Comparative planetology

The sun, with all those planets revolving around it and dependent on it, can still ripen a bunch of grapes as if it had nothing else in the universe to do.

Galileo Galilei

Before the advent of space exploration, Earth scientists had a handicap almost unique in science: they had only one object to study. Compare this with the number of objects available to astronomers, particle physicists, biologists and sociologists. Earth theories had to be based almost entirely on evidence from Earth itself. Although each object in the solar system is unique, we have learned some lessons that can be applied to Earth.

(1) Study of the Moon, Mars and the basaltic achondrites demonstrated that early melting is ubiquitous.

(2) Although primitive objects, such as the carbonaceous chondrites, have survived for the age of the solar system, there is no evidence for the survival of primitive material once it has been in a planet.

(3) The magma-ocean concept proved useful when applied to the Earth, taking into account the differences required by the higher pressures on the Earth.

(4) The importance of great impacts in the early history of the planets is now clear.

(5) Material was still being added to the Earth and Moon after the major accretion stage and the giant impacts, and is still being added, including material much richer in the noble metals and noble gases than occur in the crust or mantle.

(6) The difference in composition of the atmospheres of the terrestrial planets shows that the original volatile compositions, the extent of outgassing – or the subsequent processes of atmospheric escape – have been quite different.

(7) We now know that plate tectonics, at least the recycling kind, is unique to Earth. The thickness and average temperature of the lithosphere and the role of phase changes in basalt seem to be important. Any theory of plate tectonics must explain why the other terrestrial planets do not behave like Earth.

Although the Earth is a unique body, and is the largest of the terrestrial planets, we can apply lessons learned from the other objects in the solar system to the composition and evolution of the Earth. The Earth is also an average terrestrial planet; if we take one part Mercury, one part each of Venus and Mars, and throw in the Moon, we have a pretty good Earth, right size and density, and about the right size core. The inner solar system has the equivalent of two Earths.

Planetary crusts

The total crustal volume on the Earth is anomalously small, compared with other planets, and compared with its crust-forming potential, but
it nevertheless contains a large fraction of the terrestrial inventory of incompatible elements. The thin crust on Earth can be explained by crustal recycling and the shallowness of the basalt-eclogite boundary in the Earth. Most of Earth's 'crust' probably resides in the transition region of the mantle. Estimates of bulk Earth chemistry can yield a basaltic layer of about 10% of the mass of the mantle.

The crust of the Earth is enriched in Ca, Al, K and Na in comparison to the mantle, and ionic-radii considerations and experimental petrology suggest that the crust of any planet will be enriched in these constituents. A maximum average crustal thickness for a fully differentiated chondritic planet can be obtained by removing all of the CaO, with the available Al₂O₃, as anorthite to the surface. This operation gives a crustal thickness of about 100 km for Mars. Incomplete differentiation and retention of CaO and Al₂O₃ in the mantle will reduce this value, which is likely to be the absolute upper bound (Earth's crust is much thinner due to crustal recycling, delamination and the basalt-eclogite phase change). In the case of the Earth, up to 60-70% of some large-ion elements are in the crust, implying that about 30-40% of the crustal elements are in the mantle. This does not require that 30-40% of the mantle is still in a primordial undegassed state as some geochemists believe.

The average thickness of the crust of the Earth is only 15 km, which amounts to 0.4% of the mass of the Earth. The crustal thickness is 5-10 km under oceans and 30-50 km under older continental shields. The thickest crust on Earth – about 80 km – is under young actively converging mountain belts. The parts deeper than about 50 km may eventually convert to eclogite, and fall off. The situation on the Earth is complicated, since new crust is constantly being created at midoceanic ridges and consumed at island arcs. The continental crust loses mass by erosion and by delamination of the lower eclogitic portions. Continental crust is recycled but its total volume is roughly constant with time. Both the Moon and Mars have crustal thicknesses greater than that of the Earth in spite of their much smaller sizes, and probable less efficient differentiation.

**Mercury – first rock from the Sun**

Mercury is 5.5% of the mass of the Earth, but it has a very similar density, 5.43 g/cm³. Its radius is 2444 km. Any plausible bulk composition is about 60% iron and this iron must be largely differentiated into a core. Mercury has a perceptible magnetic field, appreciably more than either Venus or Mars, probably implying that the core is molten. Mercury's surface is predominantly silicate, but apparently not basaltic. A further inference is that the iron core existed early in its history; a late core-formation event would have resulted in a significant expansion of Mercury.

Mercury's shape may have significantly changed over the history of the planet. Tidal de-spinning results in a less oblate planet and compressional tectonics in the equatorial regions. Cooling and formation of a core cause a change in the mean density and radius. A widespread system of arcuate scarps on Mercury, which appear to be thrust faults, provides evidence for compressional stresses in the crust. The absence of normal faults suggests that Mercury has contracted. This is evidence for cooling of the interior.

One factor affecting the bulk composition of Mercury is the probable high temperature in its zone of the solar nebula; it may have formed from predominantly high-temperature condensates. If the temperature was held around 1300 K until most of the uncondensed material was blown away, then a composition satisfying Mercury's mean density can be obtained, since most of the iron will be condensed, but only a minor part of the magnesian silicates. Since the band of temperatures at which this condition prevails is quite narrow, other factors must be considered. Two of these are (1) dynamical interaction among the material in the terrestrial planet zones, leading to compositional mixing, and (2) collisional differentiation. A large impact after core formation may have blasted away much of the silicate crust and mantle. Our Moon may have been the result of such an impact on proto-Earth. On Mars, the crust is locally thinner under the large impact basins.
Terrestrial bodies were subjected to a high flux of impacting objects in early planetary history. The high-flux period can be dated from lunar studies at about 3.8 billion years ago. The large basins on the surface of Mercury formed during this period of high bombardment. Later cooling and contraction apparently were responsible for global compression of the outer surface and may have shut off volcanism. On the Earth, volcanism is apparently restricted to the extending regions.

**Venus**

Venus is 320 km smaller in radius than the Earth and is about 4.9% less dense. Most of the difference in density is due to the lower pressure, giving a smaller amount of self-compression and deeper phase changes. Venus is a smoother planet than the Earth but has a measurable triaxiality of figure and a 0.34 km offset of the center of the figure from the center of mass. This offset is much smaller than those of the Moon (2 km), Mars (2.5 km) and Earth (2.1 km).

In contrast to the bimodal distribution of Earth’s topography, representing continent-ocean differences, Venus has a narrow unimodal height distribution with 60% of the surface lying within 500 m of the mean elevation. This difference is probably related to erosion and isostatic differences caused by the presence of an ocean on Earth. For both Earth and Venus the topography is dominated by long-wavelength features. Most of the surface of Venus is gently rolling terrain. The gravity and topography are positively correlated at all wavelengths. On Earth most of the long-wavelength geoid is uncorrelated with surface topography and is due to deep-mantle dynamics or density variations.

The other respects in which Venus differs markedly from the Earth are its slow rotation rate, the absence of a satellite, the virtual absence of a magnetic field, the low abundance of water, the abundance of primordial argon, the high surface temperature and the lack of obvious signs of subduction. From crater counts it appears that the age of the surface of Venus is 300-500 million years old, much less than parts of the Earth’s surface. The oceanic crust on the Earth is renewed every 200 million years but the continents survive much longer.

If Venus had an identical bulk composition and structure to the Earth, then its mean density would be about 5.34 g/cm³. By ‘identical structure’ I mean that (1) most of the iron is in the core, (2) the crust is about 0.4% of the total mass and (3) the deep temperature gradient is adiabatic (an assumption). The high surface temperature of Venus, about 740 K, would have several effects; it would reduce the depth at which the convectively controlled gradient is attained, it would deepen temperature-sensitive phase changes and it may prevent mantle cooling by subduction.

The density of Venus is 1.2-1.9% less than that of the Earth after correcting for the difference in pressure. This may be due to differences in iron content, sulfur content, oxidization state and deepening of the basalt-eclogite phase change. Most of the original basaltic crust of the Earth subducted or delaminated when the upper-mantle temperatures cooled into the eclogite stability field. The density difference between basalt and eclogite is about 15%. Because of the high surface temperature on Venus, the upper-mantle temperatures are likely to be 200–400 K hotter in the outer 300 km or so than at equivalent depths on Earth, or melting is more extensive. This has interesting implications for the phase relations in the upper mantle and the evolution of the planet. In particular, partial melting in the upper mantle would be much more extensive than is the case for the Earth except for the fact that Venus is probably deficient in the volatile and low-molecular-weight elements that also serve to decrease the melting point and viscosity. Crust can be much thicker because of the deepening of the basalt-eclogite phase boundary.

Schematic geotherms are shown in Figure 2.1 for surface temperatures appropriate for Earth and Venus. With the phase diagram shown, the high-temperature geotherm crosses the solidus at about 85 km. With other plausible phase relations the eclogite field is entered at a depth of about 138 km. For Venus, the lower gravity and outer-layer densities increase these depths by about 20%; thus, we expect a surface layer
of 100–170 km thickness on Venus composed of basalt and partial melt. On the present Earth, the eclogite stability field is entered at a depth of 40–60 km. If the interior of Venus is dry it will be stronger at a given temperature and will have a higher solidus temperature.

A large amount of basalt has been produced by the Earth's mantle, but only a thin veneer is at the surface at any given time. There must therefore be a substantial amount of eclogite in the mantle, the equivalent of about 200 km in thickness. If this were still at the surface as basalt, the Earth would be several percent less dense. Correcting for the difference in temperature, surface gravity and mass and assuming that Venus is as well differentiated as Earth, only a fraction of the basalt in Venus would have converted to eclogite. This would make the uncompressed density of Venus about 1.5% less than Earth's without invoking any differences in composition or oxidation state. Thus, Venus may be close to Earth in composition. It is possible that the present tectonic style on Venus is similar to that of Earth in the Archean, when temperatures and temperature gradients were higher. If the Moon, Mercury and Mars have molten iron core, it is probable that Venus does as well.

The youthful age of the surface of Venus has been attributed to a global resurfacing event. The cratering record indicates that the global resurfacing event, about 300 my ago, was followed by a reduction of volcanism and tectonism. Delamination of thick basaltic crust, foundering of a cold thermal boundary layer and a massive reorganization of mantle convection are candidates for the resurfacing event. Mantle convection itself is strongly controlled by surface processes and changes in these processes.

**Mars**

Mars is about one-tenth of the mass of Earth. The uncompressed density is substantially lower than that of Earth or Venus and is very similar to the inferred density of a fully oxidized (minus C and \( \text{H}_2\text{O} \)) chondritic meteorite. The moment of inertia, however, requires an increase in density with depth over and above that due to self-compression and phase changes, indicating the presence of a dense core. This in turn indicates that Mars is a differentiated planet.

The tenuous atmosphere of Mars suggests that it either is more depleted in volatiles or has experienced less outgassing than Earth or Venus. It could also have lost much of its early atmosphere by large impacts. Geological evidence for running water on the surface of Mars suggests that a large amount of water is tied up in permafrost and ground water and in the polar caps. Whether Mars had standing water – oceans and lakes – for long periods of time is currently being debated. The high \(^{40}\text{Ar}/^{36}\text{Ar}\) ratio on Mars,
ten times the terrestrial value, suggests either a high potassium-40 content plus efficient outgassing, or a net depletion of argon-36 and, possibly, other volatiles. If Mars is volatile-rich, compared to Earth, it should have more K and hence more argon-40. Early outgassed argon-36 could also have been removed from the planet.

SNC meteorites have trapped rare-gas and nitrogen contents that differ from other meteorites but closely match those in the martian atmosphere. If SNC meteorites come from Mars, then a relatively volatile-rich planet is implied, and the atmospheric evidence for a low volatile content for Mars would have to be rationalized by the loss of the early accretional atmosphere. Mars is more susceptible to atmospheric escape than Venus or Earth owing to its low gravity. The surface of Mars appears to be weathered basalt. The dark materials at the surface contain basaltic minerals and hematite and sulfur-rich material, and there is evidence for the past action of liquid water. The large volcanoes on Mars are similar in form to shield volcanoes on Earth. Andesite – a possible indicator of plate tectonics – has been proposed as a component of martian soil but this is controversial; weathered basalt can explain the available data.

The topography and gravity field of Mars indicate that parts of Mars are grossly out of hydrostatic equilibrium and that the crust is highly variable in thickness. If variations in the gravity field are attributed to variations in crustal thickness, reasonable values of the density contrast imply that the average crustal thickness is at least 45 km, and the maximum crustal thickness may reach 100 km. Giant impacts may have removed most of the crust beneath the basins, replacing crustal material by uplifted mantle. If so, the crust was in place in early martian history, consistent with other evidence throughout the solar system for rapid early planetary differentiation. On Earth, delamination of lower crust produces a thinning but the whole crust is not involved.

The only direct evidence concerning the internal structure of Mars is the mean density, moment of inertia, topography and gravity field. The mean density of Mars, corrected for pressure, is less than that of Earth, Venus and Mercury but greater than that of the Moon. This implies either that Mars has a small total Fe-Ni content or that the FeO/Fe ratio varies among the planets. Plausible models for Mars can be constructed that have solar or chondritic values for iron, if most or all of it is taken to be oxidized. With such broad chemical constraints, mean density and moment of inertia and under the assumption of a differentiated planet, it is possible to trade off the size and density of the core and density of the mantle.

The mantle of Mars is presumably composed mainly of silicates, which can be expected to undergo one or two major phase changes, each involving a 10% increase in density. To a good approximation, these phase changes will occur at one-third and two-thirds of the radius of Mars. The deeper phase change will not occur if the radius of the core exceeds one-third of the radius of the planet.

The curve in Figure 2.2 is the locus of possible Mars models. Clearly, the data can accommodate a small dense core or a large light core. The upper limit to the density of the core is probably close to the density of iron, in which case the core
would be 0.36 of Mars' radius, or about 8% of its mass. To determine a lower limit to the density, one must consider possible major components of the core. Of the potential core-forming materials, iron, sulfur, oxygen and nickel are by far the most abundant elements. The assumption of a chondritic composition for Mars leads to values of the relative radius and mass of the core: \( \frac{R_c}{R} = 0.50 \) and \( \frac{M_c}{M} = 0.21 \). The inferred density of the mantle is less than the density of the silicate phase of most ordinary chondrites.

Three kinds of chondrites, HL (high iron, low metal), LL (low iron, low metal) and L (low iron or hypersthene-olivine) chondrites, all have lower amounts of potentially core-forming material than is implied for Mars, although HL and LL have about the right silicate density (3.38 g/cm³). If completely differentiated, H (high iron) chondrites have too much core and too low a silicate density (3.26-3.29 g/cm³). We can match the properties of H chondrites with Mars if we assume that the planet is incompletely differentiated. If the composition of the core-forming material is on the Fe side of the Fe–FeS eutectic, and temperatures in the mantle are above the eutectic composition, but below the liquidus, then the core will be more sulfur-rich and therefore less dense than the potential core-forming material.

Carbonaceous chondrites are extremely rich in low-temperature condensates, as well as carbon and 'organic matter.' If we ignore the water and carbon components, a fully differentiated planet of this composition would have a core of 15% by mass, composed mainly of FeS (13.6% FeS, 1.4% Ni), and a mantle with a density of about 3.5 g/cm³. However, these meteorites also contain about 19% water, most of which must have escaped if Mars is to be made up primarily of this material. Otherwise, the mantle would not be dense enough. But there is abundant evidence for water near the surface of Mars in the past.

In ordinary high-iron chondrites, the free iron content averages 17.2% by weight. The FeS content is approximately 5.4% (3.4% Fe, 2.0% S) and the nickel content is 1.6%. A planet assembled from such material, if completely differentiated, would yield a core of 24% of the mass of the planet, with Fe: S: Ni in the approximate proportions of 21:2:2 by weight. Low-iron chondrites would yield a core of 15% of the mass of the planet, with proportions of 12:1:2. Enstatite chondrites contain between 15–25 weight% free iron. The oxidation state and the oxygen isotopes of enstatite chondrites make them an attractive major component for forming the Earth. These are not trivial considerations since oxygen is the major element, by volume, in a terrestrial planet.

Carbonaceous chondrites have little or no free iron but contain 7–25% by weight FeS and about 1.5% nickel. The average core size for a planet made of carbonaceous chondrites would be 15% by mass, Fe: S: Ni being in the proportions 18:10:3. An absolute minimum core density can probably be taken as 4.8 g/cm³, corresponding to a pure FeS core with a fractional core radius of 0.6 and a fractional mass of 26%. On these grounds, the mass of the martian core can be considered to lie between 8 and 24% of the mass of the planet.

A third possibility would be to assemble Mars from a mixture of meteorites that fall above and below the curve. The meteorites below the curve are relatively rare, although Earth may not be collecting a representative sample. The size of the core and its density can be traded off. By using the density of pure iron and the density of pure troilite (FeS) as reasonable upper and lower bounds for the density of the core, its radius can be considered to lie between 0.36 and 0.60 of the radius of the planet. It is probably that there is sulfur in the core. Therefore, a core about half the radius of the planet, with a low melting point, is a distinct possibility.

The zero-pressure density of the mantle implies an FeO content of 21–24 wt.% unless some free iron has been retained by the mantle. The presence of CO₂ and H₂O, rather than CO and H₂, in the martian atmosphere suggests that free iron is not present in the mantle. Chondrites may therefore be an appropriate guide to the major-element composition of Mars. The core of Mars is smaller and less dense than the core of Earth, and the mantle of Mars is denser than that of Earth. Mars contains 25–28% iron, independent of assumptions about the overall composition or distribution of the iron. Earth is clearly enriched in iron or less oxidized than Mars and most classes of chondritic meteorites.
The lithosphere on Mars is capable of supporting large surface loads. The evidence includes the roughness of the gravity field, the heights of the shield volcanoes, the lack of appreciable seismicity and thermal history modeling. There is some evidence that the lithosphere has thickened with time. *Olympus Mons* is a volcanic construct with a diameter of 700 km and at least 20 km of relief, making it the largest known volcano in the solar system. It is nearly completely encircled by a prominent scarp several kilometers in height and it coincides with the largest gravity anomaly on Mars. The origin of the *Olympus Mons* scarp is controversial; it may be, in part, due to spreading of the volcano, and, in part, to erosion.

The surface of Mars is more complex than that of the Moon and Mercury. There is abundant evidence for volcanic modification of large areas after the period of heavy bombardment, subsequent to 3.8 Ga. Mars has a number of gigantic shield volcanoes and major fault structures. In contrast to Mercury there are no large thrust or reverse faults indicative of global contraction; all of the large tectonic features are extensional. The absence of terrestrial-style plate tectonics is probably the result of a thick cold lithosphere. In any event, the absence of plate tectonics on other planets provides clues to the dynamics of a planet that are unavailable from the Earth. Apparently, a planet can be too small or too dry or too old or too young, for it to have plate tectonics. Liquid water, magnetic fields, plate tectonics and life are all unique to Earth and there may be a reason for this.

**Moon**

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

*John Byrom*

The Moon is deficient in iron compared with the Earth and the other terrestrial planets. It is also apparently 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% iron, which is consistent with the composition of stony meteorites and the non-volatile components of the Sun. They therefore fit into any scheme that has them evolve from solar material. Mercury is overendowed with iron, which has led to the suggestion that part of the mantle was blasted away by impacts. Because iron is the major dense element occurring in the Sun, and in the preplanetary solar nebula, the Moon is clearly depleted in iron, and in a number of other elements as well. A common characteristic of many of these elements and their compounds is volatility. Calcium, aluminum and titanium are the major elements involved in high-temperature condensation processes; 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 surface samples of the Moon are 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 cooling-gas model of planetary formation, the refractories condense before the volatiles, and it was therefore implied that the Moon was made inside out! The standard geochemical model of terrestrial evolution also invokes a volatile-rich interior, one that is rich in $^3$He, but there is no evidence for this. The strange chemistry of the Moon is consistent with condensation from a gas-dust cloud caused by a giant impact on proto-Earth.

Large-ion refractory 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. Similarly, in terrestrial samples, there are missing elements that imply eclogitic and kimberlitic material at depth.

The maria on the Moon 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.

Seismic activity of the Moon is much lower than on Earth, 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 occur at great depth, about half the lunar radius. 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 lunar crust

The thickness and composition of the lunar highland crust 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% 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 aggregates 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% of the europium and potassium contents of the Moon now reside in the highland crust, which is less than 9% of the mass of the Moon. Estimates of the 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. 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.

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. 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. However, a magma ocean is also quite likely for the Earth and probably the other terrestrial planets as well.

Tables 2.1 and 2.2 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 volatiles. 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.

The origin of the Moon
Prior to the Apollo landings in 1969 there were three different theories of lunar origin. The fission theory, proposed by G. H. Darwin, Charles Darwin's son, supposed that the moon was spun out of Earth's mantle during an early era of rapid Earth rotation. The capture theory supposed that the moon formed somewhere else in the solar system and was later captured in orbit about the Earth. The co-accretion or 'double planet' theory supposed that the Earth and moon grew together out of a primordial swarm of small 'planetesimals.' All three theories made predictions at
Table 2.2 Composition of the continental crust

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</tbody>
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A: Andesite model (Taylor and McLennan, 1985).
B: Amphibolite-granulite lower-crustal model (Weaver and Tarny, 1984).
C: Theoretical model (Taylor and McLennan, 1985).

variance with the observations that the moon has no substantial metallic iron core, and that its rocks are similar in composition to the Earth’s mantle (its oxygen isotopic ratios are identical to the Earth’s), but are strongly depleted in volatiles. It was then realized that accreting matter would form embryonic planets with a large range of sizes.

The final stages of planetary formation would involve giant impacts in which bodies of comparable size collided at high speed. A giant impact produces rock-vapor that preferentially retains the refractory elements as it condensed. Shortly after the formation of the Earth, a large object is inferred to have hit the Earth at an oblique angle, destroying the impactor and ejecting most of that body along with a significant amount of the Earth’s silicate portions. Some of this material then coalesced into the Moon. This event also melted a large fraction of the Earth. From the angular momentum of the present Earth-Moon system the projectile is inferred to have had a mass comparable to Mars and the Earth was smaller than it is today. After the collision the Earth is a very hot body indeed. The idea of a cold primordial undegassed Earth can no longer be entertained. The present lower mantle is more likely to be refractory and gas-poor than to be primordial and gas-rich. The giant impact theory is the now dominant theory for the formation of the Moon. The theory was proposed in 1975 by Hartman and Davis (see Hartman et al., 1986).