

# Theory of the Earth

Don L. Anderson

Chapter 2. Earth and Moon

Boston: Blackwell Scientific Publications, c1989

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Recommended citation:

Anderson, Don L. Theory of the Earth. Boston: Blackwell Scientific Publications, 1989. <http://resolver.caltech.edu/CaltechBOOK:1989.001>

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Abstract:

Recent compilations of cosmic abundances (Table 2-1) agree fairly closely for the more significant rock-forming elements. From these data a simple model can be made (Table 2-2) of the Earth's bulk chemical composition. The mantle is nearly completely oxidized and is therefore composed of compounds primarily of MgO, SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, CaO, Na<sub>2</sub>O and Fe-oxides.

# Earth and Moon

*Strange all this difference should be  
Twixt tweedle-dum and tweedle-dee.*

—JOHN BYROM

## BULK COMPOSITION OF THE EARTH

Recent compilations of cosmic abundances (Table 2-1) agree fairly closely for the more significant rock-forming elements. From these data a simple model can be made (Table 2-2) of the Earth's bulk chemical composition.

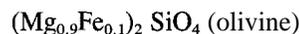
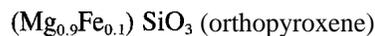
The mantle is nearly completely oxidized and is therefore composed of compounds primarily of MgO, SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, CaO, Na<sub>2</sub>O and Fe-oxides. The major minerals in rocks from the upper mantle are



and



Iron occurs in various oxidation states, FeO, Fe<sub>2</sub>O<sub>3</sub>, Fe<sub>3</sub>O<sub>4</sub>, and typically replaces about 10 percent of the MgO in the above minerals:



These minerals can be considered as solid solutions between MgSiO<sub>3</sub> and FeSiO<sub>3</sub> (ferrosilite) and Mg<sub>2</sub>SiO<sub>4</sub> and Fe<sub>2</sub>SiO<sub>4</sub> (fayalite). Another important solid-solution series is between MgO (periclase) and FeO (wustite) giving (Mg,Fe)O (magnesiowüstite or ferropiclase). These oxides usually combine with SiO<sub>2</sub> to form silicates, but they may occur as separate phases in the lower mantle.

The approximate composition of the mantle and the relative sizes of the mantle and core can be determined from the above consideration. Oxygen is combined with the ma-

ior rock-forming elements, and then the relative weight fractions can be determined from the atomic weights. The preponderance of silica (SiO<sub>2</sub>) and magnesia (MgO) in mantle rocks gave rise to the term "sima," an archaic term for the chemistry of the mantle. In contrast, crustal rocks were called "sial" for silica and alumina.

The composition in the first column in Table 2-2 is based on Cameron's (1982) cosmic abundances. These are converted to weight fractions via the molecular weight and renormalization. The Fe<sub>2</sub>O requires some comment. Based on cosmic abundances, it is plausible that the Earth's core is mainly iron; however, from seismic data and from the total mass and moment of inertia of the Earth, there must be a light alloying element in the core. Of the candidates that have been proposed (O, S, Si, N, H, He and C), only oxygen and silicon are likely to be brought into a planet in refractory solid particles—the others are all very volatile elements and will tend to be concentrated near the surface or in the atmosphere. The hypothetical high-pressure phase Fe<sub>2</sub>O has about the right density to match core values. There is also the possibility that the oxidation state of iron decreases with pressure from Fe<sup>3+</sup> to Fe<sup>2+</sup> to Fe<sup>+</sup> to Fe<sup>0</sup>. At high pressure FeO is soluble in molten iron, giving a composition similar to Fe<sub>2</sub>O at core pressures (Ohtani and Ringwood, 1984).

If most of the iron is in the core, in Fe<sub>2</sub>O proportions, then the mass of the core will be 30 to 34 weight percent of the planet. The actual mass of the core is 33 percent, which is excellent agreement. There may also, of course, be some sulfur, carbon, and so on in the core, but little or none seems necessary. The FeO content of the mantle is less than 10 percent so that does not change the results much. The mass of the core will be increased by about 5 percent if it

TABLE 2-1  
Short Table of Cosmic Abundances (Atoms/Si)

Element	Cameron (1982)	Anders and Ebihara (1982)
O	18.4	20.1
Na	0.06	0.057
Mg	1.06	1.07
Al	0.085	0.0849
Si	1.00	1.00
K	0.0035	0.00377
Ca	0.0625	0.0611
Ti	0.0024	0.0024
Fe	0.90	0.90
Ni	0.0478	0.0493

has a cosmic Ni/Fe ratio, and will be larger still if the newer estimates of solar iron abundances are used, unless the iron content of the mantle is greater than is evident from upper-mantle rocks. The new solar values for Fe/Si give 15% weight percent FeO for the mantle if the core is 32% with an Fe<sub>2</sub>O composition. This FeO content is similar to estimates of the composition of the lunar and Martian mantles.

Note that the cosmic atomic ratio Mg/Si is 1.06. This means that the mantle (the iron-poor oxides or silicates) is predicted to be mainly pyroxene, MgSiO<sub>3</sub>, rather than olivine, Mg<sub>2</sub>SiO<sub>4</sub>, which has a Mg/Si ratio of 2 (atomic). The presence of FeO in the mantle means that slightly more olivine can be present. It is the (Fe + Mg)/Si ratio that is important.

Garnet also has an Mg/Si ratio near 1, although garnet usually has more FeO than do olivine and pyroxene. This suggests that much of the mantle is composed of pyroxene and garnet, rather than olivine, which is the main constituent of the shallow mantle. The CaO and Al<sub>2</sub>O<sub>3</sub> are contained in garnet and clinopyroxene and the Na<sub>2</sub>O in the jadeite molecule at upper mantle pressures. Seismic data are consistent with a mainly "perovskite" lower mantle. MgSiO<sub>3</sub> transforms to a perovskite structure, an analog of CaTiO<sub>3</sub>, at lower-mantle pressures. MgSiO<sub>3</sub>, "perovskite" is therefore the dominant mineral of the mantle.

TABLE 2-2  
Simple Earth Model Based on Cosmic Abundances

Oxides	Molecules	Molecular Weight	Grams	Weight Fraction
MgO	1.06	40	42.4	0.250
SiO <sub>2</sub>	1.00	60	60.0	0.354
Al <sub>2</sub> O <sub>3</sub>	0.0425	102	4.35	0.026
CaO	0.0625	56	3.5	0.021
Na <sub>2</sub> O	0.03	62	1.84	0.011
Fe <sub>2</sub> O	0.45	128	57.6	0.339
Total			169.7	1.001

The crust is rich in SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> (thus "sial"), CaO and Na<sub>2</sub>O. The Al<sub>2</sub>O<sub>3</sub> and CaO and Na<sub>2</sub>O content of the "cosmic" mantle totals to 5.8 percent. The crust represents only 0.5 percent of the mantle. Thus, there must be considerable potential crust in the mantle. The thinness of the terrestrial crust, compared to the Moon and Mars and relative to the potential crust, is partially related to the presence of plate tectonics on Earth, which continuously recycles crustal material. It is also due to high pressures in the Earth, which convert crustal materials such as basalt to high-density eclogite.

Estimates of upper-mantle compositions based on peridotites are deficient in SiO<sub>2</sub>, TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub> and CaO relative to cosmic abundances. Basalts, picrites and eclogites are enriched in these oxides. Cosmic ratios of the refractory oxides can be obtained with a mixture of approximately 80 percent peridotite and 20 percent eclogite. If the lower mantle is mainly MgSiO<sub>3</sub>, ("perovskite"), then a much smaller eclogitic, or basaltic, component is required to balance the Mg/Si ratio of upper-mantle peridotites. A chondritic Earth contains about 6 to 10 percent of a basaltic component.

## EVOLUTION OF THE EARTH'S INTERIOR

The evolution of Earth, its outgassing and its differentiation into a crust, mantle and core depend on whether it accreted hot or cold and whether it accreted homogeneously or inhomogeneously. It now appears unavoidable that a large fraction of the gravitational energy of accretion was trapped by the Earth, and it therefore started life as a hot body relative to the melting point. The processes of differentiation, including core formation, were probably occurring while the Earth was accreting, and it is therefore misleading to talk of a later "core formation event." Melting near the surface leads to a zone refining process with light melts rising to the surface and dense melts and residual crystals sinking toward the center. What is not so clear is whether most of the Earth accreted homogeneously or inhomogeneously. A certain amount of inhomogeneity in the chemistry of the accreting material is required, otherwise reactions between H<sub>2</sub>O and free iron would oxidize all the iron and no water would exist at the surface. This can be avoided by having more iron accrete in the early stages and more H<sub>2</sub>O in the later stages. Iron metal, of course, condenses earlier in a cooling nebula, and because of its density and ductility may have formed the earliest planetesimals and perhaps the initial nuclei of the planets.

In earlier theories of Earth evolution, it was assumed that cold volatile-rich material similar to type I carbonaceous chondrites accreted to form a homogeneous planet, perhaps with some reduction and vaporization at the surface to form reduced iron, which subsequently warmed up and

differentiated into an iron-rich core and a crust. In these theories it proved difficult to transport molten iron to the center of the planet because of the effect of pressure on the melting point: Molten iron in the upper mantle would freeze before it reached the lower mantle. A central iron-rich nucleus mixed with or surrounded by refractory-rich material including aluminum-26, uranium and thorium would alleviate this thermal problem as well as the iron-oxidation problem. The presence of an ancient magnetic field, as recorded in the oldest rocks, argues for a sizeable molten iron-rich core early in Earth's history. The abundance of siderophile elements in the upper mantle also suggests that this region was not completely stripped of these elements by molten iron draining to the core.

The early and extensive melting and differentiation of the Moon and some meteorite parent bodies attests to the importance of melting and differentiation of small bodies in the early solar system. This melting may have been caused by the energy of accretion or the presence of extinct, short-lived radioactive nuclides. In some cases, as in present-day Io, tidal pumping is an important energy source. Isotopic studies indicate that distinct geochemical reservoirs formed in the mantle early in its history. Magmas from these reservoirs retain their isotopic identity, and this proves that the mantle has not been homogenized in spite of the fact that it has presumably been convecting throughout its history. A chemically stratified mantle is one way to keep reservoirs distinct and separate until partial melting processes allow separation of magmas, which rise quickly to the surface.

Zone refining during accretion and crystallization of a deep magma ocean are possible ways of establishing a chemically zoned planet (Figure 2-1). At low pressures basaltic melts are less dense than the residual refractory crystals, and they rise to the surface, taking with them many of the trace elements. The refractory crystals themselves are also less dense than undifferentiated mantle and tend to concentrate in the shallow mantle. At higher pressure there is a strong likelihood that melts become denser than the crystals they are in equilibrium with. Such melts, trapped at depth and insulated by the overlying rock, may require considerable time to crystallize.

Most scientists agree that simplicity is a desirable attribute of a theory. Simplicity, however, is in the eye of the beholder. The end results of natural processes can be incredibly complex, even though the underlying principles may be very simple. A uniform, or homogeneous, Earth is probably the "simplest" theory, but it violates the simplest tests. Given the most basic observations about the Earth, the next simplest theory is a three-part Earth: homogeneous crust, homogeneous mantle and homogeneous core. How the Earth might have achieved this simple state, however, involves a complex series of ad hoc mechanisms to separate core and crust from the primitive mantle and then to homogenize the separate products.

On the other hand, one can assume that the processes

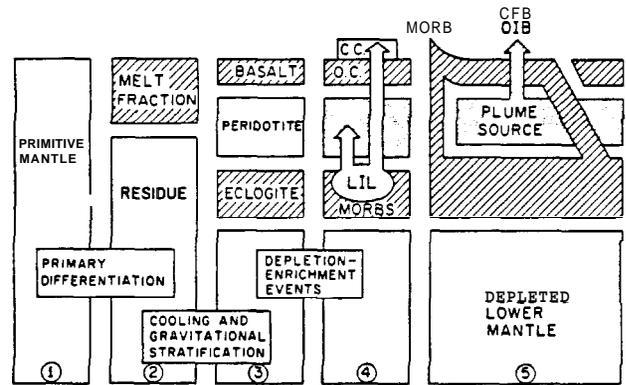


FIGURE 2-1

A model for the evolution of the mantle. Primitive mantle (1) is partially molten either during accretion or by subsequent whole-mantle convection, which brings the entire mantle across the solidus at shallow depths. Large-ion lithophile (LIL) elements are concentrated in the melt. The deep magma ocean (2) fractionates into a thin plagioclase-rich surface layer and deeper olivine-rich and garnet-rich cumulate layers (3). Late-stage melts in the eclogite-rich cumulate are removed (4) to form the continental crust (C.C.), enrich the peridotite layer and deplete MORBS, the source region of oceanic crust (O.C.) and lower oceanic lithosphere. Partial melting of the plume source (5) generates continental flood basalts (CFB), ocean-island basalts (IOB) and other enriched magmas, leaving a depleted residue (harzburgite) that stays in the upper mantle. The MORB source may be the main basalt reservoir in the mantle, and enriched or hot-spot magmas (OIB, CFB) may be MORB contaminated by interaction with shallow mantle.

that formed and differentiated the Earth were uniform and ask what the resulting Earth would look like. For example, the crust is made of the lighter and more fusible materials and was formed as a result of igneous differentiation involving the upward migration of light melts. The core is an iron-rich alloy that also melts at low temperature but drains to the interior because of its high density. This suggests a simple hypothesis: The stratification of the Earth is a result of gravitational separation of materials according to their melting points and density. The materials accreting to form the Earth *may* have been uniform, but the high temperatures associated with accretion, even a violent accretion, would result in a chemically differentiated planet. Thus, a simple, even obvious, process gives a complex result.

The mantle itself is neither expected to be nor observed to be homogeneous. Let us follow this single, simple hypothesis further. As the Earth grows, the crustal elements are continuously concentrated into the melts and rise to the surface. When these melts freeze, they form the crustal minerals that are rich in silicon, calcium, aluminum, potassium and the large-ion lithophile (LIL) elements. Melts generally are also rich in FeO compared to primitive material. This plus the high compressibility of melts means that the densities of melts and residual crystals converge, or

even cross, as the pressure increases. Melt separation is therefore more difficult at depth, and melts may even drain downward at very high pressure. However, during accretion the majority of the melt-crystal separation occurs at low pressure. All of the material in the deep interior has passed through this low-pressure melting stage in a sort of continuous zone refining. The magnesium-rich minerals,  $\text{Mg}_2\text{SiO}_4$  and  $\text{MgSiO}_3$ , have high melting temperatures and are fed through the melting zone into the interior. Even if the accreting material is completely melted during assembly of the Earth, these minerals will be the first to freeze, and they will still separate from the remaining melt.  $(\text{Mg,Fe})_2\text{SiO}_4$  has a slightly higher melting or freezing point than  $(\text{Mg,Fe})\text{SiO}_3$  and a slightly higher density, so that olivine may even separate from orthopyroxene. This separation, however, is expected to be much less effective than the separation of olivine and orthopyroxene from the melts rich in  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$ ,  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  which differ substantially from these minerals in both density and melting point. The downward separation of iron-rich melts, along with nickel, cobalt, sulfur and the trace siderophile elements, strips these elements out of the crust and mantle. The  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$  and  $\text{Na}_2\text{O}$  content in chondritic and solar material is adequate to form a crust some 200 km thick. The absence of such a massive crust on the Earth might suggest that the Earth has not experienced a very efficient differentiation. On the other hand, the size of the core and the extreme concentration of the large-ion, magmaphile elements into the crust suggest that differentiation has been extremely efficient.

The solution to this apparent paradox does not require special pleading. At pressures corresponding to depths of the order of 50 km, the low-density minerals of the crust convert to a mineral assemblage denser than olivine and orthopyroxene. Most of the original crust therefore is unstable and sinks into the mantle. Any melts below 200–400 km may suffer the same fate. Between 50 and 500 km the  $\text{Al}_2\text{O}_3$ - $\text{CaO}$ - $\text{Na}_2\text{O}$ -rich materials crystallize as clinopyroxene and garnet, a dense eclogite assemblage that is denser than peridotite. Eclogite transforms to a garnet solid solution, which is still denser and which is stable between about 500 and 800 km. Peridotite also undergoes a series of phase changes that prevent eclogite from sinking deeper than about 650 km. However, when the Earth was about Mars' size and smaller, the eclogite could sink to the core-mantle boundary.

The base of the present mantle, a region called D", is anomalous and may be composed of high-pressure eclogite. The earliest stages of accretion probably involved the most refractory materials under nebular conditions. These include compounds rich in Fe and CaO,  $\text{Al}_2\text{O}_3$  and  $\text{TiO}_2$ . The refractory lithophiles would have been excluded from a molten Fe-rich core, and these may also be concentrated in D". In either case D" would be enriched in  $\text{Al}_2\text{O}_3$  and CaO. Equilibration between material in D" and the core may also result in a high FeO content for this region. If D" is intrin-

sically denser than the rest of the lower mantle, it would be gravitationally stable at the base of the mantle. On the other hand, it is embedded in the thermal boundary layer between mantle and core and therefore has a high temperature that may locally permit D" material to rise into the lower mantle until it becomes neutrally buoyant. As it cools, it will sink back to D".

The end result for a planet experiencing partial melting, gravitational separation and phase changes is chemical stratification. The possibility of three "basaltic" regions (high CaO and  $\text{Al}_2\text{O}_3$  and possibly FeO, relative to MgO) has been identified. These regions are the crust, the transition region (between upper and lower mantle) and D". The latter two may be the result of solid subduction or sinking of high-density melts.

The bulk of the upper mantle and lower mantle may therefore be sandwiched between basalt-rich layers. If melting during accretion extended below some 300 km, the composition of the melt and residual refractory phases changes. Orthopyroxene transforms to majorite, a garnet-like phase that replaces olivine on the liquidus. Melts, therefore, are MgO rich, and we have another mechanism for separating major elements in the mantle and concentrating olivine ( $\text{Mg}_2\text{SiO}_4$ ) in the shallow mantle. Giant impacts in early Earth history have the potential for melting mantle to great depth and for concentrating dense refractory residual phases such as orthopyroxene-majorite in the lower mantle.

The subsequent cooling and crystallization of the Earth introduces additional complications. A chemically stratified mantle cools more slowly than a homogeneous Earth. Phase change boundaries are both temperature and pressure dependent, and these migrate as the Earth cools. A thick basalt crust, stable at high temperature, converts to eclogite at its base as it cools through the basalt-eclogite phase boundary. The initial crust of the Earth, or at least its deeper portions, therefore can become unstable and plunge into the mantle. This is an effective way to cool the mantle and to displace lighter and hotter material to the shallow mantle where it can melt by pressure release, providing a continuous mechanism for bringing melts to the surface.

The separation of melts and crystals is a process of differentiation. Convection is often thought of as a homogenization process, tantamount to stirring. Differentiation, however, can be irreversible. Melts that are separated from the mantle when the Earth was smaller or from the present upper mantle crystallize to assemblages that have different phase relations than the residual crystals or original mantle material. If these rocks are returned to the mantle, they will not in general have neutral buoyancy, nor are they necessarily denser than "normal" mantle at all depths. Eclogite, for example, is denser than peridotite when the latter is in the olivine,  $\beta$ -spinel and  $\gamma$ -spinel fields but is less dense than the lower mantle, and it transforms to dense perovskite-bearing assemblages at higher pressure than peridotite.

## ORIGIN OF THE CRUST

Although the crust represents less than 0.5 percent of the mass of the Earth, it contains a large fraction of the elements that preferentially enter the melt when a silicate is melted—the large-ion lithophile elements (LIL). For example, it can be estimated that the continental crust contains 58 percent of the rubidium, 53 percent of the cesium, 46 percent of the potassium, 37 percent of the barium and 35 percent of the uranium and thorium in the crust-mantle system. Other highly concentrated elements include bismuth (34 percent), lead (32 percent), tantalum (30 percent), chlorine (including that in seawater, 26 percent), lanthanum (19 percent), and strontium (13 percent). These high concentrations in such a small volume mean that the Earth is an extensively differentiated body. Apparently, most or all of the mantle has been processed by partial melting and upward melt extraction. It has also been estimated that 77 percent of the argon-40 produced by the decay of potassium-40 in the mantle and crust resides in the atmosphere. This also points toward a well-differentiated, and outgassed, Earth. It is possible that most of the H<sub>2</sub>O is in the ocean and the crust.

The age of the oldest continental crust is at least  $3.8 \times 10^9$  years, and perhaps as old as  $4.2 \times 10^9$  years; isotopic results from mantle rocks are also consistent with ancient differentiation. Yet the mean age of the continental crust is  $1.5 \times 10^9$  years (Jacobsen and Wasserberg, 1979). The energy of accretion of the Earth is great enough to melt a large fraction of the incoming material, therefore the processes of melting, crust formation and outgassing were probably contemporaneous with accretion. The absence of older crust may reflect extensive bombardment in the later stages of accretion and a high-temperature crust and upper mantle rather than a late onset of the crust-forming process. A stable crust may have been delayed by the freezing of a deep magma ocean.

Most of the continental crust formed between 2.5 and 3 Ga ago. In more recent times a small amount of continental crust has been added by accretion of island arcs, oceanic islands and plateaus and by continental flood basalts. Some crustal material is eroded, subducted and recycled into the mantle, but the net effect is nearly constant crustal volume over the past 1–2 Ga. Estimates of growth rates of the continental crust are 0.1–0.5 km<sup>3</sup>/yr at present, 0.3–0.7 km<sup>3</sup>/yr over the past 2.5 Ga and 5.7–6.6 km<sup>3</sup>/yr at the end of the **Archean**, or approximately ten times the present rate (McLennan and Taylor, 1980).

Isotopic ages of the crust suggest that crustal growth has been episodic with major additions in the time intervals 3.8–3.5, 2.8–2.5, 1.9–1.6, 1.6–1.2, 1.2–0.9 and 0.5–0 Ga. The composition of the crust has varied with time, most abruptly at the Archean-Proterozoic boundary. In particular there are decreases in chromium and nickel and increases in REE, thorium, uranium, Th/U and <sup>87</sup>Sr/<sup>86</sup>Sr (see Weaver

and Tarney, 1984, and McLennan and Taylor, 1980, for reviews).

The formation of the original crust is probably linked to the formation and evolution of a magma ocean. However, it is harder to form a crust on a magma ocean than on a watery ocean because a silicate crust is probably denser than the liquid it freezes from. Feldspar crystals of sufficient size might be able to break away from convection and float, but they only form after extensive freezing and grow only slowly. The density contrast is also relatively low. Gas-crystal packets, or foams, may provide sufficient buoyancy even though the individual crystals are small. The interstitial fluids formed in deep olivine-orthopyroxene cumulates may be the source of the protocrust—the crust may have formed by the freezing of light fluids that rose to the surface of the magma ocean. Certainly, the crust is extremely enriched in the incompatible elements, and forming it from a liquid from which the refractory crystals have been removed is one way to account for this enrichment.

The major additions to the continental crust in recent times are due to the lateral accretion of island arcs. The average composition of the continental crust may therefore be similar to the average island-arc andesite (Taylor and McLennan, 1981). The andesite model can be used to estimate the composition of the lower crust if it is assumed that andesite melts during orogenesis to form a granodioritic upper crust and a refractory, residual lower crust. The total average crust, in these models, is average island-arc andesite, and the upper-crust composition is approximated by the average post-Archean sediment. There is some evidence, however, that island arcs are more basic than *andesite*. Seismic data also suggest that the average crust is closer to quartz diorite and granodiorite than to andesite; that is, it is more siliceous. Heat flow suggests that the lower crust is rather depleted in the radioactive heat-producing elements (K, U, Th). An alternative model for continental crust chemistry was derived by Weaver and Tarney (1984) using a variety of xenolith and geophysical data. Both models are listed in Table 2-3.

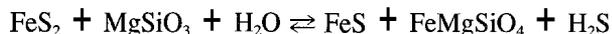
## ORIGIN OF THE MANTLE AND CORE

Larimer (1967) has outlined the condensation history of a cooling gas of cosmic composition. Compounds such as CaTiO<sub>3</sub>, MgAl<sub>2</sub>O<sub>4</sub>, Al<sub>2</sub>SiO<sub>5</sub>, and CaAl<sub>2</sub>Si<sub>2</sub>O<sub>3</sub> condense first at temperatures between 1740 and 1620 K (see Figure 1-6). Iron condenses next at 1620 K. Magnesium-rich pyroxenes and olivines condense between 1740 and 1420 K; FeS condenses at 680 K and H<sub>2</sub>O at 210 K. All the above temperatures were calculated on the assumption of a total pressure of  $6.6 \times 10^{-3}$  atmosphere. Larimer and Anders (1970) concluded that the fractionation patterns in meteorites occurred in the solar nebula as it cooled from high temperatures and could not be produced in the meteorite parent

bodies. Especially, they inferred the following accretion temperatures from the abundance patterns: carbonaceous chondrites, 400 K or below; enstatite and ordinary chondrites, 400 to 650 K; and the major fraction of iron meteorites, 1100 K or above.

Accretion of the planets presumably involved planetesimals that condensed over the entire temperature range, although the different planets may have incorporated different proportions of the various condensates. As a planet grows, the accretional energy increases and the temperature at the surface of the body is controlled by a balance between the available gravitational or kinetic energy, the heat capacity and thermal conductivity of the surface layer, the heats of reaction involved in chemical reactions occurring at the surface, and the reradiation of energy to the dust-gas cloud. As the planet grows, it will be less and less capable of retaining the volatiles brought in by the accreting particles, and the fractionation, in time, will be roughly the reverse of the condensation procedure. Refractories enter the planet and the volatiles contribute to the atmosphere. When the surface temperature reaches 680 K, the reaction  $\text{FeS} + \text{H}_2 \rightarrow \text{Fe} + \text{H}_2\text{S}$  will occur if the planet is accreting in a  $\text{H}_2$ -rich environment; the planet will, thereafter, not be able to incorporate much FeS into its interior.

Reactions such as



between the atmosphere and the surface of the accreting planet buffer the surface temperature and keep it below 500 K for long periods of time during the accretion process. Substantial amounts of heat, buried in the planet by the impacting bodies, are unavailable for reradiation if the rate of accretion is faster than the rate of heat conduction to the surface. The above considerations prolong the period available for trapping FeS in the interior. Eventually, however, dense impact-generated atmospheres serve to keep the surface insulated and hot during most of the accretion process.

Implicit in the above discussion is the assumption that the planets did not start to accrete until the condensation process was complete and that the solar nebula was relatively cold. A growing planet would incorporate the solid portion of available material—that is, the "local cosmic abundances"—in its interior in its initial stages of growth but could retain only the more refractory compounds as it grew. A large-scale redistribution of material must occur later in its history in order to form a core and to transport the retained volatiles and the less refractory compounds from the center of the body to the surface to form the crust and atmosphere. Thus the Earth, for example, must have turned itself inside out to obtain its present configuration.

**TABLE 2-3**  
Estimates of Average Bulk Composition of the Continental Crust

Species	A	B	C
SiO <sub>2</sub> (percent)	58.0	63.7	57.3
TiO <sub>2</sub>	0.8	0.5	0.9
	18.0	15.8	15.9
FeO	7.5	4.7	9.1
MnO	0.14	0.07	—
MgO	3.5	2.7	5.3
CaO	7.5	4.5	7.4
Na <sub>2</sub> O	3.5	4.3	3.1
K <sub>2</sub> O	1.5	2.0	1.1
P <sub>2</sub> O <sub>5</sub>	—	0.17	—
Rb (ppm)	42	55	32
Sr	400	498	260
Th	4.8	5.1	3.5
U	1.25	1.3	0.91
Pb	10	15	8

A: Andesite model (Taylor and McLennan, 1985).

B: Amphibolite-granulite lower crustal model (Weaver and Tarney, 1984).

C: Theoretical model (Taylor and McLennan, 1985).

If the planets accreted while condensation was taking place, a chemically zoned planet would result, with compounds that condense above 1620 K, such as  $\text{CaTiO}_3$ ,  $\text{MgAl}_2\text{O}_4$ ,  $\text{Al}_2\text{SiO}_5$ ,  $\text{CaAl}_2\text{Si}_2\text{O}_8$ , and iron, forming the central part of the body. In the next 160 K range of cooling, compounds such as  $\text{Ca}_2\text{SiO}_4$ ,  $\text{CaSiO}_3$ ,  $\text{CaMgSi}_2\text{O}_6$ ,  $\text{KAlSi}_3\text{O}_8$ ,  $\text{MgSiO}_3$ ,  $\text{SiO}_2$ , and nickel condense. Between 1420 and 1200 K the rest of the important mantle minerals,  $\text{MgSiO}_4$ ,  $\text{NaAlSi}_3\text{O}_8$ , and  $(\text{Na, K})_2\text{SiO}_3$  condense. Further cooling would bring in various metals (Cu, Ge, Au, Ga, Sn, Ag, Pb), the sulfides of zinc, iron and cadmium, and, finally,  $\text{Fe}_3\text{O}_4$  (400 K), the hydrated silicates ( $\sim 300$  K),  $\text{H}_2\text{O}$  (210 K), and the rare gases.

In both of the above models of accretion, the composition of the planet changes as it grows, and there is no compelling reason to assume that the composition of the various planets should be the same or the same as any single class of meteorites. Carbonaceous chondrites probably represent the last material to condense from a cooling nebula rather than representative planetary material. In the cold-accretion model, this material would be mixed with material that condensed at higher temperatures, such as ordinary chondritic and iron meteorites, to form the nucleus of the accreting planet, which would contain most of the volatiles such as  $\text{H}_2\text{O}$  and FeS. As the planet grows, the material that can be retained on the surface and incorporated into the interior becomes progressively less representative of the material available.

The amount of volatiles and low-temperature minerals retained by a planet depends on the fraction of the planet that accretes during the initial and terminal stages. In the accretion-during-condensation model, the full complement

of volatiles and sulfides must be brought in during the terminal stage. In both of these models much of the material that forms the core (Fe + FeS in the cold-accretion model and Fe in the condensation model) is already near the center of the Earth. In alternative proposals the core-forming material is either distributed evenly throughout the mantle or at the surface of the accreting planet.

One important boundary condition for the formation of the Earth is that the outer core, now and probably 3 Ga ago, be molten. If we take the temperature in the mantle to be 1880°C at 620-kilometer depth and assume an adiabatic temperature gradient of 0.5°C per kilometer from that depth to the core-mantle boundary, plus 1000° across D", the temperature at the top of the core is about 4000°C, which is of the order of the melting temperature of iron at core pressures. For the condensation model the temperature of the primitive iron core is 1150° to 1350°C and increases to 2300° to 2500°C, owing to adiabatic compression, as the planet assumes its present size. These temperatures can be compared with estimates of about 7000°C for the melting temperature of pure iron at the boundary of the inner core. These temperatures are all subject to great uncertainty, but it appears difficult with the condensation model to have a molten iron core. It is even more difficult to raise the central part of Earth above the melting temperature of pure iron in the cold-accretion model. It appears that the extra component in the core must also serve to decrease its melting temperature. This is easily accomplished with sulfur and, possibly, with oxygen as well, particularly if the core behaves as a eutectic system. An Fe-O-S core probably melts at a lower temperature than a two-component core. The presence of siderophiles in the upper mantle and water at the surface also provides constraints on the mechanism and timing of core formation.

The eutectic temperature for the system Fe-FeS is 990°C and is remarkably insensitive to pressure up to at least 30 kbar (Usselman, 1975). The eutectic composition is 31 percent by weight of sulfur at 1 atmosphere and 27 percent by weight of sulfur at 30 kbar. In an Earth of meteoritic composition, a sulfur-rich iron liquid would be the first melt to be formed. Core formation could proceed under these conditions at a temperature some 600°C lower than would be required to initiate melting in pure iron. If the Earth accreted from cold particles under conditions of radiative equilibrium, the temperatures would be highest in the upper mantle, which is where melting would commence. In most plausible thermal history calculations, the melting point of iron increases with depth much more rapidly than the actual temperatures, and any sinking molten iron will refreeze. Unless the Earth was very hot during most of the accretional process, it is difficult to get the iron to the center of the planet. This difficulty does not occur for the iron-sulfur model of the core, since gravitational accretion energy, adiabatic compression, and radioactive heating bring the temperatures throughout most of the Earth above 1000°C early in its history and, probably, during most of

the accretional process. Since the eutectic temperature increases only slowly with pressure, core formation is self-sustaining. The increase of gravitational energy due to core formation leads to a temperature rise of about 1600° to 2000°C throughout the Earth (Birch, 1965). Such temperatures would be adequate to melt the rest of the iron in the mantle, which would drain into the core, and to cause extensive melting of silicates and differentiation of the crust and upper mantle. The short time scale of accretion (or giant impacts) required by Hanks and Anderson (1969) to form a molten iron core within the first 10<sup>9</sup> years of earth history is not required for a sulfur-rich core.

Even if most of the light element in the core is oxygen, as seems probable, the above suggests that some sulfur will be present. This would serve to decrease the melting point of the core and would help explain the excessive depletion of sulfur in the crust-mantle system. An Fe-O-S core is likely, but there are few data on this system at high pressure.

One of the early criticisms of inhomogeneous accretion for the Earth was that it did not seem possible to melt the initial metallic core after it was accreted and buried. Attempts to do this by the thermal history of accretion were unsatisfactory. It was pointed out that the condensation sequence predicts a refractory calcium-aluminum-rich initial condensate before the condensation of metallic iron (Anderson and Hanks, 1972). Therefore, there would be calcium-aluminum-rich silicates accreting with the metallic iron, and both the theoretical calculations and measurements of the Allende inclusions show this initial condensate to be enriched in uranium and thorium as well as other heavy metals. Therefore, the inhomogeneous accretion model also predicts a long-lived heat source within the initial core. The short-lived nuclide aluminum-26, if present, could dominate the early thermal history and be responsible for melting of the core. Although uranium and thorium are unlikely to be in the present core, they may be concentrated at the core-mantle boundary, if held in sufficiently dense silicates.

The extensive melting of the Earth by giant planet-sized impactors would remove many of the problems addressed above.

How might the core of Mars have formed? The average temperature in Mars below some 600 km is about 1500°C if the planet accreted in  $3 \times 10^5$  years (Hanks and Anderson, 1969), which is well above the Fe-FeS eutectic temperature. The liquidus for an FeS-rich FeS-Fe system is 1100°C at atmospheric pressure, 1600°C at 60 kbar, and about 3400°C at the center of Mars. Thus, most of the deep interior of Mars is closer to the eutectic temperature than to the liquidus temperature, and only a fraction of the planet's complement of iron and FeS will be molten. For a carbonaceous chondrite I composition, the total FeS plus iron content is 24 percent by weight. The core of a small planet of that composition would therefore be richer in sulfur than

the core of the Earth, and much FeS would be left in the mantle. The small size of the martian core and the small size of the planet suppress the importance of gravitational heating due to core formation. If the density of the martian core is bracketed by the density of Earth's core and by the density of the eutectic mix in the system Fe-FeS at 30 kbar, then the core can be 9 to 15 percent, by mass, of the planet. A satisfactory model for Mars can be obtained by melting most of the FeS and some of the iron in ordinary chondrites (Anderson, 1971). Most of the arguments for putting substantial amounts of oxygen into the core do not apply to Mars.

## MINERALOGY OF THE MANTLE

The composition of the mantle is conventionally given in terms of oxides (SiO<sub>2</sub>, MgO, Al<sub>2</sub>O<sub>3</sub> and so forth), and these oxides are the building blocks of mantle minerals. The main minerals in the upper mantle are olivine, orthopyroxene, clinopyroxene and garnet with minor ilmenite and chromite. Clinopyroxene is basically a solid solution of CaMgSiO<sub>3</sub> (diopside) and NaAlSi<sub>2</sub>O<sub>6</sub> (jadeite). Al<sub>2</sub>O<sub>3</sub> is mainly in garnet, clinopyroxene and spinel, MgAl<sub>2</sub>O<sub>4</sub>. These minerals are only stable over a limited pressure and temperature range. The mineralogy of the mantle therefore changes with depth due to solid-solid phase changes. The elastic properties and density of the mantle are primarily controlled by the proportions of the above minerals or their high-pressure equivalents. To a lesser extent the physical properties depend on the compositions of the individual minerals.

Several cosmochemical estimates of mineralogy are given in Table 2-4. These tend to be less rich in olivine than estimates of the composition of the upper mantle (column 4). The mantle is unlikely to be uniform in composition, and therefore the mineralogy will change with depth, not only because of solid-solid phase changes, but also because of intrinsic chemical differences.

Although Al<sub>2</sub>O<sub>3</sub>, CaO and Na<sub>2</sub>O are minor constituents of the average mantle, their presence changes the mineralogy, and this in turn affects the physical properties such as the seismic velocities. The effect on density can result in chemical stratification of the mantle and concentration of these elements into certain layers. They also influence the melting point and tend to be concentrated in melts.

Olivine is an essential component in most groups of meteorites except the irons. Pallasites are composed of nickel-iron and olivine. Olivine is a major constituent in all the chondrites except the enstatite chondrites and some of the carbonaceous chondrites, and it is present in some achondrites. For these reasons olivine is usually considered to be the major constituent of the mantle. The olivines in

**TABLE 2-4**  
Mineralogy of Mantle

Species	Whole — Mantle Models			Upper Mantle
	(1)	(2)	(3)	(4)
Olivine	47.2	36.5	37.8	51.4
Orthopyroxene	28.3	33.7	33.2	25.6
Clinopyroxene	12.7	14.6	11.8	11.0
Jadeite	9.8	2.2	1.8	0.65
Ilmenite	0.2	0.5	0.24	0.57
Garnet	1.53	11.6	14.2	9.6
Chromite	0.0	1.6	0.94	0.44

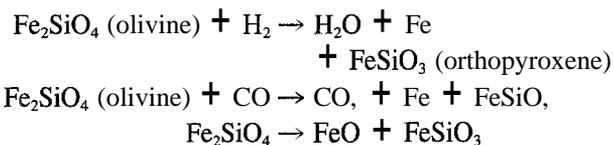
(1) Equilibrium condensation (BVP, 1980).

(2) Cosmochemical model (Ganapathy and Anders 1974).

(3) Cosmochemical model (Morgan and Anders, 1980).

(4) Pyrolite (Ringwood, 1977).

meteorites, however, are generally much richer in Fe<sub>2</sub>SiO<sub>4</sub> than mantle olivines. The olivine compositions in chondrites generally lie in the range 19 to 24 mole percent Fe<sub>2</sub>SiO<sub>4</sub> (fayalite or Fa) but range up to Fa<sub>100</sub>. The achondrites have even higher fayalite contents in the olivines. The seismic velocities in the mantle would be too low and the densities would be too high if the olivine were so iron rich. Mantle olivines are typically Fa<sub>100</sub>. Removal of iron from meteoritic olivine, either as Fe or FeO, would decrease the olivine content of the mantle:



Earth models based on cosmic abundances, with Mg/Si approximately 1 (molar), give relatively low total olivine contents. For example, the model of Morgan and Anders (1980) yields 37.8 mole percent olivine. Since the upper mantle appears to be olivine rich, this results in an even more olivine-poor transition region and lower mantle. It is usually assumed, however, that the basaltic fraction of the Earth is still mostly dispersed throughout the mantle. This is the assumption behind the pyrolite model (Ringwood, 1977) discussed in Chapter 8. Basalts were probably liberated during accretion of the Earth and concentrated in the upper mantle. Efficient remixing or rehomogenization can be ruled out with the isotopic evidence for ancient isolated reservoirs.

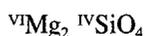
A given mineral is stable only over a restricted range of temperature and pressure. As pressure is increased, the atoms rearrange themselves, and ultimately a new configuration of atoms is energetically favorable, usually with a denser packing (see Chapter 16). Some of the phases of common mantle minerals are described below. The formulas and cation coordinations are given in Table 2-5.

**TABLE 2-5**  
Formulas and Coordination of Mantle Minerals

Formula	Species
${}^{\text{IV}}\text{Mg} {}^{\text{VI}}\text{Al}_2\text{O}_4$	Spinel
${}^{\text{VI}}\text{Mg} {}^{\text{IV}}\text{SiO}_3$	Enstatite
${}^{\text{VIII}}\text{Mg}_3 [{}^{\text{VI}}\text{Al}]_2 {}^{\text{IV}}\text{Si}_3\text{O}_{12}$	Pyrope
${}^{\text{VIII}}\text{Mg}_3 [{}^{\text{VI}}\text{Mg} {}^{\text{VI}}\text{Si}] {}^{\text{IV}}\text{Si}_3\text{O}_{12}$	Majorite
${}^{\text{VI}}\text{Mg} {}^{\text{VI}}\text{SiO}_3$	"Ilmenite"
${}^{\text{VIII-XII}}\text{Mg} {}^{\text{VI}}\text{SiO}_3$	"Perovskite"
${}^{\text{VI}}\text{MgO}$	Periclase
${}^{\text{VI}}\text{Mg}_2 {}^{\text{IV}}\text{SiO}_4$	Olivine
${}^{\text{VI}}\text{Mg}_2 {}^{\text{IV}}\text{SiO}_4$	$\beta$ -spinel
${}^{\text{VI}}\text{Mg}_2 {}^{\text{IV}}\text{SiO}_4$	y-spinel
${}^{\text{VIII}}\text{Ca} {}^{\text{VI}}\text{Mg} {}^{\text{IV}}\text{Si}_2\text{O}_6$	Diopside
${}^{\text{VIII}}\text{Na} {}^{\text{VI}}\text{Al} {}^{\text{IV}}\text{Si}_2\text{O}_6$	Jadeite
${}^{\text{VI}}\text{SiO}_2$	Stishovite

### (Mg,Fe)<sub>2</sub>SiO<sub>4</sub>–Olivine

The olivine structure is based on a nearly hexagonal-closest packing of oxygen ions:



The  $\text{MgO}_6$  octahedra are not all equivalent. Each  $\text{Mg}(1)\text{O}_6$  octahedron shares four edges with adjacent Mg-octahedra and two edges with Si-tetrahedra. Each  $\text{Mg}(2)\text{O}_6$  octahedron shares two edges with adjacent Mg-octahedra and one edge with a Si-tetrahedron. The  $\text{Mg}(1)\text{O}_6$  octahedra are linked to each other by sharing edges, making a chain along the *c* axis. The  $\text{Mg}(2)\text{O}_6$  octahedra are larger and more distorted than the  $\text{Mg}(1)\text{O}_6$  octahedra. Olivine is formed of separate  $(\text{SiO}_4)^{4-}$  groups. The four oxygen atoms surrounding a silicon atom are not linked to any other silicon atom. Olivine is a very anisotropic mineral and apparently is easily aligned by recrystallization or flow, making peridotites anisotropic as well.

### (Mg,Fe)<sub>2</sub>SiO<sub>4</sub>–Spinel

In 1936 Bernal suggested that common olivine (a) might transform under high pressure to a polymorph possessing the spinel ( ${}^{\text{VI}}\text{Al}_2 {}^{\text{IV}}\text{MgO}_4$ ) structure with an increase of density of about nine percent. Jeffreys (1937) adopted this suggestion to explain the rapid increase in seismic velocity at a depth of 400 km—"the 20° discontinuity" so named because the effect is seen in an increase of apparent velocity of seismic waves at an arc distance of 20° from a seismic event. Many  $\text{A}_2\text{BO}_4$  compounds have now been transformed to the spinel structure at high pressure, and natural olivine does indeed transform at about the right temperature and pressure to explain the 400-km discontinuity. Although  $\text{Fe}_2\text{SiO}_4$  transforms directly to the spinel structure, olivines having high  $\text{MgO}$  contents occur in two modifications re-

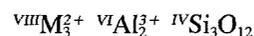
lated to the spinel structure, p-spinel and y-spinel. Both of these have much higher elastic wave velocities than appropriate for the mantle just below 400 km, so there must be other components in the mantle that dilute the effect of the  $\alpha$ - $\beta$  phase change. Although normal spinel is a structural analog to silicate spinel, it should be noted that Mg occurs in 4-coordination in  $\text{MgAl}_2\text{O}_4$  and in 6-coordination in  $\beta$ - and y-spinel with a consequent change in the Mg–O distance and the elastic properties.  ${}^{\text{IV}}\text{Mg}$  is an unusual coordination for Mg, and the elastic properties and their derivatives cannot be assumed to be similar for  ${}^{\text{IV}}\text{MgO}$  and  ${}^{\text{VI}}\text{MgO}$  compounds. In particular, normal spinel has an unusually low pressure derivative of the rigidity, a property shared by other 4-coordinated compounds but not  ${}^{\text{VI}}\text{MgO}$ -bearing compounds.

The spinel structure consists of an approximate cubic close packing of oxygen anions. In the *normal* spinel structure, exhibited by many  $\text{A}_2^{3+}\text{B}^{4+}\text{O}_4$  compounds, we have  ${}^{\text{VI}}\text{A}$ ,  ${}^{\text{IV}}\text{BO}_4$ . In *inverse* spinels one of the  $\text{A}^{2+}$  ions occupies the tetrahedral site, and the octahedral sites are shared by  $\text{A}^{2+}$  and  $\text{B}^{4+}$  ions. Intermediate configurations are possible.

$\beta$ -Spinel is crystallographically orthorhombic, but the oxygen atoms are in approximate cubic close packing. It is sometimes referred to as the  $\beta$ -phase or the distorted or modified spinel structure. It is approximately 7% percent denser than  $\alpha$ -phase, or olivine. The arrangements of the A and B ions are different from those in spinel. Two  $\text{SiO}_4$  tetrahedra, which would be isolated in the spinel structure, share one of their oxygen atoms, resulting in a  $\text{Si}_2\text{O}_7$  group and an oxygen not bonded to any Si.  $\beta$ -Spinel is an elastically anisotropic mineral. The transformation to y-spinel results in a density increase of about 3½ percent with no overall change in coordination.

### Garnets

Garnets are cubic minerals of composition

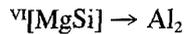


where M is Mg,  $\text{Fe}^{2+}$  or Ca for the common garnets, but can be almost any 2+ element. Some natural garnets have  $\text{Cr}^{3+}$  or  $\text{Fe}^{3+}$  instead of  $\text{Al}^{3+}$ . Garnets are stable over an enormous pressure range, reflecting their close packing and stable cubic structure. They are probably present over most of the upper mantle and, perhaps, into the lower mantle. Furthermore, they dissolve pyroxene at high pressure, so their volume fraction expands with pressure. Garnets are the densest common upper-mantle mineral, and therefore eclogites and fertile (undepleted) peridotites are denser than basalt-depleted peridotites or harzburgites. On the other hand, they are less dense than other phases that are stable at the base of the transition region. Therefore, eclogite can become less dense than other rock types at great depth. Garnet has a low melting point and is eliminated from perido-

tites in the upper mantle at small degrees of partial melting. The large density change associated with partial melting of a garnet-bearing rock is probably one of the most important sources of buoyancy in the mantle. Garnets are elastically isotropic.

### MgSiO<sub>3</sub>–Majorite

Pyroxene enters the garnet structure at high pressure via the substitution



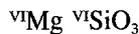
The mineral



is known as **majorite**, and it exhibits a wide range of solubility in garnet. Note that one-fourth of the Si atoms are in 6-coordination. However, the elastic properties of MgO plus SiO<sub>2</sub>-stishovite are similar to Al<sub>2</sub>O<sub>3</sub>, so we expect the elastic properties of majorite to be similar to garnet. Majorite also has a density similar to that of garnet.

### MgSiO<sub>3</sub>–Ilmenite

The structural formula of "ilmenite," the hexagonal high-pressure form of enstatite, is



making it isostructural with true ilmenite (FeTiO<sub>3</sub>) and corundum (Al<sub>2</sub>O<sub>3</sub>). Silicon is also 6-coordinated in stishovite and perovskite, dense high-elasticity phases. Because of similarity in ionic radii, we expect that extensive substitution of <sup>VI</sup>Al<sub>2</sub> for <sup>VI</sup>(MgSi) is possible. MgSiO<sub>3</sub>–ilmenite is a platy mineral, suggesting that it may be easily oriented in the mantle. It is also a very anisotropic mineral, rivaling olivine in its elastic anisotropy.

The arrangement of oxygen atoms is based on a distorted hexagonal closest packing having a wide range of O–O distances.

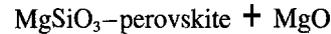
"Ilmenite" is a stable phase at the base of the transition region and the top of the lower mantle. The transformations  $\gamma$ -spinel plus stishovite, or **majorite**, to "ilmenite" and "ilmenite" to "perovskite" (see below) may be responsible for the high gradients of seismic velocity found in this depth range. "Ilmenite" appears to be a lower temperature phase than **majorite** and appears to occur mainly in deeply subducted slabs, contributing to the high velocities found in the vicinity of deep-focus earthquakes.

### MgSiO<sub>3</sub>–Perovskite

The structural formula of the high-pressure phase of enstatite, or "perovskite," is



It appears to be stable throughout the lower mantle and is therefore the most abundant mineral in the mantle. Related structures are expected for CaSiO<sub>3</sub> and garnet at high pressure. Olivine also transforms to a "perovskite"-bearing assemblage



MgSiO<sub>3</sub>–perovskite is 3.1 percent denser than the isochemical mixture stishovite plus periclase. There are a variety of Mg–O and Si–O distances. The mean Si–O distance is similar to that of stishovite.

The structure is orthorhombic and represents a distortion from ideal cubic perovskite. In the ideal cubic perovskite the smaller cation has a coordination number of 6, whereas the larger cation is surrounded by 12 oxygens. In MgSiO<sub>3</sub>–perovskite the 12 Mg–O distances are divided into 4 short distances, 4 fairly long distances and 4 intermediate distances, giving an average distance appropriate for a mean coordination number of 8.

## MOON

The Moon is one of the more obvious of our neighbors in space and is certainly the most accessible. In spite of intensive analysis and probing by virtually every conceivable chemical and physical technique, the maneuvering room for speculation on lunar origin was scarcely diminished as a result of the Apollo and Luna programs. This was not primarily due to lack of information but to the unexpected and confusing nature of the new data, most of which were open to multiple interpretations.

### Composition and Structure

The Moon's unique characteristics have become even more unique as a result of lunar exploration. It is like no other body in the solar system that we know about. It has long been known that the Moon is deficient in iron (in comparison with the proportions of iron in the Earth and the other terrestrial planets). It is also deficient in all elements and compounds more volatile than iron.

The density of the Moon is considerably less than that of the other terrestrial planets, even when allowance is made for pressure. Venus, Earth and Mars contain about 30 percent iron, which is consistent with the composition of stony meteorites and the nonvolatile components of the Sun. They therefore fit into any scheme that has them evolve from solar material. Mercury is even better endowed with iron. Because iron is the major dense element occurring in the Sun, and presumably in the preplanetary solar nebula, the Moon is clearly depleted in iron. Many theories of lunar origin have been based on this fact, and numerous attempts have been made to explain how iron can be separated from other elements and compounds. Density, mag-

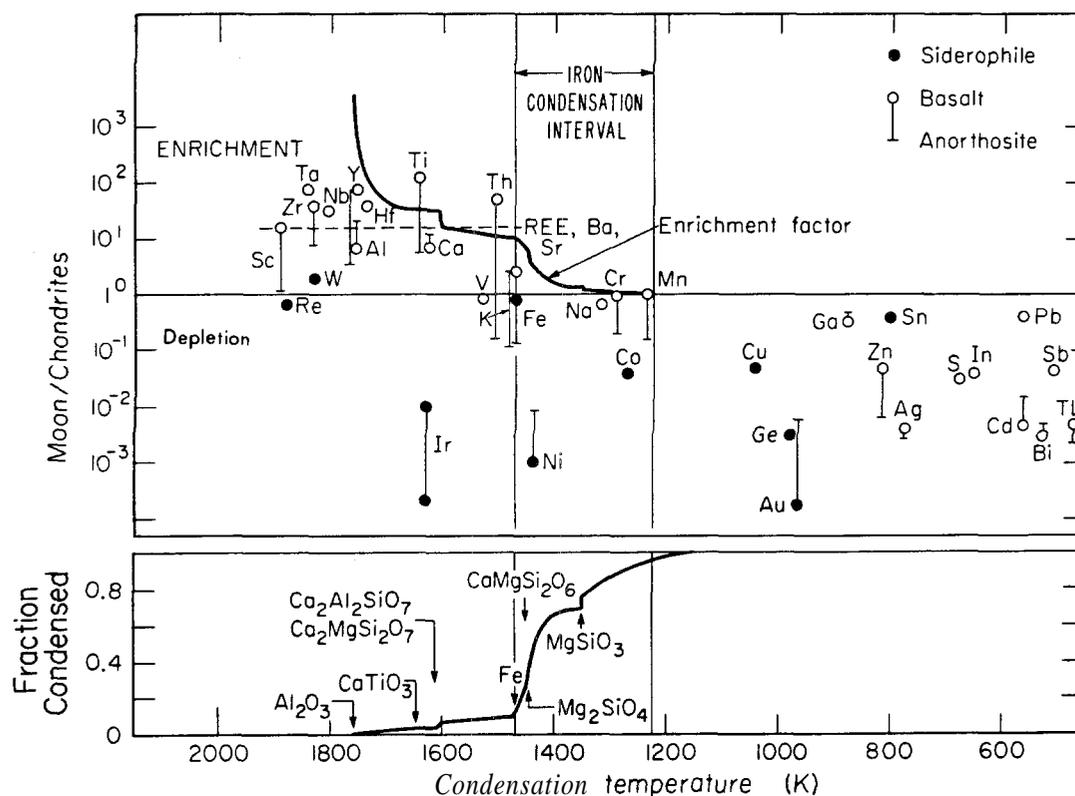
netic properties and ductility have all been invoked to rationalize why iron should behave differently than silicates in early solar-system processes.

Once samples were returned from the Moon, however, it became clear that the Moon was deficient not only in iron but also in a number of other elements as well (Figure 2-2). A common characteristic of many of these elements and their compounds is volatility. The returned samples showed that the Moon is depleted in compounds more volatile than iron. Calcium, aluminum and titanium are the major elements involved in high-temperature condensation processes in the solar nebula; minor refractory elements include barium, strontium, uranium, thorium and the rare-earth elements. The Moon is enriched in all these elements, and we are now sure that more than iron-silicate separation must be involved in lunar origin.

The abundance of titanium in the returned lunar samples was one of the first surprises of the Apollo program. Titanium is not exactly rare on Earth, but it is usually considered a minor or trace element. The first samples returned from the Moon contained 10 percent of titanium-rich compounds. The surface samples were also remarkably depleted in such volatile elements as sodium, potassium, rubidium,

and other substances that, from terrestrial and laboratory experience, we would expect to find concentrated in the crust, such as water and sulfur. The refractory trace elements—such as barium, uranium and the rare-earth elements—are concentrated in lunar surface material to an extent several orders of magnitude over that expected on the basis of cosmic or terrestrial abundances.

Some of these elements, such as uranium, thorium, strontium and barium, are large-ion elements, and one would expect them to be concentrated in melts that would be intruded or extruded near the surface. However, other volatile large-ion elements such as sodium and rubidium are clearly deficient, in most cases, by at least several orders of magnitude from that expected from cosmic abundances. The enrichment of refractory elements in the surface rocks is so pronounced that several geochemists proposed that refractory compounds were brought to the Moon's surface in great quantity in the later stages of accretion. The reason behind these suggestions was the belief that the Moon, overall, must resemble terrestrial, meteoritic or solar material and that it was unlikely that the whole Moon could be enriched in refractories. In these theories the volatile-rich materials must be concentrated toward the interior. In a



**FIGURE 2-2**

Chondrite-normalized lunar abundances as a function of condensation temperature of the element or the first condensing compound containing this element at  $10^{-3}$  atm total pressure. Ratios greater than unity indicate enrichment and less than unity indicate depletion (after Anderson, 1973b).

cooling-nebula model of planetary formation, the refractories condense before the volatiles, and it was therefore implied that the Moon was made inside out!

However, it now appears that the depletion of iron and volatiles can be taken at face value and that the whole Moon is deficient in elements and compounds more volatile than iron. Petrological considerations show that not only the surface rocks but also their igneous source regions, deep in the Moon, are also depleted in volatiles. The Moon is probably enriched in calcium, aluminum, titanium and the refractory trace elements throughout. This composition would explain the mean density of the Moon and the high heat flow, and it would help to explain why the Moon melted and differentiated very rapidly. However, if the Moon's origin is linked to a high-energy impact on the Earth, we have a ready explanation both for an initially molten Moon and a silicate and refractory-rich Moon.

In view of the abundant geological evidence that the surface rocks resulted from melting processes in the interior, it was no surprise that the geophysical evidence indicated that the Moon has a low-density crust. Its great age, as measured by geochemical techniques, and great thickness, as required by the physical evidence, were, however, unexpected. These are important boundary conditions on the origin and composition of the Moon and, probably, the Earth as well.

The Moon's principal moments of inertia indicate that crustal thickness varies from about 40 km at the poles to more than 150 km on the lunar farside. Large variations in crustal thickness are also required to satisfy gravity data and the noncoincidence of the centers of mass and figure of the Moon. The present orientation of the Moon and the restriction of basalt-filled maria to the Earth-facing hemisphere are undoubtedly the result of this asymmetry in crustal thickness.

The maria are remarkably smooth and level; slopes of less than one-tenth of a degree persist for hundreds of kilometers, and topographic excursions from the mean are generally less than 150 m. By contrast, elevation differences in the highlands are commonly greater than 3 km. The mean altitude of the terrace, or highlands, above maria is also about 3 km.

The center of mass is displaced toward the Earth and slightly toward the east by about 2 km. This gross asymmetry of the Moon has long been known from consideration of the principal moments of inertia. The differences between the principal moments of inertia are more than an order of magnitude greater than can be accounted for by a simple homogeneous body, rotating and stretched by Earth tides. The simplest interpretation is in terms of a crust of highly variable thickness, an interpretation supported by nearside gravity results.

Asymmetry is not a unique characteristic of the Moon; the asymmetric distribution of continents and oceans on the Earth is well known, and Mars, likewise, is very asymmet-

ric in both its topography and its gravity field. Large-scale convection associated with early gravitational differentiation could lead to the observed asymmetries and may be one common characteristic of all the terrestrial planets.

In the case of the Earth and the Moon, and probably for Mars as well, the physical asymmetry correlates well with, and is probably the result of, chemical asymmetry. The lunar highlands are dominantly plagioclase-feldspar-rich rocks with densities considerably less than the nearside mare basalts and the mean density of the Moon. These feldspars crystallize at higher temperatures than basalt and can therefore be expected to float to the surface of their parent liquid. The residual liquids would likely be the source region of the mare basalts, which erupt to the surface later.

This scenario explains not only the physical measurements but also some subtle details of the chemistry. Large-ion refractory elements are preferentially retained by the liquid, and therefore such elements as barium, strontium, uranium and thorium would be concentrated in the last liquid to crystallize. These elements are concentrated in the lunar-mare basalts by several orders of magnitude over the highland plagioclase-rich material, with the notable exception of europium, which is retained by plagioclase. Compared to the other rare-earth elements, europium is depleted in basalts and enriched in anorthosites. The "europium anomaly" was one of the early mysteries of the lunar sample-return program and implied that plagioclase was abundant somewhere on the Moon. The predicted material was later found in the highlands.

Seismic activity of the Moon is much less than the Earth's, both in numbers of quakes and their size, or magnitude. Their times of occurrence appear to correlate with tidal stresses caused by the varying distance between the Moon and the Earth. Compared with the Earth, they seem to occur at great depth, about half the lunar radius. Both the age-dating evidence and the seismic data indicate that the Moon today is a relatively inactive body. This conclusion is consistent with the absence of obvious tectonic activity and with the low level of stresses in the lunar interior implied by gravity and moment-of-inertia data.

The outer kilometer has extremely low seismic velocities, less than 1 km/s. This value is more appropriate for rubble than consolidated rock. Velocities increase to 4 km/s at 1 km depth and 6 km/s at 20 km; the lower velocity is appropriate for consolidated rubble or extensively fractured igneous rock, such as basalt. The increase of velocity with depth is the result of consolidation and crack closure. The 6-km/s velocity is consistent with laboratory measurements on returned samples of lunar basalt.

At 20 km depth the velocity increases abruptly to about 6.7 km/s, and it remains relatively constant to 60 km depth. The constancy of velocity means that most cracks have been eliminated and also that the effects of temperature and pressure are either small or mutually canceling. In this region the velocities can be matched by anorthositic gabbro, a pla-

gioclase-rich rock type that has a low iron content and relatively low density (about  $2.9 \text{ g/cm}^3$ ). This layer may be similar in composition to the lunar highlands.

At 60 km the velocity jumps, at least locally, to about 9 km/s. When all the seismic data are considered together, we may find that this layer is very thin (less than 40 km) or it occurs only locally, or both. Perhaps it occurs only as pods or lenses under maria basins or only under mascon basins; at the moment we have no way of telling. In any event, such high velocities are unexpected and are unusual by any standards. They may also be fictitious, because seismic waves refracted by dipping interfaces can give apparent velocities slower or faster than real velocities. A velocity of 9 km/s is much greater than the 8 km/s velocity typical of the Earth's upper mantle and of rocks thought to be common in the upper lunar mantle. Only a few minerals, exotic by terrestrial standards, have such high velocities. These include spinel ( $\text{MgAl}_2\text{O}_4$ ), corundum ( $\text{Al}_2\text{O}_3$ ), kyanite ( $\text{Al}_2\text{SiO}_5$ ) and Ca-rich garnet ( $\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$ ). These are all calcium- and/or aluminum-rich minerals and occur as the dense residual crystals when a Ca-Al-rich liquid partially solidifies. The 9-km/s layer may therefore be related petrologically to the overlying crustal layer.

The apparent seismic velocity at greater depth is only 7.7 km/s—intermediate between the velocities we usually associate with the crust and the mantle. This velocity continues to a depth of at least 150 km (the deepest depth of penetration of seismic energy from artificial impacts to date) and is appropriate for pyroxene or for plagioclase-rich (meaning crustal) rocks. In the latter case we would still be monitoring the crust and therefore would have only a lower bound, 150 km, on its thickness. It should be recalled that the crust is thicker on the farside. Even if the crust is only 60 km thick, the conventional interpretation of the seismic results, it is much thicker than the average terrestrial crust, particularly in relative terms; this great thickness indicates that the Moon is extensively differentiated. In combination with the age data this means that the Moon was extensively melted early in its history. The source of this early heat is a matter of some controversy. Forming the Moon from debris resulting from a giant impact on Earth is the current "best idea."

Evidence for the constitution of the deeper interior is very sketchy. Seismic shear waves apparently cannot pass efficiently below some 1000 km depth. This can be taken as tentative evidence for a hot, if not molten or partially molten, deep interior. One of the long-standing controversies regarding the Moon is whether its interior is hot or cold. The widespread occurrence of basalt certainly indicates that it was at least partially molten early in its history. Conduction alone is only efficient in lowering the internal temperatures of the outer 300 km.

The lunar crust represents about 10 percent of the lunar volume. The mare basalts constitute about one percent of the crust. The average seismic velocities are about 7.7 km/s

for P-waves and 4.45 km/s for S-waves down to about 1000 km. There may be a minor discontinuity at a depth of 400–490 km in the Moon, with a drop in velocity below that depth range. Most of the moonquakes occur between 800 and 1000 km depth. There is some evidence for increased attenuation of seismic waves below 1100 km (Goins et al, 1981). The geophysical data place a weak upper bound of about 500 km radius on the size of a dense, conductive core. The depletion of iridium in lunar rocks is consistent with either removal by iron draining to the core or an overall depletion of siderophiles. The moment of inertia, 0.391, is consistent with a density increase in the interior over and above the crust-mantle density increase. This could be due to a large eclogite core or a small iron-rich core.

The anorthositic highlands are probably also the result of igneous and differentiation processes. But their emplacement mechanism would be plagioclase flotation, resulting in a thick insulating blanket for the remaining liquid. The age of maria material indicates that melts still existed at moderate depths for more than  $10^9$  years after creation of the plagioclase-rich highlands.

It is quite possible that the mare-forming igneous episodes were a result of thermal, tidal and impact stresses, all of which were intense in the earliest history of the Moon. Igneous activity may have ceased when stresses were no longer adequate to breach the thick lithosphere. If this is true, then the Moon below some 300 to 1000 km may still be partially molten. A lithosphere of this thickness could easily support the stresses implied by the nonhydrostatic shape of the Moon and the presence of mass concentrations.

The nonequilibrium shape of the Moon, the offset of the center of mass from the center of figure and the presence of large surface concentrations of mass (mascons) have been used as arguments that the Moon is a cold, strong body. However, when viewed more carefully, all of this evidence suggests just the contrary. The stresses required to support the nonequilibrium shape are only some tens of bars, and a relatively thin, strong, outer layer would suffice to support these stresses. The Earth, by contrast, supports stress differences of hundreds of bars, and stresses at kilobar levels are required to break rocks in the laboratory. Thus, taken at face value, the lunar data suggest that the Moon is a hot weak body. This conclusion is consistent with the lunar heat-flow values, the low level of seismic stresses and the high radioactivity inferred for the interior.

## Early History

If a small body such as the Moon formed hot, it is likely that the other planets also formed hot and that they too are differentiated. This is an important boundary condition on the vaguely understood processes of planetary formation. The accretion of planetesimals into planets must have been extremely rapid, at least on the time scale of cooling of the

nebula. Otherwise the gravitational energy of accretion would be conducted to the surface and radiated away. A rapid time scale of accretion is consistent with modern theories of planetary accretion in a collapsing gas-dust cloud. High initial temperatures are also consistent with the hypothesis of accretion during condensation, explained earlier in this chapter.

How can we explain the early high-temperature history of the Moon and at the same time its rapid death as an active body? Prior to lunar exploration the preferred theory for the formation of the planets involved the gradual accretion of cold particles with subsequent heating, melting and differentiation resulting from decay of radioactive elements. Because the first  $10^9$  years of terrestrial history are inaccessible to us, there was no compelling reason to suppose that the Earth was initially hot. Some indirect clues, however, point to a high-temperature origin. The oldest terrestrial rocks are clearly the result of igneous processes, and they contain evidence that the Earth had a magnetic field—hence a molten core—when they crystallized. Many ancient meteorites also have clearly gone through a very early high-temperature event or are the result of magmatic processes, which must have taken place in a body much smaller than the Earth.

Age dating of lunar material indicates that most of the differentiation of the Moon occurred in the first  $10^9$  years of its existence. For the Earth nearly all of this early record is erased by later activity, but the Moon has been remarkably quiescent for the last  $3 \times 10^9$  years. This is true not only for internal processes but also for the external bombardment processes that are mainly responsible for the surface features of the Moon. Both the internal structure and exterior morphology were apparently the result of an extensive early history of activity.

Many sources have been proposed as early heating processes. These include short-lived radioactive elements, solar radiation, high temperatures associated with the early nebula, and energy of gravitational accretion. For any of the latter three processes to be effective, the bodies in the solar system must have assembled very quickly. Gravitational collapse of a gas-dust cloud is an attractive way both to localize protoplanetary material and to have initially high temperatures. The main problem is to be able to build planetary nuclei to a size big enough that they can effectively scavenge the remaining material in their vicinity.

Some of the most obvious facts about the Moon may also be among the most relevant for our attempts to understand its origin. It is small, light and is close to a more massive body. These simple facts may all be related, for it is unlikely that preplanetary processes resulted in a single nucleus for each of the present planets. The nebula probably evolved from a disk to a series of rings; each ring in turn collapsed to a series of local gas-dust concentrations that collapsed further to form protoplanets, the building blocks of the planets.

Variations in the eccentricities and inclinations of the

orbits of protoplanets at this stage of development ensure that they periodically approach each other; encounter velocities between bodies in a ring are low, and concentration rather than dispersal is the natural result. It is not difficult to believe that eventually one body will predominate; the remaining bodies will impact, orbit temporarily before impact, or orbit permanently. The scenario repeats for those bodies in orbit about the primary body. The largest nucleus, the Earth in this case, grows at the expense of the smaller particles, and if all this is happening in a cooling nebula it will inherit most of the later, lower-temperature condensates.

In a cooling nebula of solar composition, the first compounds to condense are calcium-, aluminum- and titanium-rich oxides, silicates and titanates. These compounds comprise approximately 6 percent of the nonvolatile composition of the nebula, which is roughly everything but hydrogen, helium, carbon and that oxygen not tied up in the refractory compounds. Carbonaceous chondrites are usually taken as an approximation to the "nonvolatile" content of the nebula. Type I and II carbonaceous chondrites contain about 10 percent of unique white inclusions, composed primarily of exotic Ca-Al-rich minerals such as gehlenite, spinel and anorthosite. These inclusions are rich in barium, strontium, uranium and the rare-earth elements, and they have most of the properties that have been inferred for the Moon. They are also very ancient. This material may represent the most primitive material in the solar system and be the initial building blocks of the planets. At the very least they are analogs of the refractory material in the solar system.

Planetesimals formed at this point will be deficient in iron, magnesium, silicon, sulfur, sodium and potassium, all of which are still held in the gaseous phase. Solid particles will rapidly concentrate toward the median plane to accrete into refractory-rich planetary nuclei. While cooling of the nebula continues, iron and magnesium silicates condense; these are the most abundant constituents of meteorites and of the terrestrial planets. The largest body at any distance from the Sun will obtain the major share of these later condensates. In this scenario the Moon is one of the original smaller bodies that avoided impact with the Earth. It may also represent the coagulation of many smaller bodies that were trapped into Earth orbit from Earth-crossing solar orbits. In this hypothesis many of the satellites in the solar system may be more refractory than their primary bodies. The angular momentum of the Earth-Moon system is difficult to understand with this scenario. The Moon may also have formed by condensation of material vaporized from the Earth by a large impact.

## The Lunar Crust

The Moon is such a small body that both its major-element chemistry and its incompatible trace-element chemistry are affected by the crustal composition. The thick lunar high-

land crust (Taylor, 1982) is estimated to have 13 percent aluminum, 11 percent calcium and 3400 ppm titanium, which, combined with concentrations of trace elements and estimates of crustal thickness, immediately indicate that the Moon is both a refractory-rich body and an extremely well-differentiated body. The amount of aluminum in the highland crust may represent about 40 percent of the total lunar budget. This is in marked contrast to the Earth, where the amount of such major elements as aluminum and calcium in the **crust** is a trivial fraction of the total in the planet. On the other hand the amount of the very incompatible elements such as rubidium, uranium and thorium in the Earth's crust is a large fraction of the terrestrial inventory. This dichotomy between the behavior of major elements and incompatible trace elements can be understood by considering the effect of pressure on the crystallization behavior of calcium- and aluminum-rich phases. At low pressures these elements enter low-density phases such as plagioclase, which are then concentrated toward the surface. At higher pressures these elements enter denser phases such as **clinopyroxene** and garnet. At still higher pressures, equivalent to depths greater than about 300 km in the Earth, these phases react to form a dense garnet-like solid solution that is denser than such upper-mantle phases as olivine and pyroxene. Therefore, in the case of the Earth, much of the calcium and aluminum is buried at depth. The very incompatible elements, however, do not readily enter any of these phases, and they are concentrated in light melts. The higher pressures in the Earth's magma ocean and the slower cooling rates of the larger body account for the differences in the early histories of the Earth and Moon.

In the case of the Moon, the anorthositic component is due to the flotation of plagioclase during crystallization of the ocean. Later basalts are derived from cumulates or cumulus liquids trapped at depth, and the KREEP (K, REE, P rich) component represents the final residual melt. The isotopic data (Pb, Nd, Sr) require large-scale early differentiation and uniformity of the KREEP component. About 50 percent of the europium and potassium contents of the bulk Moon now reside in the highland crust, which is less than 9 percent of the mass of the Moon. The energy associated with the production of the highland crust is much greater than needed for the mare basalts, which are an order of magnitude less abundant. Estimates of the required thickness of the magma ocean are generally in excess of 200 km, and mass-balance calculations require that most or all of the Moon has experienced partial melting and melt extraction. A 200-km-thick magma ocean will freeze in about  $10^7$ – $10^8$  years. Other evidence in support of the magma ocean concept, or at least widespread and extensive melting, include: (1) the complementary highland and mare basalt trace-element patterns, particularly the europium anomaly; (2) the enrichment of incompatible elements in the crust and KREEP; (3) the isotopic uniformity of KREEP; and (4) the isotopic evidence for early differentiation of the mare basalt source region, which was complete by about 4.4 Ga.

There is thus a considerable body of evidence in support of large-scale and early, prior to 4.4 Ga, lunar differentiation (Taylor, 1982). The separation of crustal material **and** fractionation of trace elements is *so* extreme that the concept of a deep magma ocean plays a central role in theories of lunar evolution. The cooling and crystallization of such an ocean permits efficient separation of various density crystals and magmas and the trace elements that accompany these products of cooling. This concept does not require a continuous globally connected ocean that extends to the surface nor one that is even completely molten. The definition of magma extends to melt-crystal slurries, and oceans can be covered by a thin crust of light crystals. Such semantic niceties have resulted in unproductive quibbling about whether it is a magma ocean or a series of magma marshes, unconnected magma lakes, buried magma chambers or widespread partial melt zones in the mantle. I refer to all of these possibilities under the collective name "magma ocean," which is descriptive and serves to focus attention on the geochemical requirements for widespread and extensive partial melting of the Moon that leads to at least local accumulations of magma and cumulative crystal layers and efficient separation of certain elements. Part of the evidence for a magma ocean on the Moon is the thick anorthositic highland crust and the widespread occurrence of KREEP, an incompatible-element-rich material best interpreted as the final liquid dregs of a Moon-wide melt zone. The absence of an extensive early terrestrial anorthositic crust and the presumed absence of a counterpart to KREEP have kept the magma ocean concept from being adopted as a central principle in theories of the early evolution of the Earth. I show in Chapter 11, however, that a magma ocean is also quite likely for the Earth and probably the other terrestrial planets as well.

The gross physical properties of the Moon such as its mass and moment of inertia do not place significant constraints on the major-element abundances of the deep lunar interior, except that the Moon is clearly deficient in iron in relation to cosmic, chondritic, or terrestrial abundances. In particular, this kind of data cannot be used to place upper bounds on the CaO and  $Al_2O_3$  content. However, the surface evidence suggests that the Moon is also deficient in elements and compounds more volatile than iron. This led to the suggestion (Anderson, 1973a, b) that the interior of the Moon was also enriched in CaO and  $Al_2O_3$ . Such a body would be highly enriched in the refractory trace elements relative to chondritic abundances and upon melting or fractional crystallization would yield a plagioclase-pyroxene outer shell of the order of 250 km in thickness. A mixture of about 14 percent basalt and 86 percent anorthosite satisfies the bulk chemistry of this layer and eliminates the europium anomaly. These proportions are broadly consistent with the areal extent of the maria and the volumetric relationships implied by elevation differences, offset of center of mass, and the moments of inertia. **High-CaO-and- $Al_2O_3$**  peridotites have broad intermediate density fields, and

**TABLE 2-6**  
Crustal Compositions in the Moon and Earth

	Lunar Highland Crust	Lunar Mare Basalt	Terrestrial Continental Crust	Terrestrial Oceanic Crust
<b>Major Elements (percent)</b>				
Mg	4.10	3.91	2.11	4.64
Al	13.0	5.7	9.5	8.47
Si	21.0	21.6	27.1	23.1
Ca	11.3	8.4	5.36	8.08
Na	0.33	0.21	2.60	2.08
Fe	5.1	15.5	5.83	8.16
Ti	0.34	2.39	0.48	0.90
<b>Refractory Elements (ppm)</b>				
Sr	120	135	400	130
Y	13.4	41	22	32
Zr	63	115	100	80
Nb	4.5	7	11	2.2
Ba	66	70	350	25
La	5.3	6.8	19	3.7
Yb	1.4	4.6	2.2	5.1
Hf	1.4	3.9	3.0	2.5
Th	0.9	0.8	4.8	0.22
U	0.24	0.22	1.25	0.10
<b>Volatile Elements (ppm)</b>				
K	600	580	12,500	1250
Rb	1.7	1.1	42	2.2
Cs	0.07	0.04	1.7	0.03

Taylor (1982), Taylor and McLennan (1985).

the gabbro-eclogite transformation pressure increases with  $Al_2O_3$  content. Therefore the Moon can have a thick plagioclase-rich outer shell without violating its density. The seismic velocities in the lunar mantle can be compared with the range 7.5–7.6 km/s for terrestrial anorthosites with more than 95 volume percent plagioclase. A material composed of about 80 percent gabbroic anorthosite and about 20 percent Ca-rich pyroxene yields a velocity of about 7.7 km/s.

Table 2-6 gives comparisons between the crusts of the Moon and the Earth. In spite of the differences in size, the bulk composition and magmatic history of these two bodies, the products of differentiation, are remarkably similar. The lunar crust is less silicon rich and poorer in volatiles, probably reflecting the overall depletion of the Moon in volatile. The lunar highland crust and the mare basalts are both more similar to the terrestrial oceanic crust than to the continental crust. Depletion of the Moon in siderophiles and similarity of the Earth and Moon in oxygen isotopes are consistent with the Moon forming from the Earth's mantle, after separation of the core. A giant impact on Earth is the most recent variant of the "Moon came from the Earth" scenario. Earlier variants included tidal disruption and rotational instability.

## Origin of the Moon

The Moon has several unique attributes, in addition to its large size, which place constraints on its origin. The substantial prograde angular momentum of the Earth-Moon system seems to rule out co-accretion from a gas-dust cloud or from planetesimals. On the other hand oxygen-isotope data favor formation of both bodies from a similar cosmic pool of material. There are similarities in the refractory-element abundances between the mantles of Earth and Moon, but the Moon is extremely depleted in the volatile elements and iron compared to the Earth and most meteorites. The Moon is a highly differentiated body, so much so that extensive melting of all or most of the Moon and the presence of a deep magma ocean appear likely.

The small mass of the Moon indicates that the event or process that formed it was an energetic one; it was probably responsible for the extensive and early melting and depletion in volatiles. The angular difference between Earth's equator and the lunar orbital plane (about  $5^\circ$ ) suggests a role for large impacts in early Earth-Moon history.

The low density of the Moon restricts not only the size of an iron-rich core but also the total iron content. The maximum allowable size of the lunar core is one-fifteenth, by mass fraction, of the terrestrial core. The lunar mantle silicates must also be less iron rich than those in chondritic meteorites. The depletion of refractory siderophiles in the Moon suggests that the Moon was formed from material poor in iron, not just in volatiles.

Much attention has recently been focused on the role of giant impacts in early planetary evolution. Giant impactors, say 10 to 30 percent of the mass of a planet, can extensively melt the parent body, remove primitive atmospheres, and even melt, vaporize and remove large portions of the crust and upper mantle. They would also affect, and possibly control, the angular momentum of the impacted body. If even a modest fraction of the mass of a terrestrial planet, say 30 percent, is brought in by giant impactors, then the initial condition will be hot, just as if the whole planet accreted rapidly. A giant impact is, of course, an example of very rapid accretion.

A sufficiently energetic impactor, greater than the mass of Mars, would melt the crust and much of the mantle and throw enough material into orbit to form the Moon. The objection that such an event would be extremely rare is actually a point in its favor, since the Moon is unique. A single such event could explain the geochemical similarities and differences between the Earth and the Moon, the absence of an early terrestrial geological record, the angular momentum problem, the lunar depletion of volatiles and iron and an early magma ocean on both bodies.

The main conventional theories of lunar origin include fission from a rapidly rotating Earth, capture of a fully formed moon into Earth orbit and co-formation with the

Earth from the solar nebula. There are apparently insurmountable problems associated with each of these mechanisms, mainly involving the distribution of angular momentum and geochemical similarities and differences between the two bodies. One oft-stated difficulty with the various theories of lunar origin is their improbability. This, by itself, is not a serious problem since the uniqueness of the Moon in the solar system shows that its formation was a rare event. If formation of moons was a high-probability event, we would have to explain why the Earth has only one moon and why Mercury, Venus and Mars do not have massive moons. On the other hand, no theory of lunar origin can violate the laws of physics or the observables regarding the properties of the Earth and Moon and their orbits.

The most significant clues pertinent to this problem are not the detailed studies resulting from the Apollo program but more mundane observations such as the large number of impacts on the surfaces of the Moon and other bodies and the small number of terrestrial planets. The planets must have grown from something, and at present there is relatively little material to accrete and impacts are very infrequent. We must conclude that there were large numbers of bodies in the early solar system and that the present planets grew by intercepting the orbits of other bodies of all sizes. If the initial orbits of planetesimals had a variety of ellipticities and inclinations, encounters and close encounters would be frequent. Encounters would be both destructive and constructive, but the end result would be to regularize the orbits and to place more objects in a single plane with more nearly circular orbits. Intersection of these orbits would be more likely to result in accretion than in destruction, and eventually the smaller bodies would be perturbed into the orbits of the larger bodies and swept up. The present bimodal distribution of objects—a few large objects and many small objects—is the result of these competing processes. A plausible initial distribution, the result of comminution, would have an increase of about an order of magnitude in the numbers of objects for each order of magnitude decrease in the mass. For example, for every Earth-sized body there would be about 10 Mars-sized bodies and 100 Moon-sized bodies. The impacts of equal-sized bodies would probably destroy both, if they were in dissimilar orbits, because of the high impact velocities. These bodies would be more likely to coalesce if they were in similar orbits. The net result, however, is a few larger bodies and many smaller bodies.

An Earth-sized body could probably survive an impact by a Mars-sized body, but it would be traumatic to both. The Earth would be partially melted and vaporized, and considerable amounts of both the Earth and the impactor would be thrown out of the Earth. If orbital mechanics alone were responsible for the subsequent fate of this material, it would either go into hyperbolic orbit and escape or reimpact the Earth. However, the massive atmosphere caused by such an impact, the mutual interactions of the many small

solid fragments, and the freezing and condensation of liquids and gases ejected from the surface before they escape from the Earth's gravitational attraction results in a small fraction of the material being trapped in Earth orbit, where it accretes into a small version of the Moon. The presence of a protomoon in Earth orbit provides a nucleus for sweeping up debris from later Earth impacts. In this scenario, the material forming the Moon is a mixture of material melted and vaporized from the Earth and the impactor and recondensed in Earth orbit. The Moon would presumably be richer in the early condensing and freezing material and would therefore be refractory rich, as in the condensation scenario, but the material would have undergone a prior stage of differentiation, including siderophile removal, on a large planet.

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