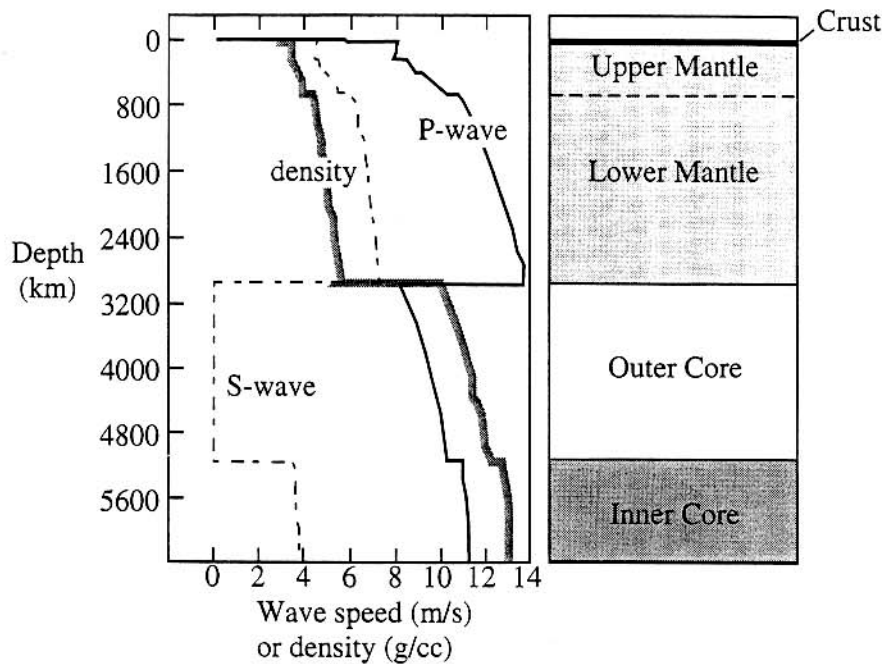


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The composition of the Earth, integrated from core to atmosphere, is comparable to that of undifferentiated meteorites (chondrites). But this simple statement offers little insight into the kind of chondrite the Earth most resembles or if even there is a good analog to the Earth in our present spectrum of chondritic meteorites. It also tells us little of how the Earth got to its present configuration (Fig. 1.1) (i.e., a three-layered, metal–rock–water system). The geophysical, geochemical, and geological characteristics of the Earth reveal much about the planet's overall system. However, when we seek to describe the whole body, such information requires considerable integration and interpretation to see through the last 4.6 Ga of geological history. Improving our understanding of the Earth's composition yields insights into how our planet formed and evolved, as well as providing insights into our planetary neighbors.

Estimating the composition of the Earth requires derivation of the core and silicate Earth (crust plus mantle) composition. A number of papers provide an estimate of the composition of the primitive mantle (or silicate Earth), which are based on samples of the mantle and meteorites, and these show good agreement (Allegre *et al.*, 1995b; Jagoutz *et al.*, 1979; McDonough and Sun, 1995; O'Neill and Palme, 1997). Estimates of the core's composition are less certain, given uncertainties as to the nature of the light element in the outer core. Iron meteorites give insights into elements that might be in the core, but these meteorites are products of low-pressure differentiation, whereas the Earth's core likely formed under markedly different conditions.

Meteorites and observable extrasolar processes tell us much about the nature, composition, and evolution of our solar system. Current models suggest that our solar system formed from the gravitational collapse of a rotating interstellar cloud, which may have been triggered by a nearby supernova (Cameron, 1988; Wetherill, 1990). The evolution from a rotating cloud of gas and dust to a highly structured solar system is modeled as a series of collisional processes having some degree of hierarchical evolution. Dust grains accrete to form small particles, and these combine to form



**Figure 1.1** Schematic diagram of the basic structure of the Earth identifying its three distinct layers: metallic core, rocky silicate shell, and hydrous gaseous exosphere. The density and seismic wave structure (PREM) is shown for comparison (Dziewonski and Anderson, 1981).

planetesimals and protoplanets (asteroidlike bodies). Following this, and perhaps at a slower accretion rate (because of decreased probabilities of collisions, for they are fewer in number), the protoplanets coalesce to form larger bodies. Such large collisions may explain the origin of our moon, the high Fe/silicate ratio of Mercury, and the retrograde rotation of Venus, among other things.

The time scale for these processes is not well known, but insight is provided from meteoritical studies. The oldest materials of our solar system are the Ca–Al inclusions in chondrites. They have radiometric ages (i.e., time of mineral closure for specific isotopic systems) that are within a few million years of  $T_0$  (i.e., where  $T_0$  is considered as the initiation of formation of our solar system and is on the order of 4.60 Ga) (see Russell *et al.*, 1996, and references therein). In addition, extinct radionuclide systems lend additional constraints on early solar system processes (e.g., timing of metal–silicate differentiation on planets and asteroids). Some of the oldest lunar and Martian materials are on the order of 0.1 Ga after  $T_0$ , leading one to conclude that bodies the size of the Earth were formed and differentiating within the first 100 Ma of solar system history.

Material that contributed to the growing Earth came from the same interstellar cloud that gave rise to the other planets and the sun, the latter of

which contains  $> 99\%$  of the solar system's mass. In its initial state, the interstellar cloud was likely to have been compositionally homogeneous to first order, but became chemically heterogeneous during the formation and evolution of the planets (Cameron, 1988). The sun and the outer planets have a substantially greater complement of gases and other volatiles than the inner, rocky planets. Most of the gaseous component in the inner solar system is believed to have been removed during the violent T-tauri stage of the early sun. Outstanding questions include: How did compositional heterogeneity in the solar system developed during planetary accretion? What degree of mixing of components between the "radial feeding zones" in the accreting solar disk is likely to have occurred during planetary coalescence? These questions remain in the realm of speculation until we can get a better handle on the composition of the bulk Earth, Moon, Mars, and other inner planetary bodies.

## 1.1 STRUCTURE OF THE EARTH

The Earth is made up of three major and distinctly different units: the core, the mantle–crust system, and the atmosphere–hydrosphere system (Fig. 1.1). These units are the products of planetary differentiation and are distinctive in composition. The mass of the core is about one-third of the Earth's mass, its volume is about one-eighth of the Earth's, and its radius is about one-half of the Earth's. The silicate part of the Earth (crust and mantle) makes up the remaining two-thirds of its mass, and the rest of its volume, aside from that of the atmosphere/hydrosphere. The Earth, thus, has two distinct boundary layers, the core–mantle boundary and the Earth's surface, with grossly contrasting physical properties above and below these regions. The core is an Fe–Ni alloy, with lesser amount of other siderophile elements and  $\sim 10\%$  by mass of a light element. The crust–mantle system is a mixture of silicates containing primarily magnesium, iron, aluminum, and calcium. The atmosphere–hydrosphere system is dominated by the mass of the oceans, but the atmosphere is unique within the solar system in that it is an 80/20 mixture of  $N_2$  and  $O_2$ .

A broad range of observations provide us with this first-order picture of the Earth. Studies that directly measure physical properties of the Earth's interior include the Earth's seismological profile, its magnetic field, and its orbital behavior, the last of which provides us with a coefficient of the moment of inertia for the Earth. Less direct information on the makeup of the Earth is provided by studies of meteorites and samples of various parts of the silicate Earth. It is from these investigations that we develop models for

the composition of the bulk Earth and primitive mantle (i.e., the silicate Earth) and from these deduce the composition of the core.

The seismological profile of the Earth (Fig. 1.1) images density with depth (Dziewonski and Anderson, 1981). Together this and laboratory studies constrain the mineralogical and chemical constituents of the core and mantle. The time it takes a seismic wave to pass through the Earth is a function of the temperature and elastic properties of its internal layers. The bulk sound velocity of a material is directly correlated to its average density, and this in turn relates to its mean atomic number (Birch, 1952). The average density for the mantle immediately below the Moho, the seismic discontinuity between the crust and mantle, is  $3.3 \text{ Mg/m}^3$ , consistent with an olivine-dominated mineralogical assemblage that also contains pyroxene and an aluminous phase (plagioclase, spinel, or garnet, depending on pressure) (Ringwood, 1975). At deeper levels olivine converts to a  $\beta$ -spinel phase, giving rise to the 410 km depth seismic discontinuity (see review in Agee, 1998). At still deeper levels this olivine component breaks down into Mg-perovskite and magnesio-wüstite, giving rise to the 660 seismic discontinuity. The latter assemblage appears to continue to the core–mantle boundary (see review in Bina, 1998). At this depth there is a dramatic response change in the seismic profile for both P (compressional) and S (shear) waves, recording a fundamental change in the Earth's physical properties. The average density increases from about  $6 \text{ Mg/m}^3$  in the mantle to over  $10 \text{ Mg/m}^3$  in the core, reflecting its Fe-rich composition. A final seismic discontinuity is recorded at 5120 km depth and reveals the transition from a fluid outer core to a solid inner core. S waves do not propagate through the outer core, demonstrating that it is liquid.

It is not surprising that the Earth has so much Fe, given that Fe is the most abundant element, by mass, in the terrestrial planets (because of the stability of the  $^{56}\text{Fe}$  nucleus during nucleosynthesis). The presence of an Fe core in the Earth is also demonstrated by the Earth's shape and magnetic field. The shape of the Earth is a function of its spin and mass distribution. The Earth possesses an equatorial circular bulge and has flattening at the poles due to rotational flattening. The coefficient of the moment of inertia for the Earth is an expression for the distribution of mass within the planet with respect to its rotational axis. If the Earth were a compositionally homogenous planet having no density stratification, its coefficient of the moment of inertia would be  $0.4 Ma^2$ , with  $M$  as the mass of the Earth and  $a$  as the equatorial radius. The equatorial bulge, combined with the precession of the equinoxes, fixes the coefficient of the moment of inertia for the Earth at  $0.330 Ma^2$  (Yoder, 1995), reflecting a marked concentration of mass at its center (i.e., the Fe core). Moreover, the presence of the Earth's magnetic field requires the convection of a significant volume of Fe (or a similarly electrically conducting material) to create a self-exciting dynamo.



## 1.2 CHEMICAL CONSTRAINTS

Combined studies of meteorites and mantle samples place important constraints on compositional models for the silicate Earth. The silicate fraction of the Earth has a composition that is similar to some stony meteorites, or achondrites. These meteorites come from the silicate shells of differentiated planets that have also had a metallic core extracted. Bulk planetary compositional models based on meteorites compare the elemental abundance pattern for the planet with that of various chondrites, which are primitive, undifferentiated meteorites (Wasson and Kallemeyn, 1988). In particular, the CI carbonaceous chondrites, the most primitive of the chondritic meteorites, are often considered as the reference group of chondrites by which to compare planetary compositions (Anders and Ebihara, 1982).

Elements can be classified as refractory, moderately volatile, or volatile, depending on their sequence of condensation into mineral phases (metals, oxides and silicates) from a cooling gas of solar composition (Larimer, 1988). Refractory elements (i.e., Ca, Al, Ti, Sc, REE) have the highest condensation temperatures and occur in all chondrites at similar relative abundance ratios (e.g., Ca/Al, Al/Ti, REE/Ti). The moderately volatile (e.g., Na, K, Rb, Fe, Ni, P) and volatile elements (e.g., F, Cl, Tl, Bi, Pb) have lower condensation temperatures and their relative abundance ratios vary considerably between the different types of chondritic meteorites. The CI carbonaceous chondrites are free of chondrules and Ca-Al rich inclusions, possess the highest abundances of the volatile and moderately volatile elements relative to the refractory elements, and have a composition that matches that of the solar photosphere when compared on a Si-based scale.

Another element classification scheme uses the chemical behavior of elements to group them as lithophile, siderophile, chalcophile, or atmophile (Larimer, 1988). The lithophile elements are concentrated in the silicate shell of the Earth and are elements that bond readily with oxygen (i.e., the silicates). The siderophile elements are those that readily bond with Fe and are most concentrated in the Earth's core. The chalcophile elements bond readily with S and are distributed between the core and mantle, with a greater percentage of them in the core. Finally, the atmophile elements are gaseous and concentrated in the thin layer of atmosphere that surrounds the planet. By combining these two different classification schemes we can better understand the relative behavior of elements, particularly during accretion and large-scale planetary differentiation.

Establishing the composition of Earth and its major compositional reservoirs can be done in a four-step process. First, establish an estimate for the composition of the silicate Earth. Second, define a volatility curve for the planet, based on the abundances of the moderately volatile and volatile

lithophile elements in the silicate Earth, assuming that none of these elements have been sequestered into the core (i.e., they are truly lithophile). Third, estimate the abundances of the siderophile and chalcophile elements in the core based on the volatility curve established in step 2, meteorite analogs, and the small amount of these elements in silicate Earth. Fourth and finally, sum the core and mantle composition with that of the atmosphere to obtain a bulk Earth composition. The resulting model can be compared to geophysical data to test for consistency.

### 1.2.1 The Composition of the Silicate Earth

Most models for the composition of the silicate Earth agree to within  $\sim 10\%$  at the major element level (Allegre *et al.*, 1995b; Anderson, 1989b; Jagoutz *et al.*, 1979; McDonough and Sun, 1995). These models assume that the refractory elements are in chondritic proportion in the bulk Earth and that those that are lithophile are excluded from the core. The silicate Earth's composition is derived by comparing the compositions of peridotites (samples of the upper mantle) with that of primitive, mantle-derived magmas, the chemical complement to peridotites (Sun, 1982). Model compositions derived in this fashion are then compared with data from chondrites.

Compositional models for the silicate Earth usually fall in one of two categories based on major elements. One class of models assumes that the silicate Earth has a complement of Mg and Si that is equal to that in CI carbonaceous chondrites. Given this, the remaining elements are grouped into either more refractory (i.e., with higher condensation temperatures during solar system formation and equal to that in CI chondrites), or volatile groups. For those that are volatile, the rock record establishes their abundances in the silicate Earth. The second model depends on mantle samples to establish the Mg and Si abundance in the silicate Earth and shows that it does not have CI chondritic relative abundances of these elements when compared to the refractory lithophile elements. Of the five major element oxides in the silicate Earth (i.e.,  $\text{SiO}_2$ ,  $\text{MgO}$ ,  $\text{FeO}$ ,  $\text{Al}_2\text{O}_3$ , and  $\text{CaO}$ ), only  $\text{Al}_2\text{O}_3$ , and  $\text{CaO}$  are truly refractory and are constrained by observations made on chondrites; the others are volatile, and their abundances are not fixed between different varieties of chondrites. Fe is siderophile and is partitioned between the core and mantle. In contrast, Mg and Si are lithophile and concentrated in the mantle, although under very reduced conditions Si will form metals and would be partitioned into the core.

In models thus derived, relevant questions include (1) how representative are peridotites of the entire silicate Earth, and (2) do upper mantle samples faithfully record the composition of the lower mantle? Although there are other lithologies in our samples of the upper mantle, in principle, the upper mantle can be considered to be a peridotite, with volumetrically minor amounts of other components. In addition, because the continental crust is