1 Introduction

1.1 PETROLOGY AND ITS SCOPE

Petrology is the science dealing with the description, classification, modes of occurrence, and theories of the origins of rocks. Its emphasis is commonly chemical and mineralogical, but it draws heavily on many disciplines, including the basic physical sciences, mathematics, geophysics, structural geology, and geochemistry. Its tools range from the simple hammer and hand lens, to sophisticated devices such as the electron microprobe or the laboratory equipment capable of reproducing conditions deep within the Earth. Its goal is to provide an understanding of the great diversity of rocks found on the surface of the Earth (and other planets), and to provide insight into the nature of those materials within the Earth that are not accessible to direct observation but play such important roles in the Earth's history.

Rocks can be divided into three main groups: igneous, sedimentary, and metamorphic. Those formed from the solidification of molten material are termed *igneous*, whereas those that originate from the deposition of material from water or air are termed *sedimentary*, and those formed from a previously existing rock by some process of change are termed *metamorphic*.

The study of igneous and metamorphic rocks, the subject of this book, is commonly treated separately from the study of sedimentary rocks, mainly because of the different approaches used. Sedimentary rocks are formed by processes that, for the most part, are observable on the surface of the Earth. Careful examination of present-day environments of deposition can, therefore, provide information on the origins of most sedimentary rocks. Igneous and metamorphic rocks, on the other hand, are formed largely by processes operating within the Earth and therefore not directly accessible to observation; the origin of these rocks must, consequently, be deduced through physical–chemical arguments. Also, at the higher temperatures existing within the Earth, reactions proceed more rapidly than they do on the surface, and thus principles of chemical equilibrium are more applicable to the study of igneous and metamorphic rocks than they are to most sedimentary ones.

Petrologic studies fall into two general categories: the identification and classification of rocks, and the interpretation of these data and the generation of theories on the origin of rocks. The early emphasis in petrology, as in other natural sciences, was on description and classification. There were, nonetheless, many lively discussions concerning the origins of rocks, such as that in the early nineteenth century between Werner and his student von Buch on whether basalt was a sedimentary or volcanic rock. An excellent summary of early ideas on the origin of basalt and other igneous rocks can be found at www.lhl.lib.mo.us/events_exhib/exhibit/exhibits/vulcan/index.shtml. This web site provides illustrations and brief descriptions from 66 rare books and journals, published between 1565 and 1835, held in the Linda Hall Library of Science, Engineering & Technology in Kansas City, MO. Most of the early work, however, involved the cataloging of the constituents of the Earth's crust. During the latter half of the nineteenth and early part of the twentieth centuries vast amounts of petrologic data were collected, from which came an enormous number of rock names and many different classifications. Despite the surfeit of names, generalizations concerning rock associations and mineral assemblages did emerge, which, in turn, allowed for simplifications in the classifications. Many different rocks that had previously been given separate names could be considered varieties of a single type, and this naturally led to theories explaining the associations. Fortunately, the rock-naming era is over, and modern petrology employs only a small number of rock names.

With recent investigations of the ocean floors, the inventory of rock types available for study in the Earth's crust is almost complete, and today most petrologists are concerned mainly with the genesis of rocks. This change in emphasis has been stimulated by development of experimental techniques that allow us to imitate, in a limited way, rock-forming conditions and processes within the Earth's crust, mantle, and even the core (see Consortium for Materials Properties Research in Earth Sciences www.compres.us). In fact, with such techniques it is possible to investigate the petrology of parts of the Earth that, while being of great importance as source regions for many igneous rocks, were virtually unavailable for examination and classification by the early petrologists.

Most processes involved in the formation of igneous and metamorphic rocks occur within the Earth and hence are not subject to direct observation (volcanic eruptions are obvious exceptions). Ideas on the origin of these rocks are, therefore, based on interpretations of field observations in the light of experimental and theoretical studies. Nature provides fragmentary evidence of events and processes that have formed rocks, and it is up to the petrologist to assemble this evidence into a coherent story. The task is similar to that faced by a

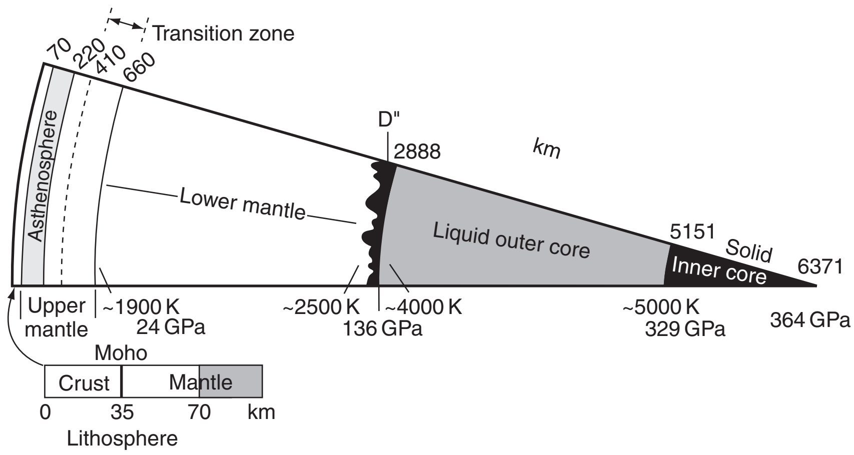

Fig. 1.1 Major divisions of the Earth based on seismic data (Masters and Shearer, 1995; Lay *et al.*, 2004). Temperatures are from Boehler (2000). Note that the lithosphere, which is the layer in which almost all rocks that are exposed on the Earth's surface are formed, constitutes a very small fraction of the entire Earth.

detective who, on arriving at a crime scene, must deduce what happened from the limited evidence that remains. Unlike the legal case, however, which requires that a verdict be reached in a court of law, petrologic questions rarely have absolute verdicts. Indeed, as new data and new research techniques become available, interpretations frequently have to be revised, and this, in itself, makes petrology an active and interesting field for study.

Before beginning a discussion of igneous petrology, it will be useful to review briefly the major structural units of the Earth and the distributions of pressure and temperature within it. This will provide a framework for later discussions of rock-forming processes.

1.2 MAJOR STRUCTURAL UNITS OF THE EARTH

The division of the Earth into three major structural units, *crust*, *mantle*, and *core* (Fig. 1.1), was made early in the history of seismology on the basis of major discontinuities in both compressional (P) and shear (S) wave velocities (e.g. Masters and Shearer, 1995). The discontinuities are best explained as boundaries separating chemically and mineralogically distinct zones. Although their compositions cannot be determined with certainty, reasonable estimates can be made from seismic velocities, the mass and moment of inertia of the Earth, and solar and meteoritic abundances of elements. For a more detailed section through the Earth, refer to Figure 23.4.

The crust contains high concentrations of alkalis, calcium, aluminum, and silicon relative to solar abundances. In continental regions, where the crust is approximately 35 km thick, these elements are so abundant that when combined to form the common minerals quartz and feldspar, the rock approximates the composition of granodiorite. Many light elements and certain heavy ones, such as uranium, thorium, and zirconium, which have difficulty substituting into the structures of common minerals, are also concentrated in this zone. Because some of these elements are radioactive and generate heat, temperature measurements at the Earth's surface can be used to show that this heat-generating granitic zone must be limited to the upper 10 to 20 km of the continental crust (Section 1.4 and Problem 1.4). Increasing seismic velocities with depth in continents indicate the presence of rocks with higher concentrations of iron and magnesium at depth; these are probably basaltic in composition. Basaltic rocks constitute most of the oceanic crust which is about 6 km thick (see Fig. 1.2).

The Mohorovičić discontinuity (Moho or M discontinuity), at the base of the crust, marks a sharp increase in seismic velocities, which can be accounted for by the disappearance of feldspar, the most abundant mineral in crustal rocks. Below the Moho, the rock is thought to consist mainly of olivine and pyroxene with lesser amounts of spinel and garnet and thus is called a peridotite. Samples of this rock are brought to the Earth's surface by certain types of explosive volcanic eruption and confirm the seismological deductions as to its composition.

The disappearance of S waves at the core–mantle boundary indicates the presence of liquid below this. If the Earth were initially formed from accreting chondritic meteorites, which is the favored model, and with its known internal distribution of mass, the core is believed to be composed largely of iron and some nickel. However, high-pressure experiments with diamond anvils and shock waves indicate that the core's density must be decreased slightly by the presence of small amounts of lighter elements such as silicon, oxygen, sulfur, and hydrogen. The Earth's magnetic field is generated by convection within this liquid. With increasing depth and pressure, the liquid undergoes a phase change to form the solid inner core; that is, the inner–outer core boundary marks the floor of a giant magma chamber, which is slowly solidifying from the bottom up as the Earth loses heat. Seismic velocities in the inner core are found to be faster in a north–south direction than in an equatorial path, indicating that the inner core is crystallographically and/or structurally anisotropic. The inner core may also rotate faster than the mantle, which may play a role in generating the Earth's magnetic field.

In addition to these traditional structural units, modern seismologists detect many other discontinuities with which they further subdivide the Earth (Fig. 1.1). Moreover, by

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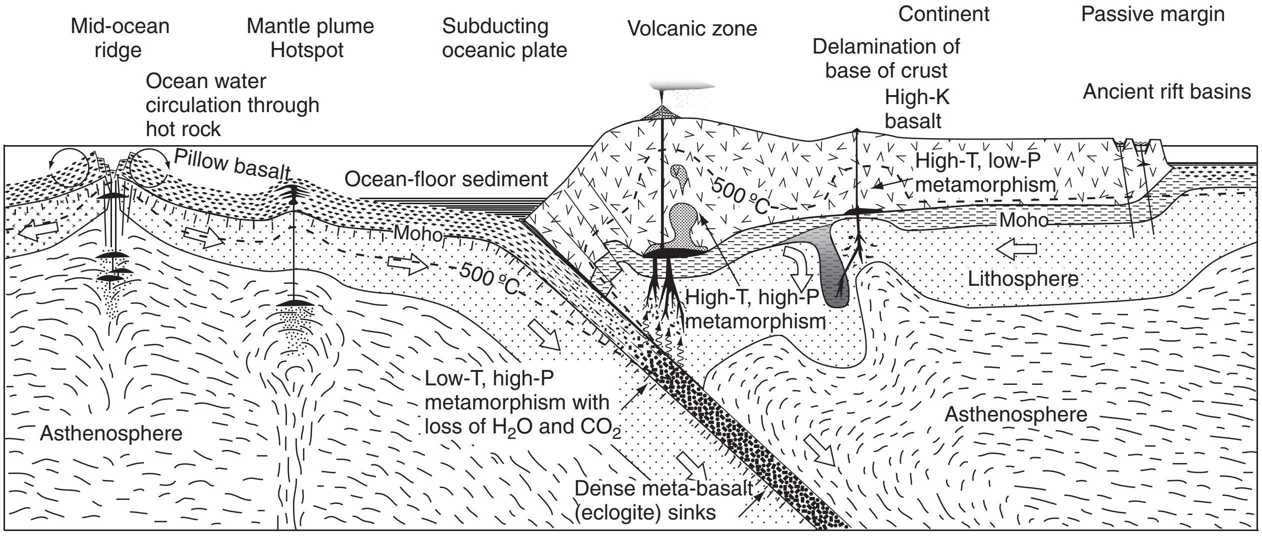

Fig. 1.2 Schematic, vertically exaggerated section through the lithosphere and asthenosphere showing an oceanic plate moving away from a spreading axis and being subducted beneath a continental plate. Most igneous rocks are formed either at spreading axes or above subduction zones, and most metamorphic rocks are formed in the vicinity of convergent plate boundaries. Slabs at the base of the continental lithosphere may be metamorphosed into rock that is dense enough to delaminate and sink into the mantle. As it sinks, the rising asthenosphere causes heating and generation of igneous and metamorphic rocks in the crust above. The range of temperatures and pressures rocks experience are determined largely by plate motions. Note how the depth (pressure) of the 500 °C isotherm varies across the section as a function of its position in a plate.

analyzing seismic recordings of earthquakes from arrays of stations around the planet, the three-dimensional distribution of seismic velocities can be determined. These tomographic studies reveal that the traditional zones within the Earth exhibit considerable complexity.

Of particular importance to petrology and plate tectonics is a zone extending from depths of approximately 70 to 250 km in which shear-wave velocities markedly decrease; this zone is known as the *low-velocity layer*. Because it has considerably less strength than overlying layers it is also referred to as the *asthenosphere* or weak zone. The more rigid overlying rocks, which include the upper part of the mantle and the crust, constitute the *lithosphere*. The reduced velocities and strengths of rocks in the asthenosphere are thought to result from small degrees of partial melting, possibly as much as 10% in regions of high heat flow. This partial melt is an important source of magma and a lubricant to ease the tectonic movements of the lithospheric plates.

Between 410 and 660 km, seismic velocities increase rapidly. Most of this change occurs at two major discontinuities. One, at 410 km, occurs at pressures where laboratory experiments indicate that the crystal structure of olivine changes to that of a distorted spinel, and the other, at 660 km, is at pressures where perovskite and simple oxides predominate. Indeed, below 660 km, most silicate minerals are converted to the dense perovskite structure and magnesiowüstite ([Mg,Fe]O). Between 660 km and the core–mantle boundary, velocities increase in a slower and steadier manner. This is known as the lower mantle. Little is known about its composition, yet it comprises 56% of the planet. Tomographic studies reveal considerable lateral variation in seismic velocities within the mantle, which may be due to changes in temperature or composition.

At the base of the lower mantle is a 200- to 350-km-thick heterogeneous thermochemical boundary layer separating the mantle from the core (Lay *et al*., 2004). This layer, designated D″ (*D Double Prime*), is marked by decreased gradients in seismic velocities with depth. Within this layer, at depths that range from 150 to 350 km above the core–mantle boundary, is a prominent discontinuity marked by an increase in seismic velocities, which may be caused by a still higher density form of the perovskite structure known as post-perovskite (Murakami *et al*., 2004). Shear waves become polarized in passing through D″ into faster horizontal – and slower vertical – traveling components. This anisotropy indicates the presence of preferred orientations either of crystals or of chemically and physically distinct structures, such as bodies of melt. At the base of D″ is a 5- to 40-km-thick ultralow-velocity zone, which is almost certainly partially melted by heat released from the rapidly convecting core beneath.

The D″ layer may play important roles in the dynamics and evolution of our planet. For example, some of the heterogeneity in D″ can be correlated with important features in the lithosphere. Seismic velocities in D″ are generally lower beneath surface hot spots, such as Hawaii, and higher in regions beneath subducted slabs, with lower and higher velocities possibly corresponding to hotter and cooler regions, respectively. Thus, major tectonic features in the lithosphere may owe their origin and location to the way in which heat is transferred out of the core through the D″ layer. Although the chemical makeup of D″ is uncertain, it has been proposed as a reservoir of ancient material, possibly dating back to the accretion of the planet (Tolstikhin *et al*., 2006). It may also be a “graveyard” for subducted slabs of lithosphere or even a chemical reaction zone between the core and mantle.

In later chapters when large numbers are used to describe the depths and pressures under which magmas are generated, it will be easy to lose sight of the fact that we will be dealing with only the extreme outer skin of the Earth. Almost all terrestrial rocks that geologists are able to examine are formed within the upper part of the lithosphere. We therefore can examine directly only a minute fraction of the whole Earth. For perspective, then, readers are advised to refer, from time to time, to Figure 1.1.

Despite the small relative volume of the lithosphere, it contains a great variety of rocks and structures. Much of its complexity has been elucidated by modern plate tectonic theories, which show the lithosphere to be a dynamic part of the Earth undergoing constant change as a result of heat transfer from the planet's interior to its surface. Most igneous and metamorphic rocks owe their origins directly to plate tectonic processes.

Figure 1.2 summarizes the principal components of the lithosphere in terms of their involvement in plate tectonics. The surface of the Earth is divided into ten major lithospheric plates and several smaller ones. Each plate moves as a rigid unit with respect to adjoining ones. Where plates move apart, new material must fill the intervening space. Thus, along mid-oceanic ridges, which are divergent plate boundaries, the rising asthenosphere decompresses and partially melts to form basaltic magma that buoyantly rises and solidifies to create new lithosphere. Elsewhere material must be consumed if the Earth's circumference is to remain constant. This occurs at convergent plate boundaries where cold lithospheric slabs are subducted into the mantle. Material at the base of the crust may also be converted into metamorphic minerals that are so dense that the lower part of the crust delaminates and sinks into the mantle. Plates may also slip past one another along transform faults, as happens on the San Andreas fault.

Lateral and vertical movements of lithospheric plates produce significant lateral variations in the temperature of rocks at equivalent depths in the lithosphere. In Figure 1.2, a line joining all points at 500 °C (the 500 °C isotherm) is traced through a typical section of lithosphere. Beneath a stable continental region this isotherm is at a depth of approximately 40 km, a position reflecting a balance between the rate at which heat rises into the lithosphere from the Earth's interior, the rate of heat production from radioactive elements within the lithosphere, and the rate at which heat is conducted to the Earth's surface (Section 1.6). At mid-ocean ridges, the emplacement of large volumes of basaltic magma, which has risen from the hotter asthenosphere, moves the isotherm much closer to the Earth's surface. As the new hot lithosphere moves away from the ridge, its temperature decreases due to conductive cooling and the circulation of cold ocean water. The circulating water also hydrothermally alters the rocks. The lithosphere thickens, becomes denser, and sinks deeper in the asthenosphere, thus causing oceans to deepen away from ridges (Stein and Stein, 1992). At subduction zones, the sinking of cool lithospheric plates into the asthenosphere depresses isotherms to considerable depths, and as a result, the ocean floor rocks are converted into low-temperature high-pressure metamorphic rocks. The metamorphism releases water and carbon dioxide, which rise into the overlying mantle wedge where they (especially water) behave as a flux that causes fusion and formation of basaltic and andesitic magma. Because these magmas are less dense than the surrounding mantle, they buoyantly rise at least to the base of the continental crust. Here, dense basaltic magma may intrude or underplate the crust. As it cools, it can melt the overlying rocks to form bodies of granitic magma that rise through the crust to form batholiths and possibly erupt on the surface as rhyolite. Andesitic magma may continue all the way to the Earth's surface to form large cone-shaped volcanoes. The rise of all this magma into the base of the continental crust above subduction zones deflects isotherms upward, which, in turn, causes high-temperature–high-pressure metamorphism and possibly melting of the crustal rocks (Marsh, 1979; Annen *et al*., 2006). If basaltic rocks at the base of the crust are metamorphosed into denser mineral assemblages, slabs of this rock are thought to be able to detach (delaminate) and sink into the mantle (Kay and Kay, 1993). The asthenosphere that rises to replace the sunken slab is hot, and consequently the isotherms above a zone of delamination are raised. Melting may occur in the ascending asthenosphere or in the sinking slab, which can produce potassium-rich basalts. The rise of the isotherms and the intrusion of magma can cause high-temperature–low-pressure metamorphism above zones of delamination. Well away from zones of subduction or delamination beneath the stable continent, the 500 °C isotherm returns to its undisturbed steady-state depth of approximately 40 km.

The variation in the depths of isotherms throughout the lithosphere gives rise to a wide range of possible pressure–temperature conditions. These, in turn, control the processes by which rocks are generated. We therefore turn our attention to the problems of how we determine pressure and temperature within the lithosphere.

1.3 PRESSURE DISTRIBUTION WITHIN THE EARTH

Rocks under high pressure do not have high shear strength, especially over long periods of time. Instead, they flow as if they were extremely viscous liquids. The depth at which this occurs is referred to as the *brittle–ductile transition*. In many parts of the Earth, this transition occurs at a depth of ~15 km, and this is why many earthquakes are restricted to depths shallower than this. Earthquakes do occur at greater depths, as for example in subduction zones, but here plate motion has taken relatively cold crustal rocks to depth, and before the rocks have time to heat, they can undergo brittle failure. Metamorphic reactions that liberate fluids can also build up fluid pressures rapidly enough to exceed the tensile strength of low permeability rocks and cause fracturing in the middle crust (Ague *et al*., 1998). Strain associated with the intrusion of magma can cause fracturing, which has been recorded seismically to depths of 60 km. Despite this fracturing, the

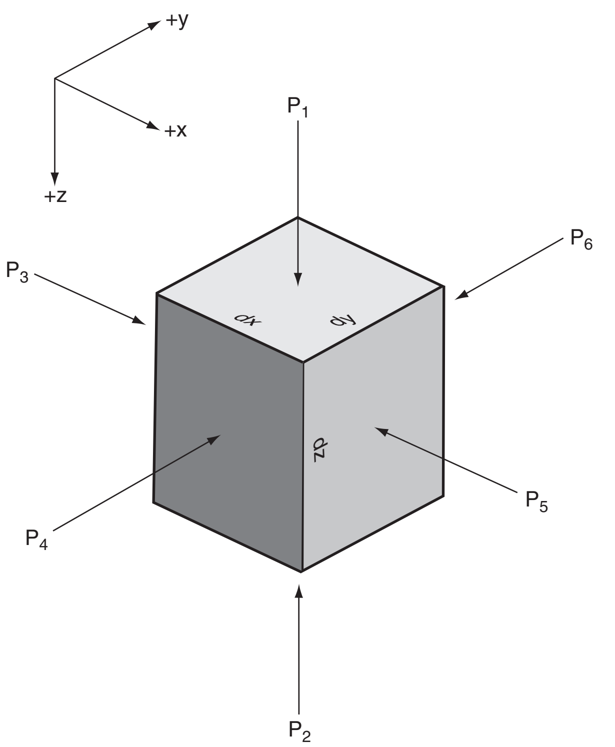

Fig. 1.3 Pressures exerted on the faces of a small volume ($dx\, dy\, dz$) of rock as a result of the load of the surrounding rocks. Note that the positive direction of z is downward (depth).

tendency of rocks to flow in response to long-term stresses allows us to calculate the pressure for any particular depth in the Earth in the same manner that hydrostatic pressure can be calculated for any depth in water. Indeed, the term *hydrostatic* is often used loosely instead of the correct term *lithostatic* to describe pressures in the Earth resulting from the load of overlying rock. At shallow depths, where open fractures exist or where the rock has high permeability, the pressure on the rock may actually be higher than that on the water in fractures or pores; that is, the lithostatic pressure may be greater than the hydrostatic pressure. In contrast, water exsolving from crystallizing magma or generated by metamorphic reactions may cause the hydrostatic pressure to exceed the lithostatic pressure and possibly cause explosions or fracturing and veining.

We will now derive the simple relation between depth and lithostatic pressure in the Earth, assuming that rocks are capable of flowing under high pressures. Although this derivation may appear trivial, it is included here to introduce a particular way of tackling a problem. This same approach will be encountered in later chapters dealing with more complicated problems involving fluid flow, heat transfer, and diffusion.

Consider the forces acting on a small volume of rock at some depth in the Earth. For convenience this volume is given the dimensions dx, dy, and dz (Fig. 1.3), where x and y are perpendicular directions in a horizontal plane, and z is vertical with positive values measured downward (i.e. z = depth). The surrounding rock exerts pressures P_1 to P_6 on each face of the volume. We further stipulate that the pressure at the center of this volume is P, and that there is a vertical pressure gradient, dP/dz.

If the volume does not move horizontally, the forces resulting from the pressures P_3, P_4, P_5, and P_6 must balance one another and need not be considered further. The forces acting in a vertical direction result from the pressures P_1 and P_2 and the acceleration of gravity acting on the mass of the small volume (remember that according to Newton's second law, a force is measured by the mass multiplied by the acceleration that the force produces; $F = ma$). Again, if the volume does not move, the upward and downward forces must balance, or as is commonly expressed mathematically, the forces acting in any particular direction must sum to zero. Recalling that the force resulting from a pressure is simply the pressure times the area on which it acts, we can sum the forces acting downward (positive direction of z) on the volume as follows:

$$P_1\, dx\, dy + (-P_2\, dx\, dy) + mg = 0$$

where m is the mass of the small volume and g is the acceleration of gravity at the depth considered. Note that the force resulting from P_2 is negative because it operates in an upward direction.

The mass of the small volume is obviously dependent on its size. We can, however, express mass in terms of the dimensions of the volume and its density (ρ), which allows the force of gravity to be expressed as follows:

$$mg = \rho\, dx\, dy\, dz\, g$$

The two pressures P_1 and P_2 can be related to the pressure P at the center of the volume by the pressure gradient (dP/dz) within the volume. Because the distance from the center of the volume to the top or bottom is $^1/_2 dz$, pressures P_1 and P_2 are given by

$$P_1 = P - \frac{1}{2}dz\left(\frac{dP}{dz}\right) \quad \text{and} \quad P_2 = P + \frac{1}{2}dz\left(\frac{dP}{dz}\right)$$

We can now write the sum of the forces acting downward on the volume as

$$\left[P - \frac{1}{2}dz\left(\frac{dP}{dz}\right)\right]dx\,dy - \left[P + \frac{1}{2}dz\left(\frac{dP}{dz}\right)\right]dx\,dy$$
$$+\rho\, dx\, dy\, dz\, g = 0$$

which simplifies to

$$\frac{dP}{dz} = \rho g \tag{1.1}$$

This indicates that the pressure gradient at any particular depth is simply the product of the density of the material and the acceleration of gravity at that point. To obtain the pressure at that depth we have simply to integrate the expression

$$\int_{P_0}^{P_z} dP = \int_0^z \rho g\, dz$$

from the surface of the Earth to the depth, z, of interest. The pressure at the Earth's surface of 1 atm is so small compared with those at depth that it can be taken as zero. We can then write

$$P_z = \int_0^z \rho g \, dz$$

To integrate the right-hand side of this equation, we must first determine whether ρ and g are constants or functions of depth (z). Most minerals and magmas are rather incompressible. Thus, no serious error is introduced by assuming that their densities remain constant, at least under the pressures encountered in the crust and upper mantle. However, there are large amounts of data on compressibilities of minerals and magmas (e.g. Bass, 1995), so density can be expressed as a function of pressure, if so desired (see Problem 3.1 in Chapter 3). Here, we will take the density to be constant.

Variation in the value of g with depth is more complex. At points above the surface of the Earth, its value varies inversely as the square of the distance from the center of the Earth. Within the Earth, however, only the underlying mass contributes any net gravitational force; those forces due to the overlying shell of rock sum to zero and hence produce no net gravitational attraction. For example, if the planet were hollow, as postulated by Abraham Gottlob Werner in his theory on the origin of Primary and Secondary rocks, you would have been weightless if you could have entered its interior. With increasing depth in the Earth, less and less mass remains to cause attraction. But this is partly offset by the increasing density of rocks with depth. The result is that although the value of g goes to zero at the Earth's center, throughout the crust and upper mantle it remains approximately constant, and will be taken as such for our purposes.

Taking the density of material and the acceleration of gravity as constants, we integrate the preceding equation to obtain

$$P = \rho g z \tag{1.2}$$

To illustrate this equation, let us determine the pressure at the base of a 35-km-thick granitic crust with an average density of 2800 kg m^{-3}. Substituting these values into Eq. (1.2) and being careful to list units, we obtain

$$P = 2800 \times 9.80 \times 35 \times 10^3 \quad \text{kg m}^{-3} \times \text{m s}^{-2} \times \text{m}$$

But a kg m s^{-2} is a unit of force, the *newton* (N). The units therefore reduce to N m^{-2}, which is the unit of pressure known as a *pascal* (Pa). The pressure at the base of a 35-km-thick crust is therefore 0.96×10^9 Pa or 0.96 GPa. In cgs units this would be 9.6 kilobars:

$$1 \text{ bar} = 10^6 \text{ dyne cm}^{-2} = 0.9869 \text{ atmosphere}$$

The pressure at the base of the 35-km-thick crust in this simple calculation is approximately 1 GPa. By taking into account details of the density distribution within the crust, a more accurate determination of pressure can, of course, be obtained (see Problem 1.1). Also, by making reasonable assumptions about the mineralogy of the deep Earth's interior and using ultrahigh-pressure experimental data on the compressibility of minerals (Bass, 1995) the density of the Earth's mantle and core can be deduced, from which we calculate the pressure (Fig. 1.1). At the base of the upper mantle (660 km), the pressure is 24 GPa (Masters and Shearer, 1995). At the core–mantle boundary it is 136 GPa, but at the inner–outer core boundary it has risen to 329 GPa and at the center of the Earth, it is 364 GPa.

1.4 TEMPERATURE GRADIENTS AND HEAT FLOW IN THE LITHOSPHERE

Temperatures in the Earth cannot be determined as easily as pressures. Deep drill holes probe only the top few kilometers of the crust, and extrapolating temperatures measured in these holes to lower parts of the crust and the upper mantle is fraught with difficulties. It is not surprising to find that the literature contains numerous, significantly different estimates of the geothermal gradient. Much interest in the internal temperatures of the Earth was piqued in the nineteenth century by attempts, such as that of Lord Kelvin (Thomson, 1863), to calculate the age of the Earth based on the length of time to cool from an initially molten state. The age was wrong, because his model did not take into account convection in the mantle (England *et al.*, 2007). These early workers were unaware of a major source of heat in the Earth, that produced by radioactive decay. Today, radiogenic heat is certainly taken into account; nonetheless, markedly different gradients can still be calculated, depending on the assumed distribution of the heat-generating elements (Problem 1.5) and the amount of heat thought to be passing through the surface of the Earth. Fortunately, high-pressure experimental investigations of the melting of rocks provide rather tight constraints on the near-surface geothermal gradient, and for temperatures in the deep mantle and core, they provide the only constraints (Boehler, 2000).

Because the boundary between the inner and outer core is interpreted to be the transition between solid and liquid iron diluted with nickel and several lighter elements, high-pressure melting experiments can be used to set a temperature for this boundary. Achieving these temperatures at the pressure of 329 GPa, however, is difficult, and at present can be reached only in shock experiments where a projectile is fired in a sophisticated two-stage gas gun at the experimental charge (Brown and McQueen, 1986). Results of lower-pressure static experiments using diamond anvils can be extrapolated to the higher pressures. Considerable uncertainty surrounds all of this experimental work and its extrapolation to conditions at the boundary of the inner core. In addition, little is known of how the melting point of iron is affected by the addition of other constituents (see Boehler, 2000, for summary). With all these uncertainties, it is only possible to say that the temperature at the boundary of the inner core is around 5000 K (Fig. 1.1).

Temperature gradients through the liquid outer core must be controlled largely by convection, and in the limiting case, would follow what is known as an *adiabat*; that is, no heat would be lost or gained from the convecting liquid as it rose toward the core–mantle boundary, but its temperature would decrease as the liquid decompressed (see adiabat in Chapter 7.5 and Problem 7.7). The fact that the outer core is still liquid at the core–mantle boundary indicates that the temperature must be above the melting point of iron (plus other components) at this depth. High-pressure experiments indicate that at the 136 GPa pressure of this boundary, the temperature must be above ~4000 K (Boehler, 2000).

Seismic data indicate that the lower mantle is solid. Temperatures throughout the lower mantle must therefore be below the melting point, except perhaps at the base of D″ where the ultralow-velocity zone indicates that partial melting probably occurs near the core–mantle boundary. Experiments and theoretical calculations show that the extrapolated melting point of pure $MgSiO_3$ with a perovskite structure, one of the major components of the lower mantle, is 5400 ± 600 K at 136 GPa, the pressure at the core–mantle-boundary (Stixrude and Karki, 2005). The addition of other components would lower this melting point by as much as 1300 K. Thus, the temperature at the base of the mantle must be below (or near, if melt is present) 3500 to 5200 K, which is consistent with the ~4000 K estimated for molten iron in the outer core (Boehler, 2000).

Although the mantle is solid, calculations show that it must convect, and therefore it will have a nearly adiabatic temperature gradient of between 0.25 and 0.3 K km^{-1} (see Problem 7.7 and 7.8). This gradient can be pinned by experimental work to 1900 ± 100 K at a depth of 660 km, where the olivine composition switches from a spinel to a perovskite structure. An adiabatic gradient of 0.3 K km^{-1} would therefore give a temperature of ~2500 K at a depth of 2600 km. Below this depth, the D″ zone is a thermal boundary layer through which the temperature must rise rapidly by ~1500 K to reach the 4000 K of the outer core.

At shallower depths in the mantle, the geothermal gradient can be extrapolated from measurements near the surface of the Earth and from knowledge of the experimentally determined melting behavior of rocks that are known to exist at these depths. Measurements in deep mines and drill holes indicate that the near-surface geothermal gradient, dT/dz, ranges from 10 to 60 °C km^{-1}, with a typical value in nonorogenic regions being near 25 °C km^{-1} (Fig. 1.9). If this gradient continued to depth, the temperature would be 625 °C at 25 km (0.65 GPa). Experiments indicate that in the presence of water, crustal rocks melt to form granitic magma under these conditions. The transmission of seismic shear waves through this part of the crust indicates that melting at this depth is neither a common nor a widespread phenomenon. Similarly, at a depth of 40 km (1.2 GPa), a geothermal gradient of 25 °C km^{-1} would reach a temperature of 1000 °C, and peridotite, the rock constituting the mantle at this depth, would begin to melt in the presence of excess water. By 52 km (1.6 GPa) the temperature would be 1300 °C, which exceeds the beginning of melting of even dry peridotite. Again, seismic data do not indicate large-scale melting at this depth. Clearly, the near-surface geothermal gradient must decrease with depth. The question, then, is why should the gradient decrease, and can we calculate or predict its change?

A combination of petrological and geophysical observations allows us to place limits on the temperature of several points along the geotherm. Seismic velocities decrease abruptly by about 10% at a depth of 60 to 100 km, the upper boundary of the low-velocity zone. This is probably the contact between solid peridotite above and peridotite that has undergone a small percentage of melting below. A thin film of melt along grain boundaries can greatly reduce the rigidity of a rock and strongly attenuate seismic energy. Experiments show that water-saturated peridotite begins to melt at about 1000 °C at a pressure of 2 GPa, which corresponds to a depth of 70 km (Fig. 1.9). The top of the low-velocity zone is the boundary between the lithosphere and the asthenosphere. The base of the low-velocity zone, at 250 km, is marked by a small increase in the rate of increase of velocity with depth. This is interpreted as the depth at which the P–T curve for the beginning of melting of mantle peridotite recrosses the geotherm, so that at depths greater than 250 km, the mantle is solid. This places an upper limit of about 1500 °C at 250 km.

Common experience teaches us that heat flows from regions of high temperature to ones of low temperature, and that it is transferred in several ways depending on the nature of the medium through which it is transmitted. For example, in a vacuum heat can be transferred only by radiation; in a gas, liquid, or even plastic solid, such as the mantle, it may be transferred by convection; and in rigid, opaque solids, it can be transferred only by conduction. These mechanisms are discussed in detail in Chapter 5. Heat transfer through the lithosphere, however, especially in old, stable continental and oceanic regions, is almost entirely by conduction.

Experience also shows that the amount of heat transferred by conduction is proportional to the negative temperature gradient (−dT/dx); that is, the greater the temperature decrease in a particular direction x, the greater will be the amount of heat transferred in that direction. This can be expressed mathematically by introducing the quantity known as *heat flux*, J_{Qx}, which is the quantity of thermal energy passing in the x direction through a unit cross-sectional area in a unit of time. We can write this as

$$J_{Qx} = -K\left(\frac{\mathrm{d}T}{\mathrm{d}x}\right) \tag{1.3}$$

where K, the constant of proportionality, is known as the *coefficient of thermal conductivity*. In the case of the Earth, heat flows out because of the geothermal gradient. This gradient (dT/dz) is normally recorded as a positive value with z increasing downward. In the upward direction, however, this gradient is negative, so that a positive quantity of heat is transferred upward and out of the Earth.

Direct measurement of heat flux in the field is not practical, but temperature gradients can be measured easily in boreholes sunk into sediments on the ocean floor and in deep drill holes on continents. These gradients can then be used, along with laboratory measurements of the thermal conductivity of the penetrated material, to calculate the heat flux according to Eq. (1.3). For example, with a typical thermal conductivity coefficient of 2.0 J m^{-1} s^{-1} °C^{-1}, a gradient of 25 °C km^{-1} results in a heat flux of 50.0×10^{-3} J m^{-2} s^{-1}; but a joule per second is a watt, so the heat flux is 50.0 mW m^{-2}. For comparison, a 60 watt light bulb emits 60 joules per second from a surface area of approximately 11×10^{-3} m^2. Its energy flux is therefore 5.5 kW m^{-2}, that is, one-hundred-thousand times more than the heat flux from the Earth. The Earth's heat flux is therefore a very small quantity, and even where the geothermal gradient is 60 °C km^{-1}, which can be found near mid-ocean ridges, as for example in Iceland where it powers geothermal power plants (Fig. 1.4), the heat flux is still only 120 mW m^{-2}. Over very active spreading axes, such as the East Pacific Rise, the flux may be as high as ~300 mW m^{-2} (Fig. 1.5). In cgs units, heat flow is measured in μcal cm^{-2} s^{-1}, which is commonly referred to as 1 *heat flow unit* or 1 HFU; 1 HFU = 41.84 mW m^{-2}.

Fig. 1.4 Geothermal power plant near the Hengill volcano in southern Iceland. The valley in which the plant is situated is part of the Mid-Atlantic Ridge system passing through Iceland. This plant, which supplies Iceland's capital, Reykjavik, with much of its heating and electrical power, taps a high-temperature geothermal field where temperatures exceed 200 °C at a depth of 1 km. Pipes from many geothermal wells can be seen leading to the power plant. Geothermal power plants supply Iceland with ~17% of its electrical power, and ~90% of its central heating.

The International Heat Flow Commission's synthesis of worldwide heat flow data is available on the Web (www.geo.lsa.umich.edu/IHFC) and is reproduced here as Figure 1.5 (Pollack *et al.*, 1993). This synthesis reveals that the new ocean floor created at divergent plate boundaries plays an important role in the loss of heat from the planet. Indeed, it is estimated that 50% of all heat lost from the Earth is liberated from these regions, which constitute only 30% of the surface of the planet. The heat flux from ancient ocean floor has a relatively constant value of 38 mW m^{-2}. As the ocean floor gets younger toward mid-ocean ridges, the heat flux steadily increases, reaching a mean value of 250 mW m^{-2} in ocean floor younger than 4 Ma, but there is considerable scatter in the younger data. The scatter has two main causes. Old ocean floor has had time to accumulate thick layers of sediment, which seals the underlying fractured igneous rocks from circulating ocean water and provides a layer into which drilling probes can be sunk easily to obtain a conductive geothermal gradient. On new ocean floor, the lack of sediment makes it difficult to drill holes in which to measure the

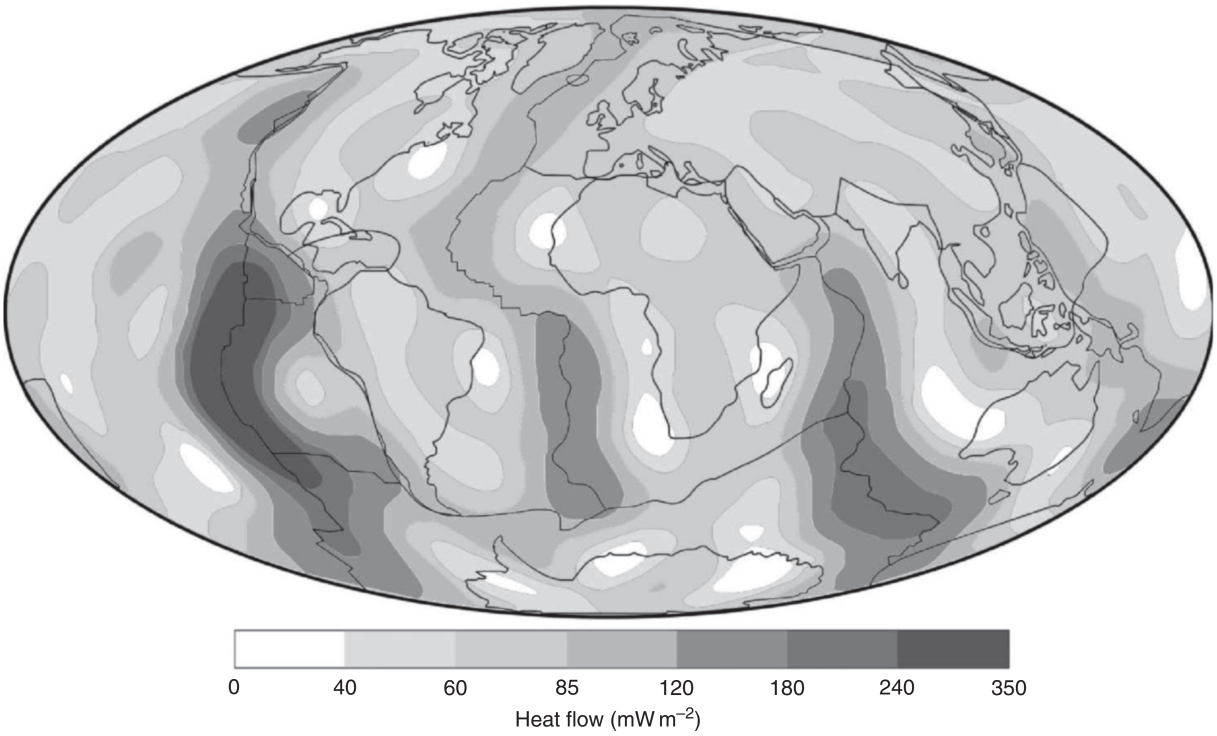

Fig. 1.5 Global heat flow. Data are shown as a degree 12 spherical harmonic representation (after Pollack *et al.*, 1993).

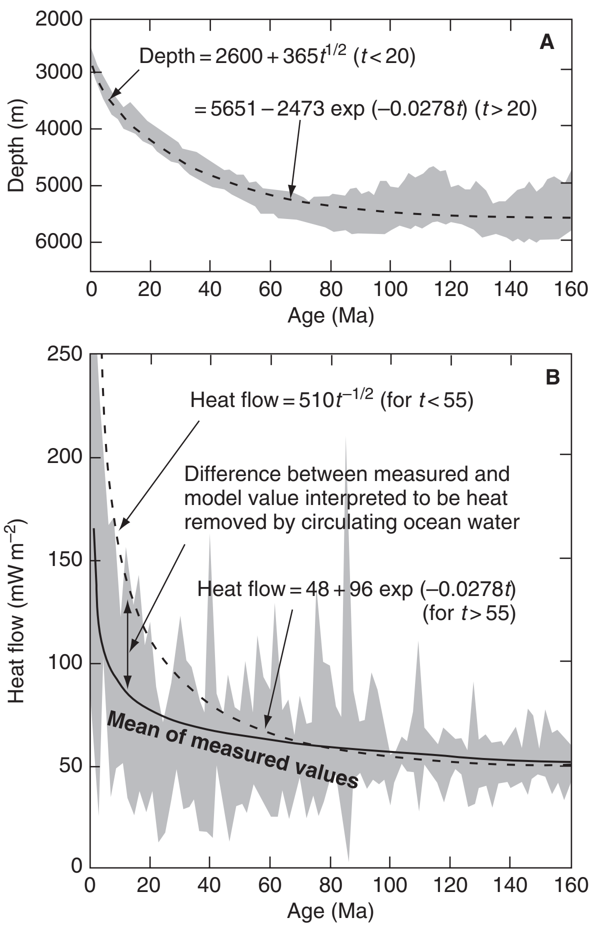

Fig. 1.6 (**A**) Shaded area shows variation in mean ocean depth (± 1 standard deviation) of the North Pacific and Northwest Atlantic as a function of ocean floor age. The dashed line shows mathematical fits to the data (see text for discussion). (**B**) Shaded area shows variation in mean heat flow (± 1 standard deviation) from the same ocean floor as in (A). The solid line is the mean of the measured data, whereas the dashed line is a model based on the cooling of the lithosphere as it moves away from an ocean spreading axis. The difference between the model and measured values is interpreted to be heat removed by circulating ocean water. (Drawn from figures presented in Stein and Stein, 1992, and Hofmeister and Criss, 2005.)

temperature gradient. In addition, without sediment sealing fractures, ocean water is free to circulate through the igneous rocks and remove heat that would otherwise have to be conductively transferred through the rock. Vents have been found along many oceanic ridges emitting high temperature water from which sulfides commonly precipitate to form what are called "black smokers" (Von Damm, 1990). Clearly, a considerable amount of heat is removed from the Earth by these circulating waters, which is in addition to the heat that is transferred conductively through the rock. To understand the Earth's heat budget, which essentially controls the rate of many processes on the dynamic planet, we need to know the magnitude of this hydrothermal component. To see how scientists have tried to answer this question, we must first examine one of the most remarkable features of the planet, the relation between the age and depth of the ocean floor.

One of the most important findings to come from the exploration of the ocean floor is the discovery that the ocean's depth is proportional to the age of the ocean floor; that is, the older the ocean floor, the greater is its depth (Fig. 1.6(A)). Indeed, for ocean floor younger than 20 Ma, its depth is proportional to the square root of its age, with depth being expressed accurately by:

$$\text{Depth} = 2600 + 365 \times t^{1/2} \tag{1.4}$$

where depth is in meters and time in Ma BP (million years before present; Stein and Stein, 1992; Parsons and Sclater, 1977). Even for ocean floor as old as 70 Ma, the relation still holds quite well, but older ocean floor sinks more slowly. The square root of time relation has been interpreted in terms of the cooling of the lithosphere as it moves away from the ocean ridges; the crust becomes denser as it cools, and isostatic readjustment causes the ocean floor to sink deeper with increasing age. Numerous attempts have been made to model the cooling and sinking of lithospheric plates (see Oxburgh, 1980). We will discuss these models in more detail in Chapter 23, but for the moment it is necessary to recognize only that when an object conductively cools by transferring heat across a plane boundary, the heat flux across that boundary typically decreases in proportion to the square root of time (see Eq. (5.13)). Although we may not have been aware of the square root term, we are all familiar with this effect from the cooling of a cup of coffee. Initially, when the coffee is hot, heat is lost rapidly and the coffee cools rapidly, but with time, as the coffee cools, it loses less and less heat and cools more slowly.

If cooling is the correct explanation for the increased depth of the ocean floor with age, then the heat flux through the ocean floor would be expected to be inversely proportional to the square root of the age of the ocean floor (Fig. 1.6(B)). This, in fact, is found to be the case where the ocean floor is covered with a significant thickness of sediment. Stein and Stein (1992) have shown that in these regions the heat flux is well described by

$$J_Q = 510 \times t^{-1/2} \tag{1.5}$$

where J_Q is in mW m^{-2} and t is in Ma BP, as long as t is less than 55 Ma. However, where the ocean floor is young and has little sediment covering it, the heat flux measurements fall farther and farther below the predicted curve as the ocean ridge is approached (Fig. 1.6(B)). This shortfall is interpreted to represent heat that is removed by the circulating ocean water. Stein and Stein (1994) estimate that 34% of the total global oceanic heat flux occurs by hydrothermal circulation (see also papers in Davis and Elderfield, 2004). In the synthesis of worldwide heat-flow data shown in Figure 1.5, the heat-flux values for young ocean floor are not measured values but are extrapolations from measurements on older ocean floor using Eq. (1.5).

Despite the common acceptance of calculating young ocean-floor heat fluxes in this way, other models have been proposed (Hamilton, 2003). Hofmeister and Criss (2005) argue that the heat flux in these regions is actually much

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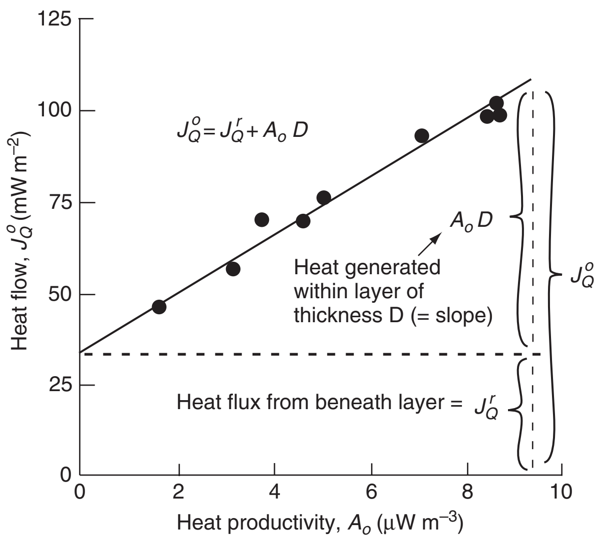

Fig. 1.7 Heat flow plotted as a function of the radiogenic heat production of surface rocks in New England (after Birch *et al.*, 1968). See text for discussion.

closer to the measured values than it is to the extrapolated values and that far less heat is being removed by hydrothermal circulation than predicted by the square root of time cooling model (Fig. 1.6(B)). Depending on which model is used for the heat flux in young ocean floor, huge differences result in the calculated total heat budget for the Earth. For example, according to Pollack *et al*. (1993), the total heat flux from the Earth is 44 TW, whereas Hofmeister and Criss (2005) estimate it to be only 31 TW. The lower estimate agrees well with the amount of heat that would be expected to be generated from radioactive decay if the planet were formed from chondritic meteorites (see Chapter 6). The higher heat flux estimates require additional heat sources, which could be left over from an earlier age, such as planetary accretion and large impacts, short-lived radioactive isotopes, core formation, and crystallization of the outer core.

Heat-flux measurements on continents do not share the problems of those on new ocean floor. As first shown by Polyak and Smirnov (1968), and substantiated by other studies (Sclater *et al*., 1980), heat flux on continents decreases with increasing age, at least back to about 800 Ma. The rate of decrease, however, is much less than that found in the ocean floor and therefore cannot be modeled as simple cooling of newly formed crust. The relation is more complex, possibly involving the removal, by erosion, of heat-generating radioactive elements that are concentrated in the upper parts of young crust. In young orogenic belts, heat fluxes may be as high as 150 mW m^{-2}, but in crust older than 800 Ma, they tend to be about 40 mW m^{-2}, which is similar to the value in ancient oceanic crust.

One of the most important findings to come from the study of heat flow is the relation between the surface heat flux and the concentration of heat-generating radioactive elements in the local rocks (Birch *et al*., 1968; Lachenbruch, 1968; Lachenbruch and Sass, 1977). The heat flux from old eroded plutonic bodies of igneous rock is very nearly linearly related to the local concentration of heat-generating elements, as shown in Figure 1.7. This linear relation can be expressed mathematically as

$$J_Q^0 = J_Q^r + DA_0 \tag{1.6}$$

where J_Q^0 is the surface heat flow (depth = 0), J_Q^r the intercept at $A_0 = 0$, A_0 the local radiogenic heat productivity, and D the slope of the line.

The simplest interpretation of this relation is that the slope of the line indicates the thickness of the heat-generating layer. Thus the heat flux at any locality can be interpreted as consisting of two parts; a constant flux of heat, which is given by the intercept J_Q^r, comes from below the layer; to this is added the heat flux generated by radioactive decay in the layer. This type of relation is known as an *energy balance* or *conservation equation*, for it states that the energy coming out of the top of the layer must equal the energy entering the layer from below plus the energy created within the layer, that is,

$$J_Q\big|_{z=0}^{\mathrm{out}} = J_Q\big|_{z=D}^{\mathrm{in}} + \int_{z=0}^{z=D} A_z \,\mathrm{d}z \tag{1.7}$$

Values of J_Q^r, which are known as the *reduced heat flow*, are characteristic of a given geological province. For example, its value in the eastern United States is ~33 mW m^{-2}, in the Basin and Range Province it is ~59 mW m^{-2}, and in both the Baltic and Canadian Precambrian Shields it is ~22 mW m^{-2}. Values of D, the thickness of the layer containing the radioactive elements, in these same provinces are 7.5, 9.4, 8.5 and 12.4 km respectively (Oxburgh, 1980). The fact that the surface concentration of radiogenic elements cannot extend to depths much greater than 10 km is surprising considering that seismic data indicate that "granitic" rocks at the surface extend to depths of approximately 25 km under most continental areas. Clearly, the radioactive elements must be concentrated near the surface of the Earth. This being so, our model of a layer with a constant concentration, A_0, of radioactive elements throughout its thickness D is unlikely to be valid. The concentration is more likely to decrease downward (see Problem 1.6).

1.5 HEAT SOURCES IN THE EARTH

Heat flows into the base of the crust from the mantle below in response to temperature gradients that were set up early in the Earth's history. This heat has two sources, the release of gravitational potential energy and the decay of naturally occurring radioactive nuclides. Conversion of kinetic to thermal energy during the Earth's accretion, which was largely complete by 4.567 Ga (Jacobsen, 2003), release of gravitational energy on separation of the core very early in Earth's history (Solomon, 1979), energy from a Moon-forming mega impact after the core had formed but before 4.45 Ga (Zhang, 2002), and generation of heat by decay of short-lived radioactive isotopes, such as ^{26}Al, have all contributed to the