

Planetary Heating and Differentiation

We explain how heat is produced by radioactive decay, segregation and exothermic crystallization of metallic cores, impacts, and tidal forces. Planetesimals in the early Solar System were most affected by the decay of short-lived radionuclides. Larger, rocky planets were heated primarily by large impacts and core segregation. Because rocks are poor conductors, heat retention in rocky bodies is a function of planet size. Large-scale melting to produce magma oceans was likely a common process facilitating differentiation to form cores, mantles, and crusts. Metallic liquids are probably necessary for core segregation. Primary crusts, formed during planetary differentiation, are rarely preserved. Mantles are residues from the extraction of silicate crustal melts and core materials. Differentiation of the giant planets was driven by density variations in high-pressure forms of gases, ices, and rock more than by heating and melting. The importance of the various planetary heat sources changes over time; in modern planets the effective heat sources are decay of long-lived radioisotopes and, for the Earth, exothermal crystallization of the liquid outer core.

6.1 Too Hot to Handle

The geologic evolution of planetesimals and planets is fueled by heat from various sources. Rocks are notoriously poor conductors of heat, so heat generated by processes in the deep interiors of large rocky planets can be retained over geologic timescales. Besides conduction, other mechanisms of heat loss such as convection can cause molten bodies to cool more rapidly. On the other hand, small bodies lose heat more rapidly than large planets, because their smaller rock masses allow heat to be conducted outward more efficiently. Once the heat reaches the surface, it is radiated into space, and the greater ratio of surface area to volume

for small bodies promotes more rapid cooling. This is illustrated in Figure 6.1, a diagram showing the apparent duration of igneous activity on various rocky bodies as a function of body size. For small bodies like asteroid Vesta, igneous activity lasted for only a few million years. Igneous activity ceased on the Moon after several billion years, and still persists today on large planets like the Earth.

The giant planets, Jupiter, Saturn, Uranus, and Neptune, are warmer than can be explained by solar radiation alone, so they too must have internal heat sources. These sources are roughly comparable in magnitude to the energy they receive from the Sun.

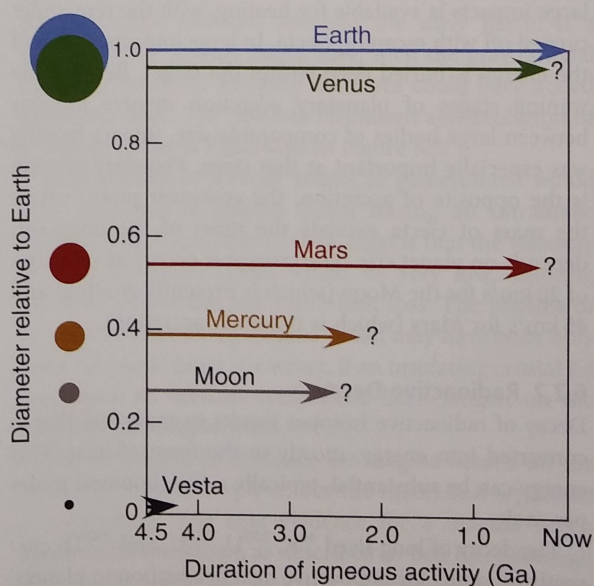


Figure 6.1 Relationship between planetary size and the duration of igneous activity, demonstrating the effect of cooling in smaller bodies.

6.2 Heat Sources

Heat is generated by varying combinations of processes, some endogenic and some exogenic. Regardless of the source, the added energy is converted to atomic motions and thence to heat.

6.2.1 Accretion and Impacts

Impacts provide a planet with energy as well as mass, and can be a potent heat source. This is a particularly important heat source during planetary accretion, but later impacts also cause heating. The energy (E) comes from the encounter velocity (V) and the gravitational energy gained as the body falls to the planet's surface at roughly the **escape velocity** (GM/R):

$$E = V^2/2 + GM/R \quad (6.1)$$

where G is the gravitational constant, and M and R refer to the target planet's mass and radius. Note that impact velocity is squared, so this is a particularly important factor. This energy can be converted into a temperature change using the expression:

$$\Delta T = E/(MC_p) \quad (6.2)$$

where C_p is the heat capacity. Accretion of small bodies involves modest velocities (as well as modest masses) and does not generate nearly as much energy as large impacts. Also, impacts provide only limited heat to planetesimals, because the GM/R term refers to the small target bodies.

Approximately 70 percent of the kinetic energy from large impacts is available for heating, with the remainder carried off with escaping ejecta. In large impacts, much of the energy is buried deep within the target. Because the waning stages of planetary accretion involve impacts between large bodies of comparable size, impact heating was especially important at that stage. Planetary erosion is the opposite of accretion; the crossover point, where the mass of ejecta exceeds the mass of the impactor, depends on planet size. The crossover occurs at velocities of 20 km/s for the Moon (which is presently eroding) and 45 km/s for Mars (which is presently accreting).

6.2.2 Radioactive Decay

Decay of radioactive isotopes results in mass loss that is converted into energy, mostly in the form of heat. This energy can be substantial, typically a few thousand joules per mole.

The decay of long-lived ^{40}K , ^{235}U , ^{238}U , and ^{232}Th currently accounts for radioactive heat production in planets. The heat generated depends on the decay rates of the radioisotopes and their abundances. As an illustration, ^{40}K and ^{87}Rb have comparable half-lives, but ^{40}K accounts

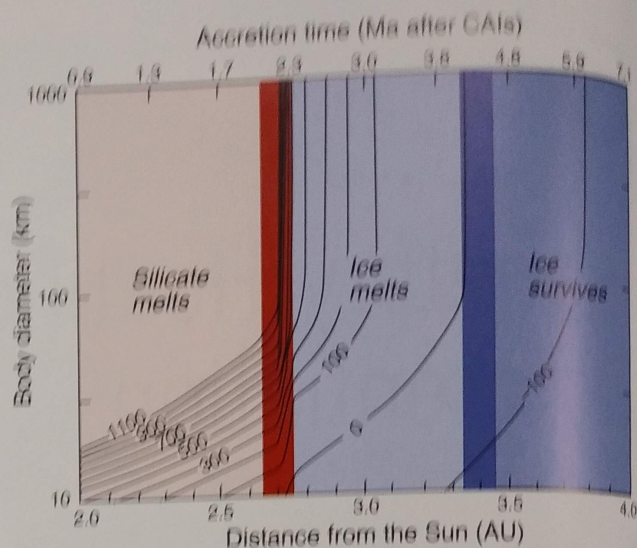


Figure 6.2 Model for asteroids heated by decay of ^{26}Al , assuming accretion starts closer to the Sun and progresses outward. Contours show the maximum temperatures reached within bodies of various sizes, and the vertical red and blue bars separate regions where rock and ice melt. After Grimm and McSween (1993).

for 15 percent of the heat generated in the Earth's crust, whereas the contribution from ^{87}Rb is insignificant because of its relatively low abundance. Long-lived radioisotopes are not particularly important in heating small planetesimals, because the heat can diffuse out faster than it is generated by radioactive decay.

Despite its low abundance, the short-lived radioisotope ^{26}Al decayed very fast and thus caused significant heating within the first few million years of Solar System history. One estimate of the heating effect of ^{26}Al decay in planetesimals is shown in Figure 6.2. In this model, accretion is assumed to have occurred first nearer the Sun, where the density of gas and dust was greater, and to have swept outward to greater heliocentric distance with time. This accretion model is certainly a simplification, but it may explain the pattern of heating inferred from asteroid spectroscopy (discussed in Section 5.5). Bodies that accreted earlier would have had a greater amount of live ^{26}Al , which continued to decay as the accretion front moved outward. Thus, bodies of 100 km diameter nearer the Sun, falling within the "silicate melts" field of this diagram, were melted. Bodies that were smaller or accreted later were still heated and thermally metamorphosed, but their interiors never got hot enough to melt. In bodies accreted even later in the middle of the asteroid belt and beyond the snow line, temperatures were high enough to melt ice and cause aqueous alteration. At even greater heliocentric distance, bodies accreted after ^{26}Al had mostly decayed, so even ice survived.

6.2.3 Core Segregation and Core Crystallization

In the early Earth, core segregation caused heating, as gravitational potential energy (that is, massive metal distributed away from the planet's center of mass) was converted to thermal energy as metal fell to the planet's center.

The energy necessary to melt a solid includes that required to raise its temperature to the melting point, plus more energy to convert it from solid to liquid (the latter is called the latent heat of melting). Conversely, crystallization of a melt releases latent heat, so crystallization is exothermic. As the Earth has cooled over billions of years, its liquid outer core has been crystallizing, adding to the volume of the solid inner core. This crystallization releases heat that drives convection in the molten outer core and heats the mantle from below.

6.2.4 Tidal Forces

Tides extend beyond distorting water – the familiar ocean tides. **Tidal forces** can also distort solid bodies. Distortion occurs because the gravitational attraction on a revolving body is stronger on one side than on the other. If the revolving body is close enough to the more massive object, the revolving body responds by deforming into a football shape, analogous to the way the oceans deform on Earth.

If the body's orbit is eccentric (to maintain eccentricity over time, a third body in orbital resonance is needed), the gravitational field changes during the orbit, and this produces flexing, as the gravity differential changes along the orbit. This flexing produces heating by shear friction. Tidal heating is the cause of volcanic eruptions on Jupiter's moon Io, where the present-day heat flow is ~200 times that expected for radioactive decay. Internal heating by tidal forces also accounts for thermal anomalies and erupting ice fountains on Saturn's moon Enceladus.

6.3 Magma Oceanography

When planetesimals formed in the early Solar System, the decay of short-lived radionuclides was the most important heat source. Calculations suggest that small bodies that accreted within the first million years after CAIs should have melted completely.

For larger planets like the early Earth, the kinetic energy delivered by large impacts was the most potent heat source, and the decay of short-lived radionuclides played a significant but lesser role because of the time lapse before planet formation. The terrestrial planets may also have experienced large-scale melting to form **magma oceans** (Elkins-Tanton, 2012). Because these collisions

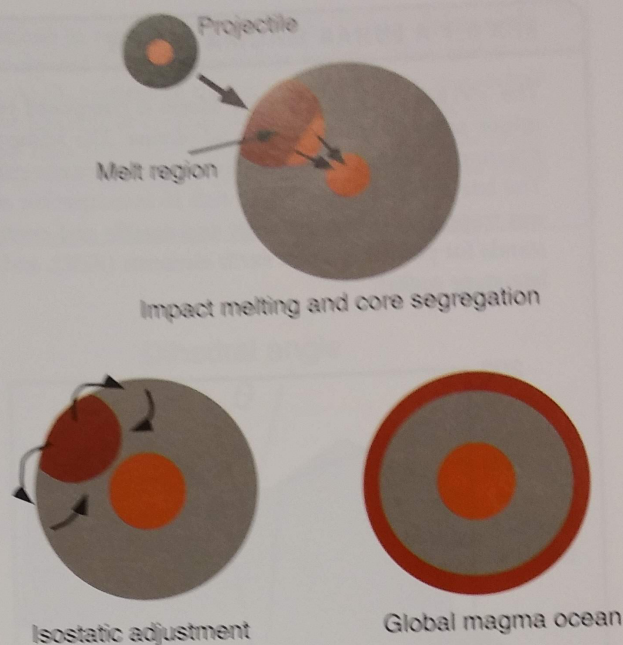


Figure 6.3 Origin and evolution of a magma ocean formed as a consequence of a giant impact. A hemispherical magma ocean forms, and the iron core of the projectile merges with the target body core. Isostatic adjustment causes the magma ocean to form a globally distributed layer.

occurred over some time interval, the planets may actually have had several, transient magma oceans. Thus, the Moon-forming impact (Section 5.6) may have been only the last major collision affecting the early Earth. If a colliding body already had a core, then the gravitational potential energy of merging the cores could have added additional heat. The possible formation and evolution of a magma ocean is illustrated in Figure 6.3.

Bulk melting of a rocky planet or planetesimal would have produced a magma ocean having an ultramafic composition. Experimental data suggest that the viscosity of such a magma ocean would be very low, allowing vigorous convection and rapid heat loss. The lifetime of the Earth's convecting magma ocean may have been only a few thousand years. However, if an insulating crustal lid developed, its lifetime could have been longer, on the order of 100 million years.

The observational evidence for magma oceans on the terrestrial planets is sparse, but the theoretical argument is compelling. The best evidence for a magma ocean comes from the Moon, as described in Box 6.1. Although lunar rocks reveal the structure of the solidified magma ocean, they are not applicable to understanding the crystallization of magma oceans on the terrestrial planets. The low pressures within the Moon allowed feldspar to

BOX 6.1 A LUNAR MAGMA OCEAN

The ~50 km thick crust of the Moon is composed predominantly of plagioclase, forming anorthosites, with minor amounts of pyroxene and olivine. The feldspathic crust is thought to have formed by plagioclase flotation in a magma ocean; olivine and pyroxene either crystallized at the bottom or sank to form the mantle. The late-stage residual liquid, rich in incompatible elements that were excluded from fractionating minerals, was trapped between the crust and mantle and crystallized to form a selva of KREEP rocks (the acronym stands for potassium, rare earth elements (REE), and phosphorus – all incompatible and thus concentrated in late-stage melts).

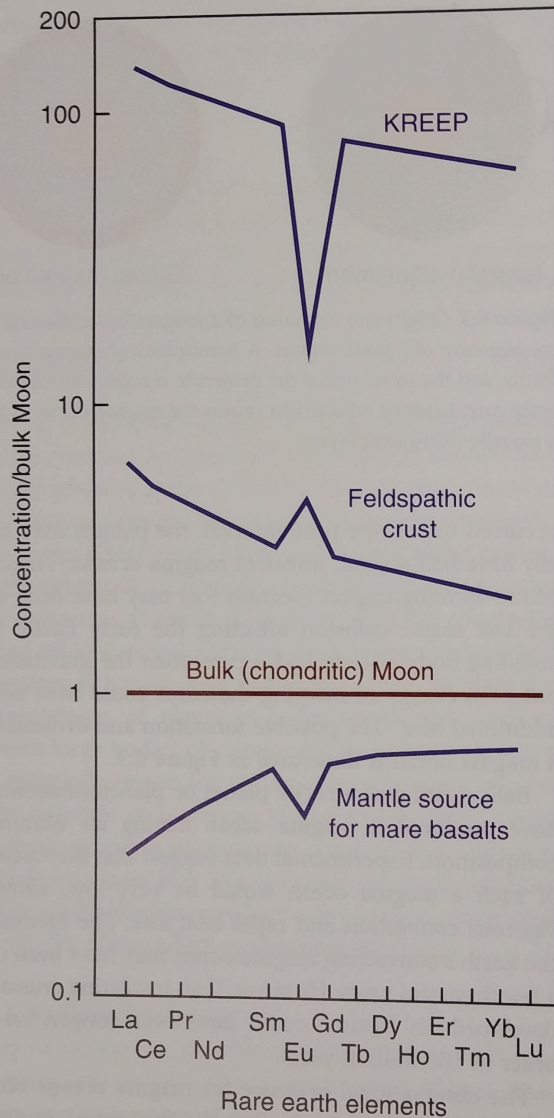


Figure 6.4 Rare earth element patterns, normalized to the bulk Moon composition, provide evidence of a magma ocean. The europium (Eu) anomalies indicate accumulation or depletion of plagioclase, which floated to form the lunar crust. KREEP, the residual liquid from the magma ocean, is extremely enriched in incompatible REEs.

Rare earth element patterns of lunar rocks (Figure 6.4) provide geochemical support for this model. These elements are generally similar in size and charge (trivalent), and thus travel together during melting and crystallization. An exception is europium (Eu), which is divalent in the highly reduced Moon, and substitutes readily for calcium in plagioclase. The feldspathic crust has a positive Eu anomaly, reflecting the accumulation of plagioclase in the crust. The mantle has a complementary negative Eu anomaly, which was inherited by subsequent basaltic melts of the mantle that have erupted to form the maria. KREEP rocks, representing the last dregs of the magma ocean, are extremely enriched in REEs.

Crystallization models for a completely melted and half-melted Moon are contrasted in Figure 6.5. The depth of the magma ocean is constrained by the amount of plagioclase that can be formed, which

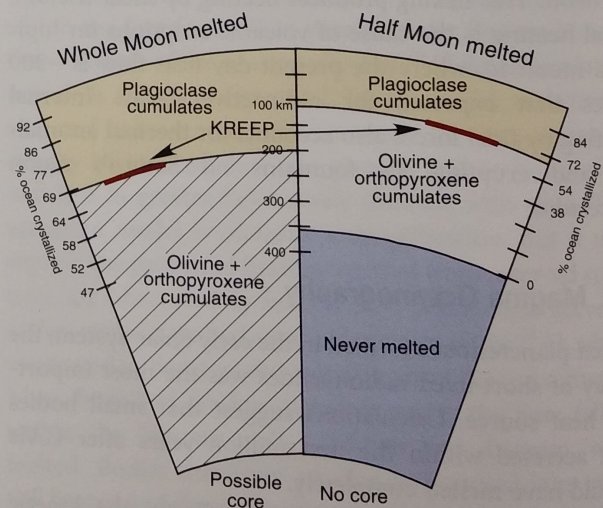


Figure 6.5 Two models for crystallization of the lunar magma ocean, involving different amounts of melting. In both scenarios, cumulus olivine and pyroxene sink and cumulus plagioclase floats. KREEP, the last gasp of melt, is trapped between the crust and mantle cumulates. Modified from Ryder (1991).

in turn depends on the extent to which the Moon is enriched in refractory elements (including aluminum). In both models, olivine and orthopyroxene crystallized early and formed a cumulate mantle. After about 70 percent fractional crystallization, which increased the Al_2O_3 content of the remaining magma, plagioclase began forming and floated upward to form the crust. The amount of crystallization before the onset of plagioclase is shown as tick marks. The late-stage residual melt (KREEP) was sandwiched between the crust and mantle. Later mixing (not illustrated) may have resulted from sinking of the dense KREEP layer and rising of plumes of mantle cumulates.

crystallize and accumulate to form its crust. The corresponding high-pressure aluminous mineral is garnet, which would have remained in the mantle in larger planets with deeper magma oceans.

6.4 Differentiation of Rocky Planets and Planetesimals

Planetary **differentiation** to form cores, mantles, and crusts is arguably the most fundamental geologic process. What we know, or infer, about differentiation is pieced together from samples, experiments, and theory.

6.4.1 Getting to the Heart of the Matter: Cores

The formation of planetary cores occurs because of the large density contrast between silicates and metal. Cores are predominantly iron, because of its high Solar System abundance and density. The separation of solid iron from solid silicate is too sluggish to have been a significant process (Stevenson, 1990). However, iron alloyed with other elements, like oxygen, sulfur or silicon, has a lower melting temperature than mantle silicates. Core formation may thus have involved the movement of molten iron through a solid silicate mantle. In this case, percolation of liquid metal depends on melt connectivity (Rubie et al., 2007). Liquid can “wet” silicate grains if the dihedral angle (θ) between solid-liquid boundaries connected at a triple junction is less than $\sim 60^\circ$ (Figure 6.6). In this case, the liquid is fully interconnected and can migrate through the solid. If $\theta > 60^\circ$, the liquid forms isolated pockets and cannot segregate unless the melt fraction constitutes at least several volume percent. At high pressures, appropriate for the mantles of the terrestrial planets, dihedral angles exceed 60° and some metal remains stranded in the mantle. The dihedral angle is less relevant if shear occurs, which tends to favor percolation.

In the case of a magma ocean, the molten core of an impacting body would emulsify into small droplets that “rain” through the melt. Liquid iron would pond at the base of the magma ocean. From this point on, large blobs

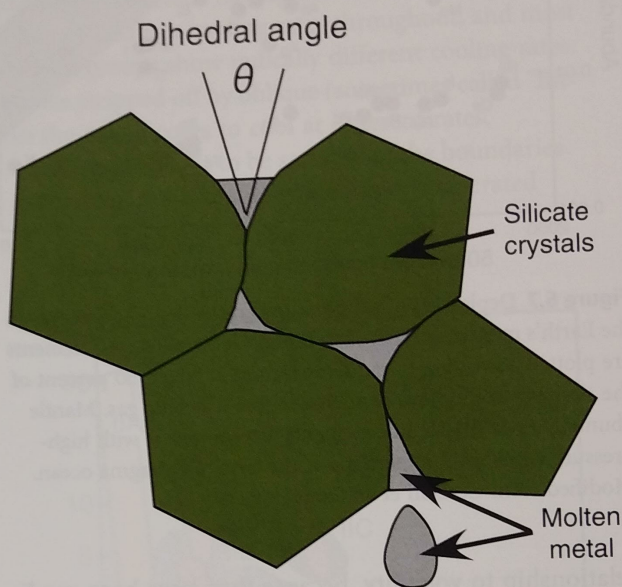


Figure 6.6 Relation between melt conductivity and dihedral angle in an aggregate of crystals containing dispersed metallic melt.

of molten metal would sink rapidly through the solid silicate mesh or intrude downward as dikes.

Small metal droplets can chemically equilibrate with the enclosing magma ocean, providing a test for this mechanism. Conversely, large descending blobs of liquid metal would probably preclude equilibration with the enclosing silicates. Siderophile element abundances in mantle rocks can be used to test for equilibration. Compared to the abundances of lithophile elements, siderophile and chalcophile elements in the Earth's mantle are strongly depleted (Figure 6.7). We have seen diagrams like this before (look back at Figures 4.14 and 5.11), where elements are plotted according to the temperature at which 50 percent of the element would have condensed. The refractory lithophile elements have constant (chondritic) relative abundances, whereas the abundances of volatile lithophile elements decrease with volatility. On the other hand, the siderophile element abundances show no

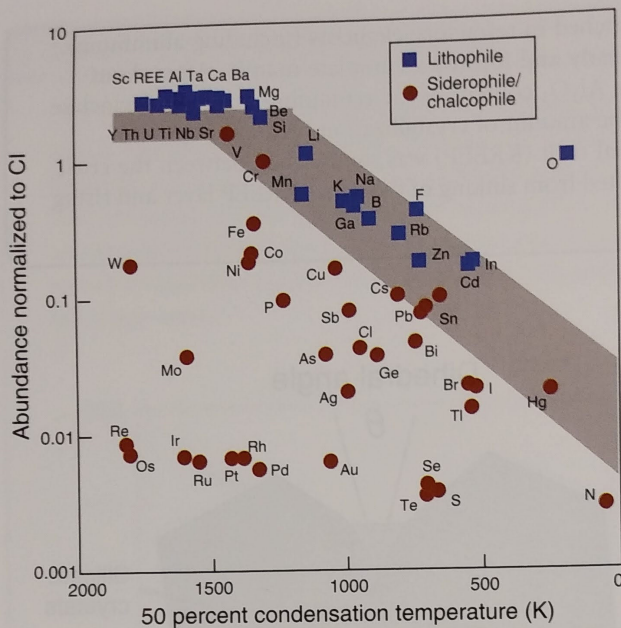


Figure 6.7 Depletions of siderophile and chalcophile elements in the Earth's mantle, compared to lithophile elements. All elements are plotted according to the temperature at which 50 percent of the element would have condensed from a nebula gas. Mantle abundances of siderophile elements are consistent with high-pressure equilibration, perhaps at the base of a magma ocean. Modified from Carlson et al. (2014).

relationship to volatility, because they have been modified by core formation.

The partitioning of siderophile elements depends on their relative affinity for metal versus silicate, which varies with pressure and temperature, as well as the oxidation state, which controls the amount of metallic Fe versus FeO. Experiments conducted under different conditions predict the partitioning of siderophile elements between mantle silicate and metal, and their results can be compared with measured siderophile element abundances in the mantle (Righter, 2011). These data indicate that the siderophile element partitioning occurred at high pressure, likely at the base of a deep magma ocean (Badro et al., 2015). However, the occurrence of the most highly siderophile elements (the elements at the bottom left of Figure 6.7) in the mantle in nearly constant (chondritic) relative proportions (Figure 6.7) suggests that the liquid metal was extracted completely into the core, and a veneer of chondritic material was accreted later. A small amount, less than 1 percent of the Earth's mass, of added chondrite would be sufficient to replace the highly siderophile elements in the mantle, but would not significantly affect other element abundances.

Another model for core formation in the Earth (Wade and Wood, 2005) posits that the core grew

gradually by accretion of already differentiated bodies. Each incoming protoplanet added its core to the growing metal mass at the center of the Earth. As the planet progressively increased in size, a global magma ocean deepened, and the pressures and temperatures at which metal and silicates equilibrated increased. This model is not applicable to core formation in small bodies, but it illustrates the complexities that are likely in planetary-sized bodies.

6.4.2 Going Up: Crusts

Crusts form from melts that ascend buoyantly and erupt on the surface or are emplaced as plutons. A critical amount of melting, usually a few percent (Maaloe, 2003), must occur before magmas can segregate from their source regions. Although melts can rise upward along grain boundaries, the formation of veins along fractures can drain melts more efficiently. Heat loss is also important in controlling magma ascent. The magma cooling rate is proportional to its surface area times the temperature difference with the surrounding rock. Calculations indicate that, to reach the surface, magma moving in a planar fracture must move 10^4 times faster than in a sphere of equivalent volume. As a consequence, spherical ascending magmas (diapirs) are more likely, at least until the magma nears the surface. On planets, tectonic control of the locations of melting allows magmas to be injected repeatedly in the same place, leading to subsurface magma chambers and volcanic edifices (McCoy et al., 2006). Multiple heat sources within planets allow crust formation to occur over billions of years. On small bodies, crust formation occurs without tectonic influences.

Taylor and McLennan (2009) distinguish “**primary**” and “**secondary**” crusts. A primary crust is formed during planetary differentiation. The Moon's anorthositic crust is primary, having crystallized directly from the lunar magma ocean. Primary crusts have commonly been destroyed by cataclysmic impacts and foundering during mantle overturn, although that appears not to have been the case for the Moon. A secondary crust is produced from magmas formed by partial melting of the mantle. The lunar mare basalts are an example of secondary crust.

Almost none of the Earth's primary (Hadean-age, earlier than 3.9 Ga) crust remains, and the evidence used to surmise its composition is enigmatic and contradictory. Its only unambiguous remnants are a handful of detrital zircon crystals with ages as old as 4.37 Ga that occur in younger sedimentary rocks. These zircons apparently crystallized in felsic magmas, but suggestions for an early, widespread granitic crust do not appear to be supported by other evidence.

BOX 6.2 METAL CORES IN ASTEROIDS

Vesta (~500 km diameter) is the only intact asteroid known to have a metallic core, with an estimated diameter of 220 km (Russell et al., 2012). Iron meteorites provide direct samples of core materials and demonstrate that core formation was common in planetesimals.

Iron meteorites are composed mostly of iron and nickel, with accessory minerals like troilite FeS, schreibersite $(\text{Fe,N})_3\text{P}$, and graphite C. Slow cooling produces the Widmanstätten pattern (Figure 6.8a), an intergrowth of low-nickel kamacite and high-nickel taenite crystals. The coarseness of the intergrowth structure and the diffusional zoning of nickel at crystal boundaries can be used to estimate the rate of cooling following solidification. The estimated cooling rates correspond to asteroids having diameters of 50–200 km. Because iron metal is so thermally conductive, cores should have uniform cooling rates throughout, and most iron groups bear this out. However, the meteorites in a few iron groups show radically different cooling rates. The explanation for this is that the insulating silicate mantle was stripped off by oblique (sometimes called “hit-and-run”) impacts, allowing metal at various depths within the naked cores to cool at different rates.

Pallasites, composed of olivine and metal (Figure 6.8b), were long thought to be samples of the boundaries between cores and the overlying mantles in asteroids. An alternative idea is that they are impact-generated mixtures of molten metal and mantle olivine.

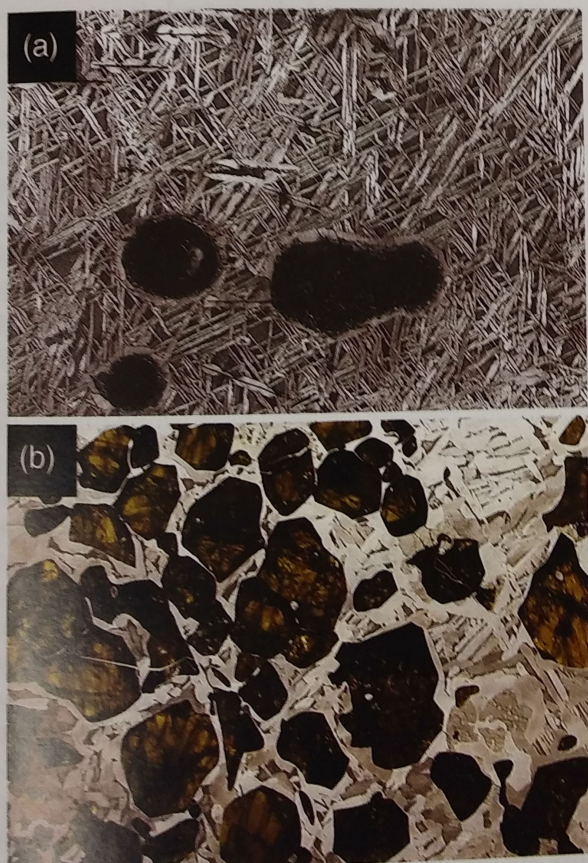


Figure 6.8 (a) Polished and etch slab of the Mount Edith (Australia) iron meteorite, exhibiting a characteristic Widmanstätten pattern formed by the intergrowth of two iron-nickel minerals, kamacite and taenite. The dark blobs are FeS (troilite). (b) Slab of the Fukang (China) pallasite, composed of olivine and metal. Smithsonian Institution images.

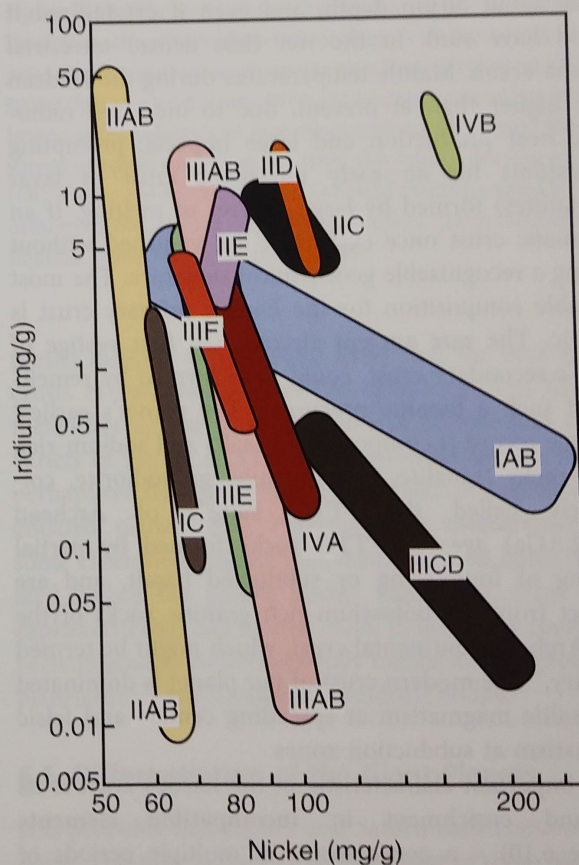


Figure 6.9 Iridium versus nickel abundances in iron meteorites. Compositional differences serve to classify irons. The trends are due to fractional crystallization and other processes during solidification of molten iron cores in asteroids. Modified from Scott and Wasson (1975).

Differences in the abundances of siderophile elements (iridium, gallium, germanium, nickel) are used to classify irons and pallasites. The iridium and nickel abundances of the most abundant iron meteorite groups are illustrated in Figure 6.9. There are lots of additional iron groups with only a few meteorites. Altogether, as many as 60 different asteroid cores may be represented in our meteorite collections. The taxonomy of irons reflects the tortured evolution of the classification system. Originally, irons were assigned to compositional groups I through IV, and some were later subdivided (III became IIIA and IIIB) and sometimes recombined (IIIAB) as more data became available.

The sloping chemical trends in Figure 6.9 reflect solidification processes in the molten cores. Fractional crystallization accounts for most of the observed trends, although more complex models including liquid immiscibility and trapping of melt between growing solids have also been advocated (Chabot and Haack, 2006). It is unclear whether asteroid cores solidified from the center outward (as in the Earth) or from the core-mantle boundary inward, and whether the crystals formed concentric layers or elongated dendrites.

Even though it seems likely that the Earth had an early magma ocean, it is unlikely to have formed an anorthositic crust like that of the Moon. Plagioclase is not stable below about 40 km depth, and even if crystallized it would have sunk in the wet (less dense) terrestrial magma ocean. Mantle temperatures during the Hadean were higher than at present, due to increased radiogenic heat production and large impacts, prompting suggestions for an early ultramafic crust of lavas (komatiites) formed by large degrees of melting. If an ultramafic crust once existed, it has vanished without leaving a recognizable geochemical signature. The most probable composition for the Earth's primary crust is basaltic. The rare ancient zircons, the first vestige of Earth's secondary crust, could have formed by remelting of such a basaltic precursor. The planet's earliest geologic record is composed of basalts and sodium-rich felsic rocks (tonalite, trondhjemite, granodiorite, collectively called the "TTG suite") of Archean (3.9–2.5 Ga) age. The TTG rocks formed by partial melting of foundering or subducted basalt, and are distinct from the potassium-rich granitic rocks of the post-Archean continental crust, which might be termed "tertiary." The modern crust of our planet is dominated by basaltic magmatism at spreading centers and felsic magmatism at subduction zones.

An important characteristic of the Earth's crust is its profound enrichment in **incompatible elements** (Figure 6.10) – a consequence of multiple periods of partial melting, each further concentrating these elements into the melt phase. The incompatible elements have large ionic size and/or charge, favoring their incorporation in magma over confining crystal structures. Sequestering of incompatible elements is also likely in the crusts of other planets, although probably not to the extent seen on Earth, where we even find ore deposits of these and

other elements concentrated by hydrothermal fluids or in highly evolved granitic magmas.

Extrapolating what has been learned about the Earth's crust to other planets and planetesimals is not very useful, because our planet's crust is compositionally unique. This is due mainly to our planet's liquid water, which enables plate tectonics and is ultimately responsible for the siliceous continental crust. Although the crusts of Mercury, Venus, and Mars are all dominated by basaltic rocks, consistent patterns in crustal evolution are elusive (Taylor and McLennan, 2009). The compositional variations among achondrites that represent crustal rocks from asteroids are even more pronounced, and the crusts of the satellites of the giant planets are stranger still. The compositions of igneous crusts on various Solar System bodies and the processes that generate and modify them are discussed more fully in Chapter 10.

6.4.3 What's Left: Mantles

We consider mantles last, because they are what remain after the extraction of metal to form cores and of silicate partial melts to form crusts. We previously noted that the Earth's mantle is strongly depleted in siderophile and chalcophile elements (Figure 6.7); conversely, its lithophile element abundances mimic those of the bulk Earth. The mantle and crust also have complementary geochemical patterns – whatever is concentrated in one is depleted in the other.

Whether or not mantles formed by solidification of a magma ocean or as residues from incomplete melting, they are composed of ultramafic rocks – a natural consequence of having chondritic bulk compositions. Any planet's primitive upper mantle is basically peridotite, composed of olivine + orthopyroxene + clinopyroxene + an aluminous mineral (plagioclase, spinel, or garnet, depending on pressure). Here we are not considering extremely high-pressure phases that occur in the lower mantles of the large terrestrial planets; these will be

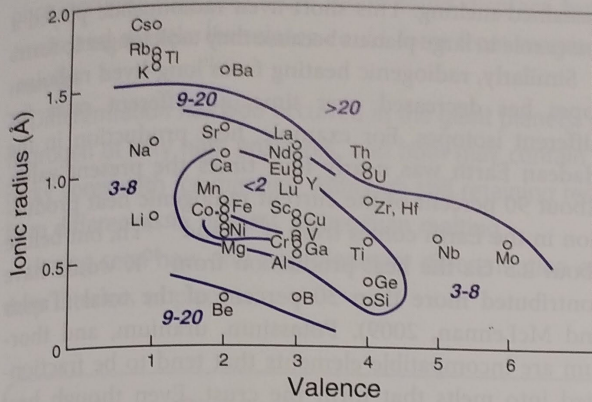


Figure 6.10 Incompatible elements have high size and charge, and are enriched in the Earth's continental crust relative to their abundances in the primitive mantle. Contours group elements with similar degrees of enrichment (italicized blue numbers). Modified from Taylor and McLennan (2009).

introduced in Chapter 7. Partial melting of peridotite yields basaltic magma. To see the effects of extracting basaltic magma on the Earth's mantle, let's examine a pressure-temperature diagram showing the stability of minerals in mantle residues (Figure 6.11). The mineralogy of the mantle at different pressures is shown on the left side of the diagram. With increasing temperature, solid rock crosses the solidus and begins to melt. The formation of basaltic magma first exhausts clinopyroxene and spinel or garnet, yielding a residue of olivine + orthopyroxene (this rock is called harzburgite). Further melting exhausts orthopyroxene, leaving only olivine (dunite). Basaltic magmas on Earth sometimes carry xenoliths of harzburgite or dunite, representing the mantle residues in their source regions. Although the various terrestrial planets may have somewhat different mantle compositions and phase relations, this example can in general explain how their mantle mineralogies vary when partial melts are extracted to form crusts. Interior pressures within planetesimals are lower, precluding the formation of garnet and spinel, but the basic melting relationships are similar except in mantles with very low oxidation states.

6.4.4 Another View: Partial Differentiation

Although planets have fully differentiated, the fate of planetesimals is less clear. Melted (achondritic) and unmelted (chondritic) meteorites are conventionally thought to have formed on different parent bodies. However, this picture has been muddled somewhat by evidence that may suggest some planetesimals experienced incomplete melting, resulting in differentiated interiors and outer, chondritic crusts (Weiss and Elkins-Tanton, 2013). The so-called primitive achondrites are residues from small degrees of

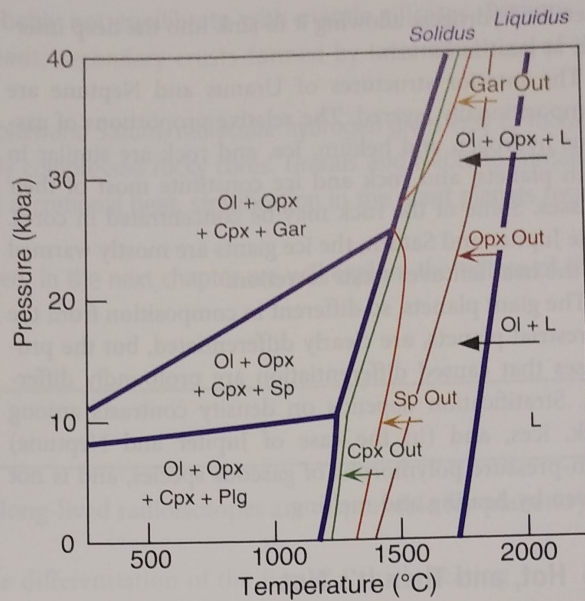


Figure 6.11 Schematic phase diagram for mantle peridotite, illustrating the effects of partial melting. With increasing pressure, the aluminous phase changes from plagioclase (Plg) to spinel (Sp) to garnet (Gar). With increasing temperature, melting begins at the solidus and first clinopyroxene (Cpx) and then garnet or spinel are exhausted from the residue, leaving olivine (Ol) + orthopyroxene (Opx). Continued melting exhausts orthopyroxene, leaving only olivine, until complete melting at the liquidus. After McSween et al. (2003).

partial melting, and it seems likely that the near-surface crusts of such bodies must be chondritic. The IIE iron meteorites appear to be mixtures of molten iron with a variety of silicate rocks, including basalts and chondrites, perhaps resulting from impact disruption of a partly differentiated planetesimal. Paleomagnetic measurements of some chondrites suggest that their magnetic fields originated within the parent bodies themselves, most easily explained if they had molten cores. In fact, there could have been all gradations between fully melted and differentiated planetesimals and unmelted chondritic bodies.

6.5 Differentiation of the Giant Planets

It is probable that Jupiter and Saturn each contain rocky cores. The main source of internal heat in these planets is gravitational potential energy gained during the bodies' accretion and, in the case of Saturn, by sinking of dense material. High-pressure experiments indicate that both gas giants exhibit a continuous transition from molecular hydrogen to metallic (ultradense, electrically conducting) hydrogen. In this case, the atmosphere can be considered a differentiated layer. Helium may be immiscible in

metallic hydrogen, allowing it to sink into the deep interior, at least in Saturn.

The interior structures of Uranus and Neptune are compositionally layered. The relative proportions of gaseous hydrogen and helium, ice, and rock are similar in both planets, and rock and ice constitute most of their masses. Some of the rock may be concentrated in cores. Like Jupiter and Saturn, the ice giants are mostly warmed by the heat left over from accretion.

The giant planets, so different in composition from the terrestrial planets, are clearly differentiated, but the processes that caused differentiation are profoundly different. Stratification depends on density contrasts among rock, ices, and (in the case of Jupiter and Neptune) high-pressure polymorphs of gaseous species, and is not driven by heating and melting.

6.6 Hot, and Then It's Not

The heat sources that caused planetary differentiation have changed with time. The primary heat source in early planetesimals was the decay of ^{26}Al . The short half-life of this radionuclide means that it was an effective heat source for only a few million years. As we saw in Figure 6.11, the aluminous phase melts early so that ^{26}Al is concentrated into partial melts that segregate into the crust, thereby further depriving the interior of the ability for more

sustained melting. This short-lived radionuclide played a lesser role in large planets because they took longer to form.

Similarly, radiogenic heating from long-lived radioisotopes has decreased over time, at different rates for different isotopes. For example, heat production in the Hadean Earth was nearly four times the present value. About 90 percent of the current radiogenic heat production in the Earth comes from ^{238}U and ^{232}Th , but before about 2.5 Ga the heat production from ^{40}K would have contributed more than 30 percent of the total (Taylor and McLennan, 2009). Potassium, uranium, and thorium are incompatible elements that tend to be fractionated into melts that form the crust. Even though heat from the crust can escape faster than heat from the mantle, these are still very important heat sources for planets.

Currently, the Earth's heat flow is 42 terawatts ($42 \times 10^{12}\text{ W}$). The decay of long-lived radionuclides is estimated to produce $\sim 18\text{ TW}$. The remainder is probably attributable to exothermic crystallization of the liquid outer core, although some portion is likely to be fossil accretional heat.

Heating of planets by impacts was much more significant in the early Solar System, especially during the waning stages of accretion, when large bodies of comparable size collided. Impact heating is no longer a significant contributor.

Summary

The sources for heating planetesimals and planets are:

- decay of long-lived (^{40}K , ^{235}U , ^{238}U , ^{232}Th) and short-lived (especially ^{26}Al) radionuclides, resulting in mass loss that is transformed into energy;
- segregation of metal cores, which transforms gravitational potential energy into thermal energy, and exothermic crystallization of molten core metal;
- large impacts that convert kinetic energy of the impactor into heat; and
- gravitational tides, where frictional heating occurs as a body flexes along its orbital path.

Short-lived radionuclide decay was particularly important in heating planetesimals in the early Solar System; long-lived radionuclides are not very important for small bodies because heat escapes faster than it can build up. Planets early in their histories were primarily heated by large impacts and core segregation; planetary heating in recent times is dominated by long-lived radionuclide decay and exothermic crystallization in bodies with partly liquid cores. Tidal heating affects some satellites of the giant planets.

Large-scale melting to produce magma oceans is thought to have been common for terrestrial planets in the early Solar System, although its evidence remains elusive (except for the Moon). Pervasive melting would have aided differentiation. Metal drops could sink through, and equilibrate with, a magma ocean. Metallic liquids that could "wet" silicates were probably required to separate core materials from a solidified mantle, and

once accumulated into large blobs could displace but probably not equilibrate with mantle silicates. Primary crusts and mantle cumulates could form from magma oceans; secondary crusts formed by later remelting of the mantle.

Differentiation has also occurred in the giant planets. In Jupiter and Saturn, molecular hydrogen gives way to metallic hydrogen at very high pressures, and both may contain highly compressed rocky cores. Uranus and Neptune contain rocky cores with ice mantles. Although still retaining residual accretional heat, stratification in the giant planets results from differences in density, rather than melting.

Having seen how and why planetary differentiation occurred, in the next chapter we will explore the nature of the deep interiors of planets.

Review Questions

1. Why does radioactive decay release heat, and which long-lived radioisotopes are responsible for planetary heating?
2. Which heating mechanisms were responsible for the differentiation of the Earth? Which heating mechanisms are important now?
3. What evidence suggests that the Moon's differentiation involved a magma ocean?
4. What controls the separation from the mantle of core materials and of magmas that become the crust?
5. How is differentiation in the giant planets different from the terrestrial planets?

SUGGESTIONS FOR FURTHER READING

Badro, J., and Walter, M. (2015) *The Early Earth: Accretion and Differentiation*. New York: John Wiley & Sons. Everything you need to know about our planet's formation and differentiation, all in one place.

Carlson, R. W., Garnero, E., Harrison, T. M., et al. (2014) How did early Earth become our modern world? *Annual Reviews of Earth and Planetary Sciences*, **42**, 151–178. This up-to-date review summarizes the Earth's geochemical inheritance from protoplanets, the crystallization of its early magma ocean, and the formation of its core, mantle, and crust.

Elkins-Tanton, L. T. (2012) Magma oceans in the inner Solar System. *Annual Review of Earth and Planetary Sciences*, **40**, 113–139. A thoughtful assessment of the formation and solidification of magma oceans on planetesimals and planets.

Elkins-Tanton, L. T., and Weiss, B. P., eds. (2017) *Planetesimals: Early Differentiation and Consequences for Planets*. Cambridge: Cambridge University Press. This book describes the compositional diversity of planetesimals and their magmatic differentiation.

Rubie, D. C., Nimmo, F., and Melosh, H. J. (2007) Formation of Earth's core. In *Treatise in Geochemistry*, Vol. 9, eds. Holland, H. D., and Turekian, K. K. Oxford: Elsevier, pp. 51–90. This chapter provides a comprehensive survey of the physics of core formation, and its compositional and temporal constraints.

Elsevier, pp. 51–90. This chapter provides a comprehensive survey of the physics of core formation, and its compositional and temporal constraints.

Taylor, S. R., and McLennan, S. M. (2009) *Planetary Crusts: Their Composition, Origin and Evolution*. Cambridge: Cambridge University Press. Here's a whole book on crusts, providing a superb overview of the nature and origin of crusts of different planets.

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