphere, which is part of the upper mantle. The rock in the asthenosphere is weak because it is close to its melting point but still mainly a solid (Figure 3.3). However, because of the high heat, the rock is mechanically weakened and has the ability to flow (ductile behavior) in response to thermal and pressure gradients. Motion in the asthenosphere contributes to movement of the overlying lithosphere or tectonic plates.

Mesosphere
Below the asthenosphere, the behavior of the rest of the mantle, referred to as the mesosphere, is mechanically similar. The mesosphere consists of the lower and middle parts of the mantle. Because of the added pressure with depth, the rocks

![FIGURE 3.3](https://example.com/fig3.3.png)

**FIGURE 3.3** Depth and temperature plot showing the geothermal gradient (red line) and melting curve with depth of rock (blue line). The melting point of rocks increases with depth because increasing pressure favors the denser, solid phase. Note that the rocks are close to their melting point in the asthenosphere and therefore mechanically weak. As the geotherm and melting point curves diverge below the asthenosphere, the rocks become less weak. The 1300°C marks the approximate temperature at which basaltic rocks begin to melt near the Earth’s surface.
are not as close to where they would begin melting as in the asthenosphere and are therefore stronger and less ductile (Figure 3.3). Nonetheless, because of the increasing temperature with depth, rocks of the mesosphere are not as strong or brittle as in the lithosphere and still have the ability to flow but at a slower rate than in the asthenosphere.

**Outer Core**

At the base of the mantle or mesosphere, temperature increases abruptly across the mesosphere–outer core boundary, reflecting the presence of molten iron and nickel. In response to gravitational and thermal gradients, the molten iron and nickel are convecting or circulating, promoting heat flow into the overlying mantle (resulting in abrupt temperature increases across the boundary). This circulation of molten iron in conjunction with Earth’s rotation produces a geodynamo that gives rise to the planet’s magnetic field. The liquid nature of the outer core is deduced from seismic wave data (see discussion below). Receiving stations on the opposite side of the planet from which an earthquake occurs will not receive any S-waves (also known as shear or secondary waves), which are attenuated when they encounter liquid material.

**Inner Core**

The inner core is compositionally the same as the outer core but is a solid rather than a liquid even though the temperature has risen to about 6000°C (depending on the model used). The transition from liquid in the outer core to solid in the inner core results from the extreme pressure at these depths. The radius of the inner core is about 1300 km.

**Evidence of Earth’s Compositional and Rheological Layers**

Our understanding of Earth’s compositional and rheological layers is not known from drilling, as the deepest drill hole is about 12 km deep, which is a mere pinprick into the Earth’s interior. Rather, our understanding comes from several sources, including meteorites, material erupted from volcanoes, and the nature of Earth’s rotation and precession (or wobble) of Earth’s axis. Primarily, though, most of what we know of the Earth’s internal compositional make-up stems from the study of seismic waves. These waves image the interior of the Earth, much like a computerized axial tomography (CAT) scan discloses internal components of the human body. The speed, direction, and propagation of these waves change in response to the density and composition of the material traversed. By collecting seismic wave data from receiving stations across the planet, the Earth’s internal compositional layers can be successfully modeled and imaged (Figure 3.4). Earthquakes generate two types of waves that travel through the interior of the Earth: P-waves, or primary (compressional) waves, and S-waves, or secondary (shear) waves. P-waves travel through solids, liquids, and gases, but S-waves travel only through solids, because liquids and gases have no elasticity to support shear stresses. Therefore, the liquid nature of the outer core is indicated because seismic receiving stations on the opposite side of the Earth from which an earthquake occurs receives no S-wave signal, only a P-wave response. The size of the resulting S-wave shadow zone is a direct indication of the diameter of the core (Figure 3.4).
There are three main sources of Earth's internal heat. First is residual heat left over from the formation of the planet (primordial heat) about 4.6 billion years ago. This heat is a product of the first law of thermodynamics, which states that energy is conserved. Our planet formed by accretion of colliding meteorites or larger chunks of space debris called planetisimals. Of movement was converted to heat energy after collision, resulting in a largely molten proto-Earth, leading to the eventual gravitational separation of heavy and light elements to form the core, mantle, and crust as described above. Because rock is a good insulator, the deep interior of our planet has stayed hot, with heat flowing outward toward the surface. This outward flow of heat, while fairly uniform at depth from the core through the mantle, becomes irregularly distributed as it flows through the crust, being concentrated in select zones due to plate tectonics (discussed in Chapter 4) and influencing the distribution of areas having high and low geothermal heat flow at the Earth's surface.
A second source of heat comes from the radioactive decay of select elements, principally from uranium, thorium, rubidium, and potassium. These elements are largely concentrated in the crust because their large atomic radii are less compatible in mineral structures in the mantle due to the high pressures there, favoring dense mineral species. As a result, about 60% of the heat in continental crust is due to radioactive decay of these four elements (Glassley, 2015). Nonetheless, these radioactive elements are present in the mantle, and even though their concentration is low there, the large volume of the mantle makes up for the low concentration, indicating that a significant amount of heat coming from the mantle is due to radioactive decay. Recent studies of Earth’s internal heat flow budget indicate that the proportion of primordial heat and radiogenic heat to total heat flow is about equal and in total amounts to about 47 terrawatts (TW) (Davies and Davies, 2010; Gando et al., 2011; Korenga, 2011). For comparison, the total installed world power capacity in 2012 was 5.55 TWe (EIA, 2016). The takeaway, clearly, is that Earth’s internal heat energy can provide a significant contribution toward supplying the energy needs of civilization. Over 50% of the total heat flow is contributed by convection in Earth’s mantle, with about 24% coming from the crust and supplied by a mixture of conduction, hydrothermal convection, and vertical and horizontal movement (advection) of localized zones of magma (Figure 3.5).

A third, albeit minor, source of heat is from gravitational pressure. When something is squeezed it heats up, and when expanded it cools. For gases, this behavior is described by Charles’ law; a similar process happens with solids, except that the changes in volume are much smaller for a given increase in pressure. Again, because rocks are good insulators, the escape of heat from Earth’s surface is less than the heat generated from internal gravitational attraction or squeezing of rock, so heat builds up with depth.

Other local sources of heat include frictional heating along earthquake faults. This frictional heating can be sufficiently intense to actually partially melt the rock, producing what is called pseudotachylite. Indeed, a small amount of heat tapped by geothermal power plants located along major active faults, such as the San Andreas fault in California or active faults in Nevada, probably comes from frictional heating as rocks grind past on either side of the fault.

HEAT TRANSFER MECHANISMS IN THE EARTH

A flux of heat is emitted from every square meter on Earth’s surface; in some places it is notably higher, particularly near the boundaries of the tectonic plates, than in other places. Overall, however, the average heat flux or flow for the Earth is about 87 milliwatts per square meter (mW/m²). Multiplying this value by the total global surface area of $5.2 \times 10^{11}$ m² yields a total heat or power output of about $4.7 \times 10^{13}$ W or 47 TW (thermal) as noted above. The heat flow for continents averages 65 mW/m², and the average heat flux for oceanic crust is 101 mW/m². The difference reflects the thinner character of oceanic crust with hot mantle rocks at comparatively shallow depths and the insulating nature of thicker continental...
crust. Indeed, if it were not for the ocean, whose depth averages about 3.7 km, much of the oceanic crust would have the potential for harnessing geothermal energy. But then again, without the oceans there would be a dearth of water, which is the primary vehicle for transferring heat energy from hot rocks at depth to the surface (see later discussion). Heat can be transferred by three main mechanisms: conduction, convection/advection, and radiation. The first two are relevant for the solid Earth, as radiation applies mainly to the transfer of electromagnetic radiation through space, such as sensing heat from a campfire or the transfer of light from the Sun.

**FIGURE 3.5** Graph showing the change in temperature (heavy solid line) of the Earth from its surface to the core (Earth’s geothermal gradient). Also shown is the solidus, or the temperature at which rocks begin to melt. Note that the geothermal gradient is highest near the surface, indicative of conductive heat flow, but becomes more gradual with depth, indicating a combination of convective and conductive heat flow. The highlighted yellow layer marks the asthenosphere where the temperature of the solidus and that of the Earth are close, resulting in rheologically weak rock. The layer above the asthenosphere is the lithosphere where the temperatures of the Earth and solidus are further apart, making for rheologically strong rock: (Adapted from Ammon, C.J., *Earth’s Origin and Composition*, SLU EAS-A193 Class Notes, Penn State Department of Geosciences, University Park, 2016, http://eqseis.geosc.psu.edu/~cammon/HTML/Classes/IntroQuakes/Notes/earth_origin_lecture.html.)
**Conductive Heat Flow**

Conduction is the transfer of heat by contact (transfer of energy from one atom to the next) and is an important means of heat transfer within the Earth. The overall geothermal gradient of Earth—the change in temperature with depth—is largely governed by conductive heat transfer. This gradient is high or changes rapidly near the surface but becomes more gradual at depth (Figure 3.5). This rapid change in temperature with depth is indicative of conductive heat flow, because, in the absence of circulating fluids, rocks are good insulators. The geothermal gradient averages about 25 to 30°C per km for the upper crust (top 10 km or so), whereas in geothermal areas the geothermal gradient is about double to perhaps three times that of non-geothermal regions. In active volcanic regions, the geothermal gradient can be as high as 150°C per km, such as at Yellowstone National Park, and the heat flux can be 500 mW/m² or even more.

Heat flux is governed by Fourier’s law, which states that the flow of heat ($Q$) depends directly on the thermal conductivity ($k$, in units of watts per meter kelvin, or W/m·K) of the material and the geothermal gradient ($\Delta T/\Delta x$ or $\nabla T$). This gives us the equation $Q = k \times \nabla T$. For example, if an exploration well is drilled in granite and encountered a temperature of 200°C at a depth of 1500 m, what is the heat flux at the site?

$$Q = \frac{k_{\text{granite}} \times (473 \text{ K} - 298 \text{ K})}{1500 \text{ m}} \quad (3.1)$$

Thermal conductivity itself is modestly sensitive to temperature and generally decreases as temperature increases for Earth materials (Clauser and Huenges, 1995). An average value of granite over this temperature range would be about 2.4 W/m·K (Glassley, 2015). Substituting these values into the equation yields the following:

$$Q = 2.4 \text{ W/m·K} \times 175 \text{ K}/1500 \text{ m} = 0.280 \text{ W/m² or 280 mW/m²}$$

which would be a very promising heat flow for developing geothermal energy.

For continental crust, the minerals feldspar and quartz are the most common, yet there is a significant difference in the thermal conductivity of quartz and feldspar (Glassley, 2015), such that the thermal conductivity of quartz averages about twice that of alkali feldspar. Thus, the thermal conductivity of a rock will be strongly dependent on the proportion of these two minerals, which in turn will directly influence the heat flow.

Related to thermal conductivity is thermal diffusivity, which measures how quickly an object changes temperature in the presence of a thermal gradient. Thermal diffusivity has the units of square meters per second (m²/s). Thermal diffusivity is defined by the ratio of thermal conductivity to the heat capacity, by volume, of a material. Heat capacity measures how much heat is required to raise
the temperature of a unit volume of material by 1 K. Minerals have thermal dif-
fusivity values of $1 \times 10^{-6}$ to $10 \times 10^{-6}$ m$^2$/s, whereas most metals have diffusivity
values in the range of $1 \times 10^{-4}$ to $5 \times 10^{-4}$ m$^2$/s, or about 100 times the diffusivity of
minerals. Also affecting conductivity and diffusivity is the porosity or open space
in rocks (porosity is discussed in more detail in Chapter 5). Pores in rocks can be
filled with water or air or a mixture of the two. Because water conducts heat more
readily than air, the thermal conductivity of a water-saturated rock will be 3 to 4
times that of its dry equivalent. Furthermore, conductivity is also dependent on
pore size such that larger pores have a lower conductivity for a given water content
(Glassley, 2015).

As a result, part of the accurate characterization of the geothermal energy poten-
tial of a given region requires measuring and understanding the properties of the
geological materials in which the system is developed. How is this important for
geothermal power production? Imagine a site having high heat flow but also char-
acterized by quartz-rich rocks, which have relatively high thermal conductivity.
Although heat is transferred efficiently to water for production, the cooler injection
water could unfavorably cool the reservoir rocks, which would lower the system’s
enthalpy and power/energy potential. Thus, the rates of production and injection
must be such that the system is not adversely perturbed, and determining the pro-
duction and injection rates requires accurate characterization of the thermal proper-
ties of the geological materials.

Examples of conductive geothermal systems include some deep sedimentary
basins and geopressed reservoirs, such as those found along the Gulf Coast of
the United States. The Paris Basin is an example of a deep sedimentary aquifer
whose geothermal fluids are used directly for space heating. The flow of water is
slow enough that there is enough time to be heated by the conductive heat flow
from the rocks. This happens because there is a general reduction in permeability
(flow of water through rock, as discussed in Chapter 5) with depth, which retards
the fluid’s ability to circulate easily. In geopressed reservoirs, permeable water-
bearing horizons are deeply buried (generally >2 km) and are isolated by sur-
rounding impermeable rock. These are self-contained systems in which the pore
water was trapped with the sediments at the time of deposition. Because they are
isolated from the surface, the pore water is under the weight of the overlying rock
(lithostatic) rather than a column of water (hydrostatic). The water is thus pretty
much stagnant and is heated conductively in response to the region’s geothermal
gradient of about 50°C/km.

A final example of conductively heated geothermal systems consists of engi-
neered geothermal systems (EGSs) in which hot rocks exist but permeability or
water content is sufficient to produce a circulating hydrothermal system. These
conductive systems are being explored in places to artificially produce convective
systems (see next section) through controlled fracturing of the rock. An EGS
project, at Newberry Volcano in central Oregon, has proved encouraging with
regard to developing improved permeability in hot rocks through the injection of
cold water. (EGSs, deep sedimentary aquifers, and geopressed reservoirs are
discussed in Chapter 11.)
**Convective (Advective) Heat Flow**

Technically, the movement of heat by bulk fluid flow is advection; however, convection is the more widely used and general term and embraces both advection and conduction, meaning that as heat is transferred by moving material some heat is also transferred by conduction through contact with surrounding material. The slower the movement of the material, the greater the proportion of heat transferred by conduction. For simplicity, we will use the more widely used term convection, understanding that the bulk of heat is transferred by movement of material and a lesser amount by conduction. Because convection involves both movement of material (advection) and thermal diffusion (conduction), it is the most effective means of energy transfer within the Earth.

Convection develops in response to buoyancy forces in the presence of a gravitational field. As material is heated it becomes less dense and will begin to rise. To replace the rising material, cooler (and more dense) material sinks, where it too might be heated and also begin to rise, resulting in a convection cycle. Without convection, a body of water, for example, can become thermally and density stratified, such that warm, less dense water lies near the surface and cooler more dense water at depth. If fluids are convecting, however, they are mixing; thus, temperature changes little with depth over the convecting interval. Recognizing zones of convection from drilling can be an effective exploration tool for identifying prospective geothermal reservoirs (discussed in Chapter 8).

As established, the solid Earth is overall density stratified with a dense iron-rich core and a low-density, outer crust; however, it is not static because the hot, liquid outer core is a potent source of energy. Although the overlying mantle is solid, it has the ability to flow slowly but significantly on the order of the geologic time scale. The rate of flow is controlled in part by the strength of the energy source but also in part on the viscosity of the material. Viscosity is a property that measures the resistance to flow of a material when stressed. For example, molasses is more viscous than water. For most materials, viscosity is inversely related to temperature; as temperature increases, the viscosity decreases, similar to heating honey. Thermally disturbed portions of the lower mantle, perhaps situated above focused zones of upwelling in the underlying and convecting molten outer core, will be gravitationally unstable relative to overlying (and adjacent) cooler and denser mantle and will begin to rise buoyantly upward, producing a system of convection cells (Figure 3.6).

**Rayleigh Number**

Factors that promote convection are low viscosity, thermally induced expansion, a gravitational field to exert buoyancy forces, and low thermal conductivity to create a strong thermal gradient and drive buoyancy forces. Quantitatively, conditions that promote convection can be represented by the ratio between buoyant and viscous forces, or what is termed the *Rayleigh number* ($Ra$), which is represented quantitatively below:

$$Ra = \left( \frac{g \times \alpha \times d^3}{\nu \times \kappa} \right) \times \Delta T$$

(3.2)
where
\[ g = \text{Acceleration of gravity (9.8 m/s}^2\text{)}. \]
\[ \alpha = \text{Coefficient of thermal expansion (1/K)}. \]
\[ d = \text{Depth interval over which the temperature change occurs (m)}. \]
\[ \nu = \text{Kinematic viscosity (m}^2\text{/s)}. \]
\[ \kappa = \text{Thermal diffusivity (m}^2\text{/s)}. \]
\[ \Delta T = \text{Vertical temperature change (K)}. \]

As a result, Ra is a dimensionless number that provides an indication of whether convection will occur or not and therefore indicates whether the dominant form of heat flow will be by conduction or convection. When Ra is $>1000$, convection is the dominant heat transfer mechanism; when Ra is $<1000$, conduction is the dominant form of heat flow.

The Rayleigh number for the mantle ranges between $10^5$ and $10^7$, indicating that the mantle is mobile and that the main form of heat transfer to the Earth’s surface is by convection. The movement of mantle material is a principal driver for plate tectonics, which accounts for much of the distribution and types of geothermal resources across the planet (discussed in Chapter 4).

**Convection in the Upper Crust**

Currently exploited geothermal systems are typically less than 3 km deep and consist of convective hydrothermal systems, either liquid or vapor dominated. The fluid must reside in rock reservoirs that allow fluids to circulate, which requires connected open space or permeability (discussed in Chapter 5); otherwise, they would be mainly undeveloped conductive reservoirs. When a convective hydrothermal
reservoir is intercepted by drilling, the geothermal gradient will decrease suddenly, reflecting the thermal mixing of fluid circulation. This is different than thermal stratification, which indicates conductive heat flow zones that commonly bound the tops and bottoms of geothermal reservoirs. In some cases, the geothermal gradient increases again below the convecting hydrothermal reservoir, whereas in other cases the gradient can decrease below the reservoir, reflecting lateral outflow of hydrothermal fluids above cooler (and more dense) groundwater recharge zones. The Casa Diablo geothermal field in the Long Valley caldera in California is an example of the latter (Figure 3.7). Most of the heat flow through the upper crust occurs by conduction (Figure 3.8), and convecting hydrothermal systems require special geologic characteristics. Such characteristics include a source of water, good permeability, properly positioned impermeable cap rocks, and a focused source of heat, such as a body of magma in the upper crust. These conditions are not met everywhere, and just because heat flow may be high in a region does not indicate whether or not potentially developable convecting geothermal reservoirs are present.

FIGURE 3.7 Temperature profiles of four drill holes into the Casa Diablo geothermal system in the Long Valley caldera, California. Note the areas of conductive heat flow at shallow depths of all drill holes where there is a rapid change of temperature with depth. The convective zones are characterized by little change in temperature with depth, reflecting circulation and thermal mixing. Note that in all holes, one or more temperature reversals occur, reflecting deeper, cooler aquifers. (Adapted from Glassley, W.E., Geothermal Energy: Renewable Energy and the Environment, 2nd ed., CRC Press, Boca Raton, FL, 2015.)
HEAT FLOW MAPS

The Southern Methodist University Geothermal Laboratory (SMUGL) has been instrumental in compiling drill-hole data and generating and updating a series of maps showing heat flow in the United States. The team of researchers there has also developed, from the compiled heat flow data, a series of temperature-at-depth maps from 3.5 to 9.5 km. These maps help illustrate prospective regions for developing geothermal energy for power and direct use, most of which are located in the western United States (Figures 3.9 and 3.10). Researchers at SMUGL have also developed maps showing the potential for engineered geothermal systems (EGSs) for each of the states based on the heat flow information. For example, Nevada, which has an installed current geothermal power capacity of 580 MWe (as of 2015) from conventional convective geothermal systems, could increase it geothermal power output to 41k MWe if only 2% of its EGS potential is recovered. That number could swell to 288,000 MWe if just 14% of its EGS potential is realized (SMU, 2016). Although such potential is impressive, EGSs are hampered by their still experimental nature and associated high costs which must compete with currently inexpensive natural gas—the fossil fuel of choice for power generation because carbon emissions are about half of those of coal (see Chapter 9). However, as described by Allis et al.
FIGURE 3.9  Heat flow map of the United States for 2011. Ochreous orange to more deeply red indicates heat flow values in excess of 80 mW/m$^2$. The highest value on the map is the pinkish red, which is >150 mW/m$^2$, which would be at Yellowstone National Park. This and the following figure illustrate the large area of geothermal energy potential covering much of the western United States. (Adapted from Blackwell, D.M. et al., Geothermal Resources Council Transactions, 35, 1545–1550, 2011.)

FIGURE 3.10  Map showing temperatures at a depth of 4.5 km. Notice the large area of temperatures of 150°C and higher across much of the western United States. Almost all of northern Nevada has temperatures that are >150°C, including numerous scattered regions with temperatures between 175 and 200°C. This depth level is largely the realm of engineered geothermal systems (EGSs), which if developed (just in small part, ~10%) could greatly expand (by one to two orders of magnitude) geothermal power production. However, accessing this potential energy resource would be expensive, mainly due to the deep levels of drilling required, making it difficult for this technology to compete economically with currently producing, more shallow geothermal reservoirs and natural gas-fired power plants. (Adapted from Blackwell, D.M. et al., Geothermal Resources Council Transactions, 35, 1545–1550, 2011.)
(2012), if sedimentary aquifers exist at these depths and are permeable, they might serve as a bridge to actual development of EGSs, offer large potential flow rates, and be cost competitive if current low energy prices were to rise modestly.

**SUMMARY**

The interior of the Earth is compositionally and rheologically partitioned into distinct layers. Compositionally, the Earth’s interior consists of an iron-rich metal core, a mantle, and a thin crust. The mantle makes up the largest volume of the Earth’s interior and consists of dense, iron- and magnesium-rich rocks. The crust consists of two types: oceanic and continental. Oceanic crust consists of more dense basalt and is relatively thin (<1 to 7 km thick). Continental crust is made of less dense granitic and metamorphic rocks and can be as much as 70 km thick underneath mountain belts. This compositional division developed very early in the Earth’s history, when the planet was still largely molten. Dense constituents settled to the center to form the core, and less dense elements rose toward the surface to form the mantle and crust.

Due to changes in temperature and pressure within the Earth, the compositional layers develop different rheological (mechanical) properties, ranging from solid (brittle and strong), to weak and ductile, to molten. These different mechanical layers are the lithosphere, asthenosphere, mesosphere, outer core, and inner core. The lithosphere consists of the crust and uppermost mantle and is strong rock that when stressed to a certain limit will break (brittle behavior) rather than bend (ductile). On average, the lithosphere is about 100 km thick and comprises the tectonic plates discussed further in Chapter 4. The underlying asthenosphere consists of weak rock near its melting point; however, it is not molten but still largely solid. Because of the hot temperatures, the rock in the asthenosphere has the ability to flow ductilely. The asthenosphere is about 200 km thick, but its lower boundary with the underlying mesosphere is gradational. The mesosphere makes up the bulk of the mantle, and the rocks there are stronger due to the increasing pressure with depth, but they still have the ability to flow, although more slowly than in the asthenosphere. The outer core consists of liquid iron–nickel metal; the inner core is the same composition but a solid due to the extreme pressure.

The compositional and rheological nature of the Earth’s interior is largely based on the study of seismic waves whose direction and speed of propagation are based on compositional and physical properties of the material through which they pass. For example, one type of seismic wave does not travel through liquids, and as such seismic receiving stations on the opposite side of the planet from where an earthquake occurs will not detect that wave, creating a shadow zone. The size of the shadow zone is a direct reflection of the size of the liquid, outer core.

Earth’s internal heat has two primary sources. The first is heat left over from the tumultuous formation of the Earth, when kinetic energy of celestial collisions was transformed to heat energy (primordial heat). The second major source is radioactive decay of select elements, mainly uranium (U), thorium (Th), rubidium (Rb), and potassium (K). The contribution of each source is about equal. Heat flows from the Earth’s interior toward the surface via two main mechanisms: conduction and
Convection. Conductive heat flow is transfer of energy by contact, also known as thermal diffusion. Conductive heat flow is mainly operative in the Earth’s core and crust. Convective heat flow is heat transferred by motion, with subsidiary contribution by conduction. Convective motion is induced by buoyant forces that arise from thermal gradients in a gravitational field. If material becomes hotter than its surroundings, its density is reduced, causing the heated material to rise. Conversely, the surrounding cooler material is more dense and sinks to replace the rising hotter material. Convective heat transfer occurs in the liquid outer core and the rheologically ductile mesosphere and asthenosphere, where buoyant forces exceed viscous forces as measured by the Rayleigh number.

Geothermal heat flow and temperature-at-depth maps illustrate that much of geothermal resources developed and yet to be developed occur in the western United States. For example, most of northern Nevada has a heat flow of $>80 \text{ mW/m}^2$, in places $>100 \text{ mW/m}^2$. At a depth of 4.5 km, the temperature of crustal rocks in northern Nevada is $>150^\circ\text{C}$ and in places as high as $200^\circ\text{C}$. Although this environment (realm of engineered geothermal systems) represents a vast reservoir of heat and potential source of energy development, it is expensive to access and cannot compete economically with currently developed sources of geothermal energy or natural gas-fired power plants. However, if sedimentary aquifers exist at these depths and are permeable, they could serve as major sources of available geothermal energy if energy prices rise modestly.

**SUGGESTED PROBLEMS**

1. Explain what factors control whether heat flow will be conductive or convective? What type offers the greatest potential for geothermal energy development and why?
2. Will the Rayleigh number of material affect the heat flow measured at the surface? Why or why not?
3. Assume that a well is drilled in dry sand to a depth of 2500 m and the temperature measured at the bottom is $150^\circ\text{C}$. For simplicity, assume that the thermal conductivity of dry sand is a constant between $10^\circ\text{C}$ and $200^\circ\text{C}$. Is there likely to be a geothermal resource? Explain why or why not.
4. Imagine you are a geologist and you have drilled hole RD08 whose temperature profile with depth is shown in Figure 3.7. Using the temperature–depth profiles of the three other wells shown in Figure 3.7, should you continue drilling deeper or stop at the current depth? Justify your position.

**REFERENCES AND RECOMMENDED READING**


