

Orionis, some four orders of magnitude greater than in quiescent T-Tauri stars. These outbursts are thought to be the result of greatly enhanced mass accretion rates, perhaps as high as $10^{-4} M_{\odot}$ per year. The cause is unclear: thermal instability in the disk, changes in the structure of the disk, and the gravitational effects of a binary companion or a giant planet or protoplanet orbiting close to the star have all been suggested as causes of the enhanced mass accretion rate. The outbursts appear to last no more than a century. By the end of Phase II, the star is perhaps 2 to 6 million years old.

Phase III is represented by *weak-lined T-Tauri stars*, so called because spectral emission and absorption lines are much weaker than in classical T-Tauri stars. In addition, the excess infrared emission that characterizes classical T-Tauri stars is absent in weak-lined T-Tauri stars. The inference from these observations is that the disk has largely dissipated by this stage. Like classical T-Tauri stars, weak-lined T-Tauri stars are cooler yet more luminous than mature main sequence stars of similar mass, but they are closer to the main sequence on the Hertzsprung-Russell diagram than classical T-Tauris.

Weak-lined T-Tauris are particularly luminous in the X-ray part of the spectrum. These X-rays are thought to be produced in hot plasma during magnetic reconnection events above the stellar surface. The same process in the Sun produces bright solar flares, but flares of weak-lined T-Tauris are 100–1000 times more powerful. Outflows and winds subside to those of typical main sequence stars as accretion ends and the star reaches its final mass. During the weak-lined T-Tauri phase, the star contracts to its final radius and density. At the end of this process, fusion ignites in the core and the luminosity and temperature of the star settles onto the main sequence. The entire process from Phase 0 through Phase III consumes perhaps 10 million years.

10.5.2 The condensation sequence

In the preceding section, we saw young stellar objects and their surrounding disks are highly energetic and at least locally very hot environments. We learned earlier that chondrites contain a variety of components, chondrules, CAIs, and AOAs, that provide evidence of at least local and transient high temperatures as

our solar system was born. It is thus useful to ask what the chemical effects of temperature are. For example, suppose we heated material of solar or chondritic composition to the point where everything evaporated. What would be the sequence of condensation? The theoretical condensation sequence has been calculated from thermodynamic data, largely by Larimer, Grossman, and Lewis, all of whom have worked with E. Anders at the University of Chicago (e.g., Larimer, 1967; Grossman, 1972). The condensation temperature of an element reflects its vapor pressure, its tendency to react with other elements to form compounds in the gas or solid solutions or alloys in the solid, and its abundance in the gas. Let's consider two examples of condensation sequence calculations. First, consider iron, which is a particularly simple case since it condenses as Fe metal:



Assuming ideality, the partial pressure of iron is simply its mole fraction in the gas times the total pressure (P_T). Since hydrogen is by far the dominant element in the gas (~98%), the mole fraction can be approximated as 2 times the solar Fe/H ratio (the 2 arises from hydrogen's presence as H_2). Thus the partial pressure of Fe is written as:

$$p_{\text{Fe}} = \frac{[\text{Fe}]_{\odot}}{\frac{1}{2}[\text{H}]_{\odot}} P_T \quad (10.3)$$

where $[\text{Fe}]_{\odot}$ and $[\text{H}]_{\odot}$ are the solar abundances of Fe and H and P_T is total pressure. Once condensation begins, we can express the equilibrium constant for this reaction as the ratio of the partial pressure of Fe in the gas to the concentration in the solid:

$$K = \frac{p_{\text{Fe}}}{[\text{Fe}]_s} \quad (10.4)$$

where $[\text{Fe}]_s$ is the concentration in the solid and p_{Fe} is the partial pressure in the gas.

Condensation begins when the partial pressure of Fe exceeds the equilibrium vapor pressure of solid Fe. Since:

$$\Delta G = -RT \ln K \quad (3.86)$$

the equilibrium constant can also be written as:

$$-\ln K = -\frac{\Delta H_V}{RT} + \frac{\Delta S_V}{R} \quad (10.5)$$

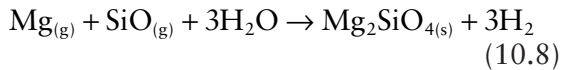
ΔH_V and ΔS_V are the enthalpy and entropy of vaporization. Once condensation of an element begins, its partial pressure drops by $(1 - \alpha)$ where α is the fraction condensed. Hence, eqn. 10.3 becomes:

$$p_{\text{Fe}} = \frac{(1 - \alpha)[\text{Fe}]_{\odot}}{\frac{1}{2}[\text{H}]_{\odot}} P_T \quad (10.6)$$

Combining eqns. 10.4, 10.5, and 10.6, the equation describing condensation is:

$$\ln(\alpha - 1) = -\frac{\Delta H_V}{RT} + \frac{\Delta S}{R} - \ln \frac{[\text{Fe}]_{\odot}}{\frac{1}{2}[\text{H}]_{\odot}} - \ln P_T + \ln [\text{Fe}]_s \quad (10.7)$$

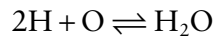
Now consider elements such as Mg and Si, which form compounds in both the gas and solid, and complicate the calculation considerably. The reaction for the condensation of forsterite is:



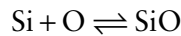
The equilibrium constant for this reaction is:

$$K = \frac{a_{\text{Mg}_2\text{SiO}_4} p_{\text{H}_2}^3}{p_{\text{Mg}}^2 p_{\text{SiO}} p_{\text{H}_2\text{O}}^3 P_T^3} \quad (10.9)$$

The first step is to compute partial pressures of the gaseous species. For example, for SiO and H₂O above, we may write reactions:



and



and calculate equilibrium constants for them from the free energies of these species, for example:

$$K = \frac{p_{\text{SiO}}}{p_{\text{Si}} p_{\text{O}}} \quad (10.10)$$

Values of K can be computed from thermodynamic data. For each element, an additional constraint is imposed by the total abundance of that element in the gas. Thus for example:

$$[\text{SiS}] + [\text{Si}] + [\text{SiO}] + [\text{SiO}_2] = [\text{Si}]_{\odot} - [\text{Si}]_s \quad (10.11)$$

where $[\text{Si}]_s$ is the total of silicon in condensed phases. Combining equilibrium constant equations such as 10.10 with mass balance equations such as 10.11 leads to a series of simultaneous equations. These can be solved by successive approximation using a computer. These kinds of calculations were first done by Larimer (1967) and have since been refined by many workers (e.g., Grossman, 1972, 2010; Petaev and Wood, 1998; Ebel and Grossman, 2000). Values of equilibrium constants such as eqn. 10.9 can then be computed from ΔH and ΔS using eqn. 10.5.

Further complications arise when solid solutions form. For example, the forsterite term in eqn. 10.8 is 1 if forsterite is treated as a pure phase. If the solid solution with fayalite is considered, then the value of the activity must also be calculated. Values for activity coefficients are difficult to obtain, and ideal solid solution is generally assumed. Thus the activity of forsterite in eqn. 10.9 would be replaced by its mole fraction. Solid solution results in the condensation of an element at a higher temperature than if a pure component were the condensed phase.

Some elements, such as Au and Cu, will condense primarily either by reaction with already-condensed Fe metal grains, or by condensing with Fe metal. At lower temperatures (around 670 K for $P_T = 10$ Pa), the Fe metal will react with H₂S gas to form FeS. Moreover, from the onset of condensation a small but increasing amount of the Fe will react with H₂O gas to form FeO that dissolves in the silicates. A marked increase in the Fe content of silicates occurs around 400–500 K.

Figure 10.34 shows two theoretical sequences calculated by Larimer (1967). In the “fast cooling” sequence, matter does not react with nebular gas after it has condensed – this is effectively fractional condensation, analogous to fractional crystallization. At any given time and temperature, the solid phases are a mixture of material condensed over a

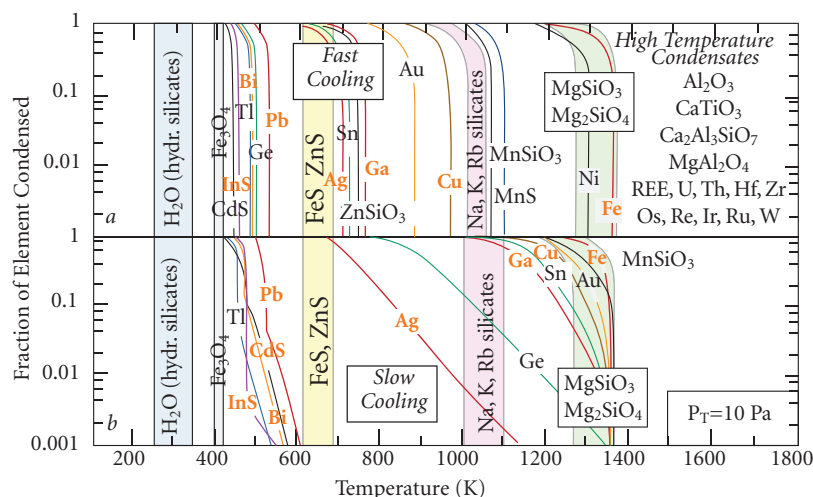


Figure 10.34 Condensation sequence of a gas with solar composition. In slow cooling (b), condensed solids are assumed to continually re-equilibrate with the gas. Fast cooling (a) leads to disequilibrium and condensation of pure elements and compounds. Black lines show condensation of silicates and oxides; red lines show condensation of metals and sulfides. Larimer (1967). With permission from Elsevier.

range of temperature. In the “slow cooling” or “equilibrium” sequence, condensed material continually reacts and re-equilibrates with the gas as temperature drops, so that at any time, only an assemblage in equilibrium with the gas at that temperature is present. Figure 10.35 illustrates the minerals expected at a given temperature somewhat more clearly. The condensation sequence depends critically on total pressure and H pressure; the sequence shown is for relatively low total and H pressure. At relatively high pressure, metallic Fe liquid is the first phase to condense.

In a nutshell, the sequence goes like this (Figure 10.35). The first elements to condense would be Re and the most refractory of the platinumoid metals (Os, Ir, Ru), which would condense as metallic phases. Since these are extremely rare elements, they would likely form very small grains. As we noted earlier, small nuggets of such metal, called “Fremdlinge” (the German word for “strangers”), are found as inclusions in CAIs. Following this would be condensation of oxides and silicates of Ca, Al, and Ti. They should be rich in refractory trace elements such as U, Th, Zr, Ba and the REE. This closely matches the composition of the CAIs, suggesting the possibility that CAIs are high-temperature condensates.

Next in the condensation sequence should come metallic Fe-Ni and compounds richer in the moderately refractory elements such as Mg and Si: olivines and pyroxenes. If the cooling takes place under equilibrium conditions, the high-temperature (CAI) assemblage should react to form anorthite as well, and at lower temperature when Na condenses, plagioclase. These phases are the ones that predominate in chondrules, with the important caveat that chondrules are poorer in metal than the condensation sequence would predict. Since these phases condense at temperatures similar to Fe-Ni metal, some process must have separated metal from silicates before formation of the chondrules.

The Fe should also largely react out to form more Fe-rich olivine and pyroxene. At lower temperature, S condenses and reacts with Fe to form sulfides. At even lower temperature, the Fe reacts with O to form magnetite and the silicates react with water vapor to form hydrated silicates. Sulfates, carbonates and organic compounds will also form around these temperatures.

If equilibrium conditions prevail, only the last-named compounds would exist when condensation was complete, but all might exist if disequilibrium prevails. The CI chondrites seem to be very similar in composition

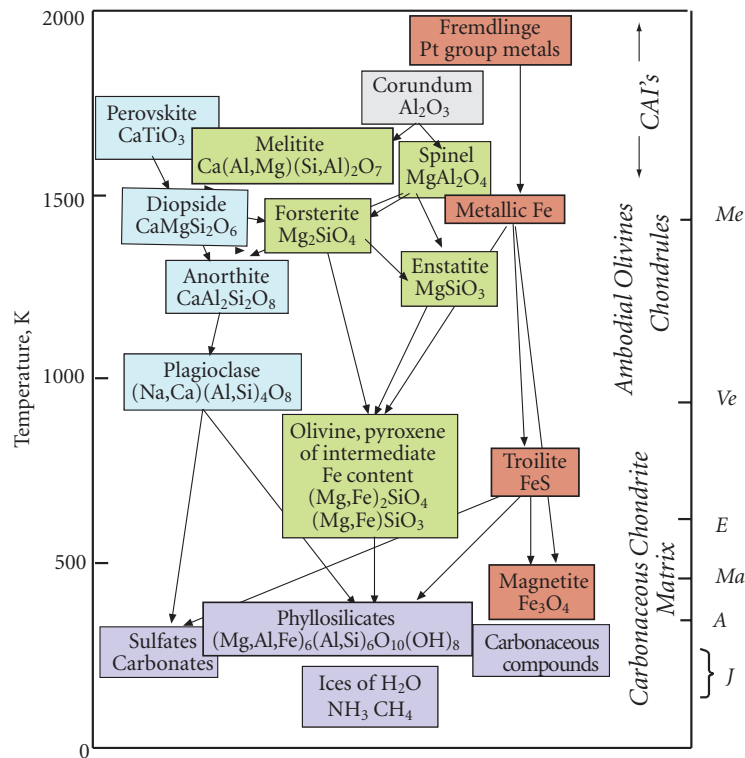


Figure 10.35 Simplified mineralogical condensation sequence. McSween (1987). Reprinted with permission of Cambridge University Press.

and petrography to equilibrium condensates (down to 300 K or so), or, more accurately, accretions of equilibrium condensates. The other carbonaceous chondrites approximate, to varying degrees, aggregates of disequilibrium condensates. In particular, the CV and CO chondrites contain both the highest temperature condensates (CAIs) and lowest temperature material (hydrated silicates) and much of everything expected to condense in between. CV and CO chondrites are depleted in the more volatile elements and compounds (such as water, but also the moderately volatiles such as Ga and Ge, alkalis).

10.5.3 The solar system

The solar system consists of a central star, the Sun, and a variety of bodies that orbit it. These are somewhat arbitrarily divided into planets, asteroids, comets, Kuiper Belt objects, and so on. Table 10.5 lists some data regarding the planets. The planets can be divided into three groups based on their size, density and composition. These are:

- The terrestrial planets:
Mercury
Venus
Earth-Moon
Mars
(asteroids)
- The gas giants:
Jupiter
Saturn
- The outer icy planets:
Uranus
Neptune

Seven of the eight planets have orbits that fall on a single plane, $\pm 3^\circ$. The same 7 have nearly circular orbits, with eccentricities all less than 0.1. Mercury's orbit is inclined some 7° with an eccentricity of 0.2. (Pluto's orbit is inclined 17° and has an eccentricity of 0.25; this highly anomalous orbit is one reason it is no longer considered a planet). Most major satellites of the planets also orbit in nearly the same plane. The Sun's equator is inclined some 7° to this plane. Thus the angular momentum vectors of major solar system objects are all rather

Table 10.5 Data regarding the sun, the planets, and their major satellites.

	Mass (kg)	Radial distance (AU)	Radius (km)	Density (g/cc)	1 atm density (g/cc)	Principal atmospheric components
Sun	1.99×10^{30}		6.96×10^5	1.4		
Mercury	3.35×10^{23}	0.39	2.44×10^3	5.42	5.3	–
Venus	4.87×10^{24}	0.72	6.05×10^3	5.24	3.95	CO ₂ , N ₂ , Ar
Earth	5.98×10^{24}	1.0	6.38×10^3	5.52	4.03	N ₂ , O ₂ , Ar
Moon	7.35×10^{22}		1.74×10^3	3.3	3.4	–
Mars	6.42×10^{23}	1.6	3.39×10^3	3.93	3.7	CO ₂ , N ₂ , Ar
Asteroids	4×10^{21}	2.8	$\leq 10^3$		3.4–3.9*	
Jupiter	1.90×10^{27}	5.2	6.99×10^4	1.31		H, He
Io	8.63×10^{22}		1.82×10^3	3.42		–
Europa	4.71×10^{22}		1.55×10^3	3.03		–
Ganymede	1.51×10^{23}		2.63×10^3	1.98		–
Callisto	1.06×10^{23}		2.40×10^3	1.83		–
Saturn	5.69×10^{26}	9.6	5.95×10^4	0.69		H, He
Titan	1.38×10^{23}		2.58×10^3	1.88		N ₂ , CH ₄
Uranus	8.73×10^{25}	19.1	2.54×10^4	1.30		H, He, CH ₄
Neptune	1.03×10^{26}	30.8	2.13×10^4	1.76		H, He, CH ₄
Triton	2.14×10^{22} (?)		1.35×10^3	2.08		–

* Densities of anhydrous chondrites.

similar, consistent with formation from a single rotating nebula. Rotational vectors of planets are generally inclined to their rotational vectors, some highly so, and Venus and Neptune have retrograde rotations (as does Pluto).

The definition of “planet” is somewhat open to debate. The asteroids have long been known to stably orbit the Sun, but are not considered planets. More than 100,000 asteroids are known and there may be millions. The largest is Ceres, discovered in 1801, which has a diameter of 950 km and a mass of 9.5×10^{20} kg. Pluto, once considered the ninth planet, is more than an order of magnitude smaller than Mercury and smaller than some of the major satellites of Jupiter, Saturn, and Neptune, and has a very anomalous orbit. It is more correctly considered a Kuiper Belt object. The Kuiper Belt, which lies between 30 to 50 AU from the Sun, is a great ring of debris, similar to the asteroid belt but of much lower density material – presumably dominated by hydrocarbons and ices of H₂O, CH₄, and NH₃ with lesser amounts of silicates. This region is thought to be the place of origin for short-period comets. There are estimated to

be over 70,000 Kuiper belt objects with a diameter greater than 100 km, with a total mass similar to that of the Earth. In the last several years, five new Kuiper Belt objects have been discovered that rival Pluto in size. *Eris*, discovered in 2003, with a radius of 1200 km, is the largest of these and is larger than Pluto. It has an elliptical orbit with semi-major axis of 68 AU, and has a moon of its own: *Dysnomia*. Sedna, also discovered in 2003, is the most distant. It has a mass, density, and radius in between those of Pluto and Chiron and a highly elliptical orbit with a semi-major axis of 937 AU. Sedna is distinctly red in color, possibly due to a surface composed of hydrocarbon sludge.

Finally, the Oort Cloud is a region from 50,000 to 100,000 AU where long-period comets are thought to originate. Comets appear to be low-density icy dust-balls. Based on spectral analysis of ejecta from Comet Tempel 1 produced by NASA’s Deep Impact mission, they appear to consist principally of water ice, HCN, a variety of hydrocarbons including polycyclic aromatic hydrocarbons, amorphous carbon, silicates, including olivine and pyroxene, spinel, carbonates, and

hydrated silicates. While most comets are quite small, a few km to tens of km in diameter, it is estimated that the total mass of the Oort cloud is between 5 and 100 Earth masses (M_E).

The terrestrial planets all consist of silicate mantles surrounding Fe-Ni metal cores and thin atmospheres that are highly depleted in H and He compared with the Sun. The gas giants Jupiter and Saturn are much more similar in composition to the Sun. The atmosphere of Jupiter is 81% H and 18% He by mass (compared with 71% and 28%, respectively, for the Sun), with CH_4 and NH_3 making up much of the rest. Saturn's atmosphere consists of 88% H and 11% He. The He depletion of both these atmospheres relative to the solar composition reflects a concentration of He in the interior. On the whole, the H/He ratio of Jupiter is close to the solar value. Elements heavier than He are about 5 times enriched in Jupiter compared with the Sun. The H/He ratio of Saturn as a whole is about a factor of 3 lower than the solar ratio and elements heavier than He are roughly 15 times enriched compared with the Sun. The nature of these planets' interiors is not entirely certain. Jupiter probably has a core consisting of liquid or solid metal and silicates with a mass roughly 15 times that of the Earth. Saturn probably has a similar core with a mass 100 times that of the Earth. Surrounding the core are layers of liquid metallic H and ordinary liquid H, both containing dissolved He. The icy planets consist of outer gaseous shells composed of H and He in roughly solar ratio with a few percent CH_4 surrounding mantles consisting of liquid H_2O , CH_4 , H_2S , NH_3 , H, and He, and finally liquid silicate-metal cores. Elements heavier than He are about 300 times enriched in Neptune and Uranus compared with the Sun. In a gross way, this compositional pattern is consistent with a radial decrease in nebular temperature: the terrestrial planets are strongly depleted in the highly volatile elements (e.g., H, He, N, C) and somewhat depleted in moderately volatile elements (e.g., K, Pb). From what can be judged from reflectance spectra, the asteroids also fit this pattern: the inner asteroids (sunward of 2.7 AU) are predominantly igneous and compositionally similar to the achondrites, which are highly depleted in volatile and moderately volatile elements. The

outer asteroids (beyond 3.4 AU) are richer in volatile elements and appear to be similar to carbonaceous chondrites.

The planet that we know the most about (other than Earth) is Mars. Observations on Mars and its composition come from (1) a variety of spacecraft orbiting Mars (as of 2012 there were three active) making spectroscopic measurements, (2) a variety of landers equipped with instruments capable of analyzing rocks and soil (to date there have been seven successful landers), and (3) the SNC meteorites (of which there are 5 falls and 87 finds). This has enabled the achievement of a vague understanding of the composition and history of Mars. Comparing the volatile inventory of Mars with that of Earth, Mars at first appears depleted in volatile elements. It has a much smaller atmosphere than the Earth (surface pressures are 0.006 atm). Like Venus, the Martian atmosphere is dominated by CO_2 , with N_2 as the second most abundant component. However, it is now clear from orbiter and surface observations that significant amounts of liquid water existed on the Martian surface during at least its first billion years or so, and there is evidence of some small ephemeral streams at present. To attain the necessary temperatures, Mars must have had CO_2 pressures at its surface of 5 to 10 atm. This early atmosphere has been lost, a consequence of several factors, including lower gravity and the lack of a geomagnetic field that serves to prevent erosion of the atmosphere by the solar wind. Thus the depletion of highly volatile elements on Mars may be partly a secondary feature. Nevertheless, Dreibus and Wänke (1987) estimated the water content of the Martian mantle to be an order of magnitude or so lower than that of the Earth. It is possible that Mars lost a larger fraction of the highly volatile elements during accretion than did the Earth.

Like the Earth, Mars has an iron core, but it appears to comprise a smaller fraction of planetary mass than Earth's. The Martian core has a diameter of 1400–1700 km and is about 15–20% of the mass of the planet. By comparison, the Earth's core is 32% of the mass of the planet. The smaller core appears to translate into a more iron-rich Martian crust and mantle and is likely due to accretion of Mars under more oxidizing conditions than the Earth, resulting in a lower Fe_{metal} .

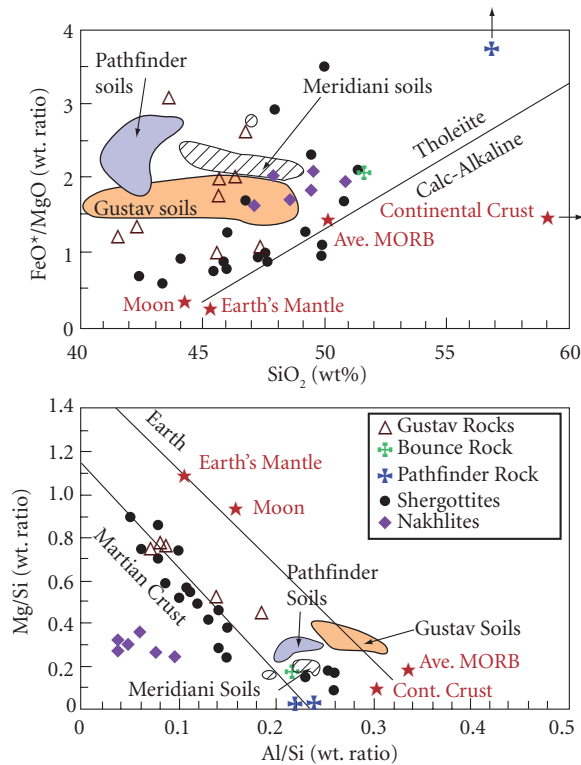


Figure 10.36 Major element chemistry of Martian meteorites and rocks and soils analyzed by NASA Martian landers. Mars appears to have a lower Al/Si ratio and a higher Fe/Mg ratio than the Earth. From McSween *et al.* (2009). Reprinted with permission from AAAS.

$\text{Fe}_{\text{silicate}}$ ratio. Both SNC meteorites and lavas analyzed by the NASA rovers are richer in Fe than comparable terrestrial basalts (Figure 10.36). Surface rocks on Mars are dominantly tholeiitic basalts formed by extensive partial melting; more differentiated rocks such as andesites and granites are uncommon or absent.

There is evidence that Mars is richer in moderately volatile elements than the Earth. Analyses of both Martian soil and the composition of SNC meteorites suggest a K/U ratio about 19,000, whereas this ratio is about 13,800 in the Earth. The significance of this ratio is that U is a highly refractory element while K is a moderately volatile one. Sr-Nd isotope systematics of SNC meteorites appear to define an array shifted to higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios compared with that of the Earth, implying Mars has a Rb/Sr ratio of about 0.07 compared with 0.03 for the Earth (Taylor,

1992). Pb isotope ratios of these meteorites indicate a $^{238}\text{U}/^{204}\text{Pb}$ ratio of about 5 for Mars, compared with the terrestrial value of ~ 8.5 . These ratios are consistent with the higher K/U of Mars, since Rb and Pb are both moderately volatile elements while Sr is a refractory one. Thus Martian moderately volatile/refractory element ratios appear to be systematically higher than terrestrial ratios.

The comparison of Venus with the Earth is particularly interesting, although we know considerably less about Venus than Mars. What we do know is based on gamma ray spectroscopy and analyses of the Venetian atmosphere by spacecraft. Although the two planets are of similar size, the Venetian atmosphere is almost 100 times more massive than that of the Earth. Whereas Venus's atmosphere is dominated by CO_2 , that of the Earth is dominated by N_2 and O_2 . However, the differences in both atmospheric mass and the abundance of CO_2 may reflect the difference in geological and, particularly, biological processes, on the two planets. Both planets appear to have similar relative abundances of carbon and nitrogen (Prinn and Fegley, 1989). In the case of the Earth, however, most of the surficial carbon is locked up in carbonates and organic carbon in rocks. This is in part due to biological activity, as is the presence of O_2 in the terrestrial atmosphere. There are, however, other differences that are more fundamental. The Venusian atmosphere is richer in noble gases than is the terrestrial one. Furthermore, the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of the Venusian atmosphere is about 1.15, whereas that of Earth is about 300. Since ^{40}Ar is produced by radioactive decay, it is related to the planetary K/Ar ratio. The terrestrial $^{40}\text{Ar}/^{36}\text{Ar}$ is high because the Earth captured relatively little primordial Ar; most of the Ar in the terrestrial atmosphere has been produced by decay of ^{40}K . Apparently, Venus captured much more primordial Ar and has a much lower K/Ar ratio than the Earth.

Although the Earth is depleted in noble gases relative to Venus, it appears to be richer in H_2O (Prinn and Fegley, 1989). Furthermore, the Earth and Venus appear to have similar K/U ratios, implying similar depletions in the moderately volatile elements (Taylor, 1991).

Thus in detail, we find the compositional differences between planets cannot simply be

explained by radially decreasing temperature. The noble gas-rich nature of Venus is just one example. Another example is Mercury. Considering its density, Mercury appears to be much richer in iron than the other terrestrial planets, which is difficult to explain as simply a temperature effect since Fe condenses at the same temperature as Mg silicates (Figures 10.33 and 10.34). In the outer Solar System, we find that although Uranus and Neptune are further from the Sun, they are poorer in H and He than Saturn and Jupiter.

10.5.4 Other solar systems

As of 2012, 809 confirmed planets had been discovered orbiting other stars and several thousand more were suspected but not yet confirmed. Multiple planets occur in about 15% of the cases, and as many as six planets have been found orbiting a single star. For stars having Fe/H ratios comparable to or greater than that of the Sun, planets have been found in about 15% of stars. The first extra-solar planets, or “exoplanets” to be discovered were found by detecting the “wobble” of their stars spectroscopically: an orbiting planet exerts a gravitational pull on the star and the resulting velocity change produces a Doppler shift in the light emitted by the star. To date, most exoplanets have been discovered by this method, which is most likely to detect large planets orbiting near the star. Subsequently, a variety of space telescopes have detected planets by direct imaging, infrared imaging, and occultation (dimming of the light of the star as a planet transits in front of it). The most promising mission is NASA’s Kepler Space telescope, which uses the transit method to detect candidate planets and has detected 105 confirmed planets and thousands of candidates in its first two years, including several planets comparable in size to Earth and Venus. Nevertheless, the range of sizes and orbital radii of known planets represents a highly skewed sample of all exoplanets. Consequently, at this point we can draw only a few conclusions from the discovery of exoplanets. First, they are not rare; second, unlike our own solar system, large planets can be present quite near the star. Thus planets and solar systems may be a normal consequence of star formation, but the distribution of planets in our solar system

is not necessarily typical. Finally, planets are generally substantially smaller than their central stars; they are not merely undersized companions. This latter observation suggests that planets form in a fundamentally different way than stars and isolated brown dwarfs.

10.6 BUILDING A HABITABLE SOLAR SYSTEM

We have now examined star and solar system formation from several perspectives: observations about the planets of our own solar system, observations about conditions in the early solar system derived from meteorites, observations about stars presently forming elsewhere in the galaxy, and attempts to reproduce processes and conditions in both laboratory or mathematical and computer simulations. Let’s first review how these observations constrain the process of formation of *our* solar system 4.56 billion years ago, and then try to deduce what actually happened back then.

10.6.1 Summary of observations

The planets in our solar system show a very strong compositional zonation. The four inner planets are strongly depleted in volatile elements, with some suggestion of increasing depletion inward. The two giant planets have compositions close to that of the Sun, indicating they formed from a nebula that had experienced little chemical fractionation. The outer icy planets appear slightly depleted in the most volatile elements (H and He), but are much more volatile-rich than the terrestrial planets.

Chondrites can be viewed as mixtures of four principal components: CAIs, chondrules, amoeboid olivine aggregates (AOAs), and matrix. The CAIs consist of highly refractory material, very similar in composition to what we would expect if chondritic dust were heated to the point where 95% of it evaporated, or, conversely, the first 5% of a gas of chondritic composition condensed. In either case, minimum *sustained* temperatures of ~1700 K are required (assuming pressures of around 10 Pa). AOAs appear to have condensed from a gas at only slightly lower temperatures. Chondrules also provide evidence of high temperatures, but in these cases the

peak temperatures must have been of short duration. Finally, chondritic groundmass material includes a variety of materials of which the most significant are presolar grains – ejecta of red giants and supernova. The grains have escaped nebular processing. While most such grains (e.g., graphite, SiC, diamond) are inherently refractory and hence would not necessarily be destroyed by high temperatures, these grains retain significant quantities of noble gases, suggesting they never experienced substantial heating.

Variations in the chemical composition of chondrites clearly indicate chemical heterogeneities within the solar nebula. Much of this is clearly related to volatility, implying significant variation in temperature in space and/or time in the nebula. Other variations relate to oxygen fugacity. Since H_2 is the principal reductant and it dominates the gas, while O constitutes a significant fraction of condensed matter, variation in oxygen fugacity most likely reflects variation in the ratio of gas to dust. In addition, there must have been significant variations in the metal/silicate ratio within the nebula to explain chondritic variations.

Planetesimals that were the parent bodies of achondrites and irons underwent sufficient heating to melt and differentiate. The oldest achondrites do not appear to be significantly younger than many chondrites. Thus planetesimals formed, melted, and differentiated within a few million years, at most, of the formation of chondrites.

We can also draw a number of conclusions from the isotopic compositions of meteorites. First, the isotopic composition of most elements in most meteorites and meteorite components is uniform. This contrasts with the isotopic heterogeneity of presolar grains preserved in some chondrites. From this we may conclude that, at least at radii less than or equal to that of the asteroid belt, the solar nebula was well mixed and extensively processed. Some isotopic variation is indicative of the presence of short-lived radionuclides when the solar system formed. Most were most likely produced in red giants or supernovae and injected into the molecular cloud that ultimately formed the solar system during or very shortly before its collapse. Other radionuclides, ^{10}Be in particular, were likely produced by spallation, perhaps by energetic

particles produced by the young Sun itself. Decay of these radionuclides, as well as conventional dating, suggests the time-scale for nebular processing and planet formation was a few million to a few tens of millions of years.

10.6.2 Formation of chondrite components

Chondrites are a mix of materials formed under different conditions in different environments. With some understanding of the star-forming process, we can begin to discern what these environments were, as illustrated in Figure 10.37.

As we found in Section 10.4.1, CAIs were the first-formed solids in our solar system. They represent material that condensed at temperatures of 1700 K or so. Many were subsequently reheated and subjected to partial melting and evaporative loss. The ages of such “processed” CAIs are indistinguishable from apparently “primary” CAIs, suggesting that the period of their formation was short, perhaps 50,000 years. They likely formed within 1 AU of the Sun as it transitioned from a Class I to Class II young stellar object.

Initial $^{26}Al/^{27}Al$ ratios in at least some amoeboid olivine aggregates are as high as in CAIs, suggesting they formed around the same time. They condensed from somewhat cooler nebular gas (~1400 K) than the CAIs.

Pb-Pb and extinct radionuclide dating indicates that chondrules formed 1–2 million years later than CAIs (and CB chondrules may have formed 4 to 6 million years later than CAIs) when the Sun had evolved to become a classical T-Tauri (class II) or weak-lined T-Tauri (class III) star. Heating experienced by chondrules was even more transient, lasting minutes to hours. This kind of transient thermal event matches well that expected for shock waves within the solar nebula (Desch and Connolly, 2002). The rapid cooling experienced by chondrules suggests ambient temperatures were low, perhaps 300 K. Overall, the conditions are consistent with those expected in the asteroid belt. Since chondrules make up much of the mass of primitive meteorites, we can infer that a very significant fraction of the solid matter in the inner solar system was processed in this way.

Finally, the chondrite matrix represents material that condensed over a range of

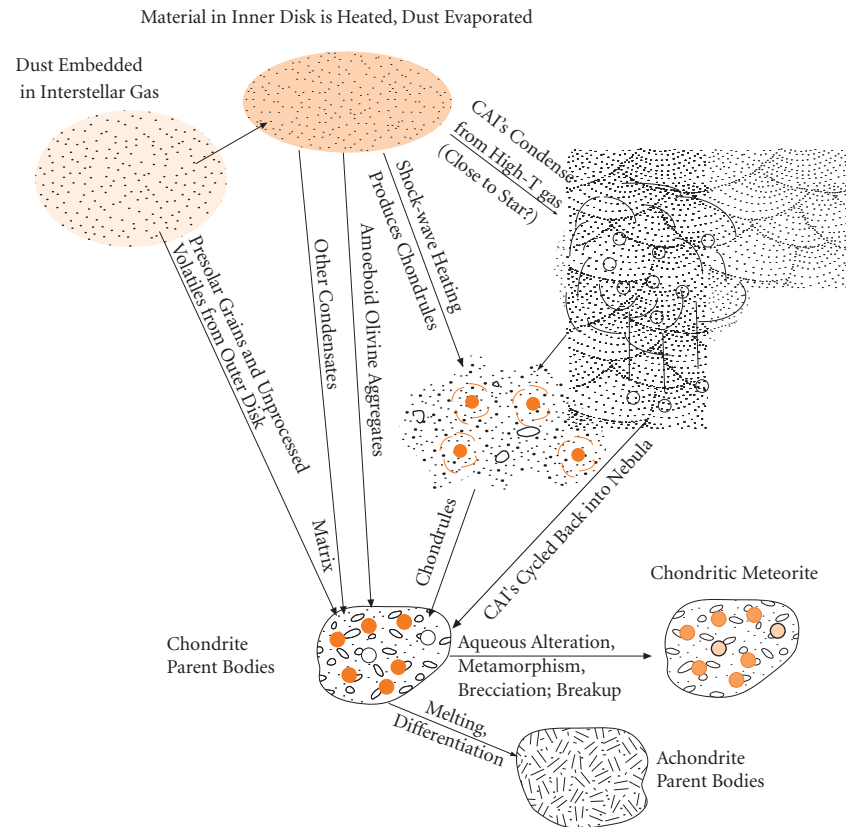


Figure 10.37 Cartoon illustrating the processes involved in the formation of chondrites and their components. McSween (1987). Reprinted with the permission of Cambridge University Press.

temperatures, including quite low ones, possibly at a variety of radial distances or at a variety of times as the nebula cooled. In addition, solid grains in the matrix reacted with nebular gas at low temperature. Subsequent reactions occurred on the parent bodies between components that condensed over a range of temperatures. The matrix also contains presolar grains that were incorporated into chondrites without significant processing in the solar nebula.

The compositional variation between chondrite classes can be partly explained by varying the proportion of these components. Variation in radial distance from the Sun and in vertical distance above or below the central plane helps to explain some of the remaining compositional variation. Gravity within the disk would tend to concentrate the dust component near the central plane of the disk; the gas to disk ratio would have increased vertically above and below the disk. This variation

effectively controls the ratio of H to O, and therefore the oxygen fugacity. More oxidized components formed near the central plane, while more reduced components formed some distance above or below it. Gravity, magnetic fields, gas drag, and other forces produced sorting between metal and non-metal grains, further enhancing compositional variation.

10.6.3 Formation of the planets

Unlike star formation, there are few astronomical observations to guide theories of planet formation, although possible planets have been imaged in very young stars still surrounded by nebular disks (e.g., Lagrange *et al.*, 2010). Hence current ideas on how the planets formed rest solely on theoretical modeling and observations of the present solar system: meteorites and the planets themselves.

Terrestrial planet formation can be divided into three stages: (1) dust condenses out of the

hot nebular disk; (2) dust grains grow by accretion from micron-sized particles to planetesimals and protoplanets; (3) protoplanets grow into planets via long-term, long-distance, cumulative gravitational interactions.

As we saw earlier, observed surface temperatures of nebular disks imply that interior temperatures in the inner parts of these disks are hot enough to vaporize much of the dust. On time-scales of perhaps 10^5 years or so, the inner disk cooled to the point where the silicates and iron recondensed, but temperatures remained warm enough so that ices of water, methane, ammonia, and so on, could not condense. At greater distances, the orbit of Jupiter and outward, the disk remained cool enough so that silicates and iron were never vaporized. More volatile compounds such as water might have been vaporized, but they recondensed long before the nebula dissipated. The radial distance beyond which ice condenses, ~ 160 K at prevailing pressures, from the nebula is known as the “snowline”. The snowline would have migrated inward as the nebula cooled.

In the inner solar system, the condensed dust comprised only about 0.5% of the mass of the nebula; in the outer solar system, the dust included ices and comprised 2% of the mass. Gravity caused the dust to settle quickly to the mid-plane of the nebula, so that its concentration there would be much higher (as we noted, the dust-to-gas ratio controlled oxygen fugacity – the mid-plane would have had higher oxygen fugacity). Van der Waals and electrostatic forces caused the dust grains to stick together and aggregate into “dust balls”. In the inner solar system many of these dust balls were melted to form chondrules, probably by gas drag during shockwaves. Turbulence in the gas may have also concentrated the dust in eddies. Dust accumulation would have proceeded more rapidly in the outer, cooler parts of the solar system beyond “snow line”. Nebular evolution models generally place the snow line between the asteroid belt and the orbit of Jupiter. Much of the water and other ices probably condensed on pre-existing dust particles. This has two effects: first, it enhances the overall concentration of solids, and second, it greatly increases the tendency of grains to stick to each other.

This process produced millimeter- to centimeter-sized particles, the size of CAIs

and chondrules. How these particles grew into kilometer-sized blocks, where gravity began to play a role, is less certain. One possibility is that gravitational instabilities developed in the dust and km-sized blocks developed by gravitational attraction. It seems likely, however, that gas turbulence would be sufficient to prevent this. What is clear is that the transition from dust to km-sized bodies must have happened quickly or the dust would have been swept into the growing Sun. Because of the thermal motion of gas molecules, the gas rotates more slowly than the velocity necessary to keep solid bodies in stable orbit: the Keplerian velocity. Thus gas drag slows the particles embedded within it, and they spiral inward toward the Sun. The effect is negligible for micron-sized dust and bodies of a km or more, but it is quite significant for meter-sized blocks, which will spiral into the Sun from the Earth’s orbit in 100 yrs. Thus growth through this stage must be fast if any solids are to form planets. Further evidence of rapid formation of planetesimals comes from Hf-W ages of some iron meteorites, suggesting they formed only ~ 1 Ma after formation of CAIs. Thus at least some planetesimals must have accreted, melted, and differentiated into silicate and iron parts very quickly, perhaps during, or even before, the time of chondrule formation.

Two mechanisms for rapid growth of planetesimals have been recently proposed. In the first, transient high-pressure regions in the disk concentrate meter-sized boulders, and this concentration of solids then affects flow of the gas so as to capture solids drifting inward. Computer simulations suggest this process leads to gravitationally bound “clumps” that collapse to form planetesimals on time-scales of a few orbital periods, that is, a few years to a few decades (Johansen *et al.*, 2007). The second mechanism is concentration of particles in low-vorticity (i.e., slowly rotating) eddies within the disk. Simulations show that particles can become sufficiently concentrated in these eddies to become gravitationally bound and eventually contract to form ~ 100 km size planetesimals (Chambers, 2010). This process could form planetesimals in a few thousand years beyond the snow line, but might require a few million years inside it.

Early-formed planetesimals larger than 1 km would have been heated by decay of ^{26}Al such that interior temperatures would have reached 1500 K within $\sim 500,000$ years and thus would have undergone metal-silicate differentiation. Such bodies are the likely parent bodies of the achondritic and iron meteorites. Collisions between planetesimals would in some cases lead to growth of ever-larger bodies. But collisions would also have destroyed some planetesimals, with the debris replenishing the nebular dust, from which a new generation of planetesimals could grow. Later-formed planetesimals might have avoided melting and differentiation: these are likely the parent bodies of the chondritic meteorites.

Once planetesimals have formed, gravity becomes important and bodies grow by collision. This process is termed “*oligarchic growth*” because it ends with relatively few large objects. It has been extensively modeled with computer simulations and is comparatively well understood. Because larger bodies have larger effective cross-sections (their gravity pulls in other bodies from a wider area), the largest bodies grow fastest. The larger bodies also tend to acquire the most regular orbits, which leads to gravitational focusing and further enhancement of growth rates. This leads to very rapid growth in the early stages. As a relatively few large objects become dominant, growth slows somewhat. Models suggest that bodies the size of the Moon ($0.01 M_{\oplus}$ or Earth masses) or Mars ($0.1 M_{\oplus}$) could have formed in the inner solar system within 10^5 to 10^7 years. The Hf-W system may provide a tighter constraint on this stage. Hf-W data on SNC meteorites suggest that Mars had formed and differentiated into a silicate mantle and iron core about 2 million years after CAI formation (Dauphas and Chaussidon, 2011). However, the HED parent body (*Vesta*) and the angite parent body, both of which are considerably smaller than Mars, apparently formed and differentiated later. Thus planetary growth may have proceeded at different rates in different parts of the solar nebula.

Once Mars-sized bodies have formed, only a few large bodies are present within any planet-forming zone, and collisions between them become infrequent. Consequently, growth slows dramatically. Indeed, simula-

tions suggest it might have required an additional 10^8 years for bodies the size of the Earth or Venus to form (Wetherill, 1990; Wetherill and Stewart, 1993; Chambers and Wetherill, 1998). Another feature of the late stages of accretion is that the collisions involve very large bodies and are consequently catastrophic. The energy released in these collisions leads to extensive melting. As we shall see later, this fits very well with the evidence that we have for the origin of the Earth and the Moon.

There is somewhat more uncertainty as to how the giant planets formed. There are essentially two classes of theories. The first and best established is that of *core accretion*. It proposes that rocky, icy cores of giant planets accreted in a process very similar to that described earlier, albeit enhanced by the presence of ice beyond the snow line. Once these cores reached a size of 10 Earth masses, they would have had sufficient gravity to capture gas from the solar nebula and eventually become gas giants. This theory nicely explains why the gas giants have enhanced concentrations of condensable elements compared with the Sun. On the other hand, simulations assuming a “minimum mass nebula” take much too long to form giant planets in this way – 10^8 years or more. As we saw, it appears that nebular disks dissipate within at most 10 Ma – the gas giants must have captured their gas shrouds within this time. The “minimum mass nebula” has just enough mass to produce the planets observed, which is the mass of the planets plus the non-condensable fraction missing from them, and is about $0.01 M_{\odot}$. However, in simulations involving more massive nebulae, $>0.1 M_{\odot}$, gas giants do form in this way within 10 Ma. The other theory, a relative newcomer, is gas instability (Boss, 1997). It posits that a density perturbation in the disk could cause a clump of gas to become massive enough to be self-gravitating. Once that happens, the clump could collapse into a planet on time-scales of 10^3 yrs. These density perturbations occur in simulations with nebular masses significantly greater than the minimum mass. One criticism of this model is that it fails to explain why the giant planets are depleted in H and He, and does not readily explain why these planets have rocky cores. Either way, it appears planet formation is an inefficient

process, with perhaps 90% the original nebular mass being either swept into the Sun or lost to space. Very massive nebulae, containing a solar mass or more and which were once popular among theorists, have fallen out of favor.

Within 10^7 years, the nebula cleared. This resulted from a variety of factors. While the planets were forming, gas and dust were steadily spiraling into the central star. Small bodies that were not swept up by the forming planets would have been flung, through gravitational interaction with the planets, into the central star or out to interstellar space. In the outer solar system, the giant planets swept up large amounts of gas. As the Sun became hotter and more luminous, the nebula would have dissipated through “photoevaporation”. (This term is a bit of a misnomer, since most of the nebula is gas to begin with, it does not evaporate in the conventional sense. But it does absorb radiational energy, which is converted to kinetic energy through collisions. The fast-moving gas molecules then escape to interstellar space.) Finally, the enhanced solar winds of the T-Tauri stage would have helped to drive out remaining gas and dust. As a result, the nebula clears mostly from the inside out. However, nearby large stars could also erode the nebula from the outside through photoevaporation. This is observed to occur in the Orion nebula. The nebula must have cleared of gas before the inner terrestrial planets were able to accumulate their full share of volatile elements, perhaps because the inner solar system cleared before it cooled enough for these elements to condense. It may have cleared before the ice giants, Uranus and Neptune, were able to capture their full complement of gas.

In many solar systems, giant gas planets have orbits smaller than that of Mercury. It is highly unlikely that gas giants could form so close to their central star; rather, astronomers believe they must have migrated inward from where they formed. How and how fast the nebular disk clears may affect giant planet migration. Where outer disks are short-lived due to erosion, as in Orion, migration may be minimal, leaving giant planets in the outer part of their solar systems. Long-lived outer disks (as in the Taurus star-forming region) should allow significant giant planet migration and lead to the “hot Jupiters” observed

to orbit other stars (Boss, 2005b). Presumably, our solar system is an example of the former, in which the giant planets presently orbit close to where they formed. Even minor inward migration, however, could have had consequences for orbits of other planets and planetesimals. Giant planet migration is one possible explanation for the “late heavy bombardment” discussed in the next section.

10.6.4 Chemistry and history of the Moon

Before we consider the formation of the Earth specifically, there is one more set of observations we need to consider: observations about the Moon. The Moon is of interest for several reasons. First, it is the only solar system body, other than the Earth, that we have returned samples from and has been explored by men. From this exploration we have found that much of the earliest history of the Moon is preserved, giving us unique insights into early planetary history. In contrast, no trace remains of the earliest period of Earth history (some SNC meteorites date to 4.5 Ga, implying that evidence of early Martian history might be preserved on its surface, but we have yet to explore and sample those parts of Mars). Second, the Moon is closely associated with the Earth, not only physically, but chemically as well. Lunar oxygen isotope compositions fall on the terrestrial fractionation line, implying the Moon and Earth share the same O isotopic composition. The Moon is nearly unique in this respect: with the exception of E-chondrites, all analyzed meteoritic material, including the SNC meteorites, falls off this line. This strongly implies that the Moon and the Earth are closely related. As we shall see, the Moon holds important clues to the earliest part of Earth’s history.

10.6.4.1 Geology and history of the Moon

The Moon can be divided into three geologic provinces: the highlands – mountainous regions apparently consisting largely of anorthosite; the uplands – areas of mild relief covered by a blanket of ejecta from large impacts; and the Mare – the large craters filled with basaltic lavas. Much of the surface of the Moon is covered with fine debris of impacts, called the regolith, consisting of rock and mineral fragments, glass, and some meteorite

Table 10.6 Representative compositions of lunar rocks.

	Highland anorthosite	Highland anorth. gabbro	Mare qz. tholeiite	Low-K Mare basalt	Fra Mauro KREEP
SiO ₂	44.3	44.5	46.1	40.5	48.0
TiO ₂	0.06	0.39	3.35	10.5	2.1
Al ₂ O ₃	35.1	26.0	9.95	10.4	17.6
FeO	0.67	5.77	20.7	18.5	10.9
MnO	—	—	0.28	0.28	—
MgO	0.80	8.05	8.1	7.0	8.7
CaO	18.7	14.9	10.9	11.6	10.7
Na ₂ O	0.8	0.25	0.26	0.41	0.7
K ₂ O	—	—	0.07	0.10	0.54
P ₂ O ₅	—	—	0.08	0.11	—
Cr ₂ O ₃	0.02	0.06	0.46	0.25	0.18
Total	100.5	99.9	100.3	99.7	99.4

particles. For the most part, it seems to be locally derived; thus the regolith in the Mare differs from that of the highlands, although large impacts would have showered debris over large regions. Basalt from the Mare encompasses a variety of magma types, including both incompatible-rich and incompatible-poor types, and both quartz-normative and olivine-normative tholeiites. Highland rocks include anorthosite (nearly monomineralic calcic plagioclase), anorthositic gabbro (plagioclase and pyroxene with lesser amounts of olivine), dunite, and K-rich basalts. The highlands are extremely brecciated; most of these rock types have been found only as clasts in breccias. Table 10.6 shows some representative compositions of lunar rocks, and Figure 10.38 shows rare earth patterns of the same rock types.

Most of the lunar Mare are thought to have been created by large impacts between 4.2 and 3.8 Ga. Subsequently, the Mare were flooded by basalt to a depth of 5–10 km. These were partial melts generated at 100 or so km depth. The flooding occurred over an extensive time: 3.9–3.1 Ga. Mare flooding was the last major lunar geologic event. Subsequent to that time, the only activity has been continual bombardment by meteorites and asteroids, which continued to produce minor disruption of the surface and build-up of the regolith, and rare volcanism.

Figure 10.39 illustrates the highlights of lunar history. The oldest lunar rocks are nearly as old as meteorites. These are clasts

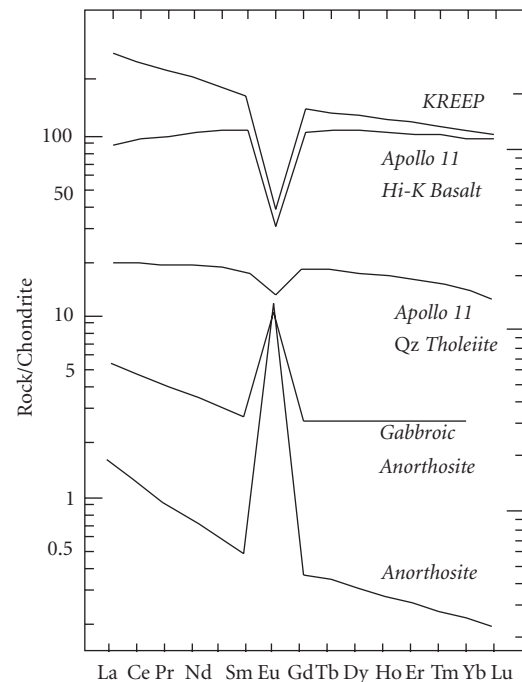


Figure 10.38 Rare earth patterns of representative lunar rocks. Taylor (1975). With permission from Elsevier.

of anorthositic highland rocks found in breccias. This suggests lunar differentiation began about 4.5 billion years ago. But most highland rocks have ages between 3.9 and 4.0 Ga. For the most part, these ages are interpreted as (and sometimes can be shown to be) the time of impact metamorphism. Apparently

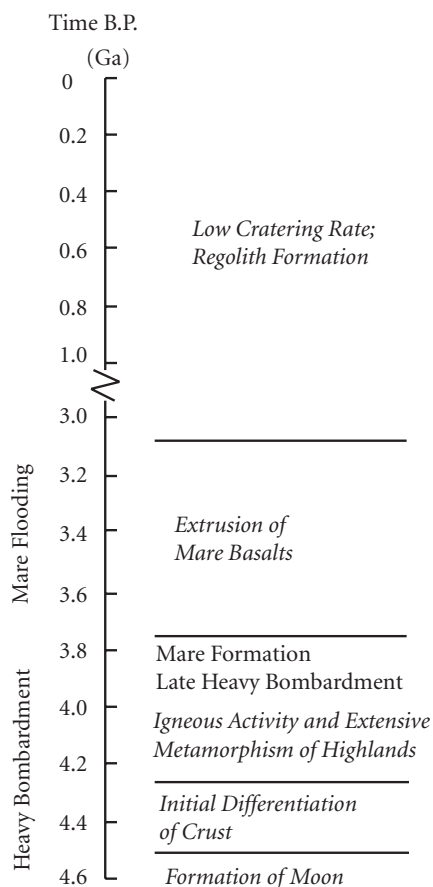


Figure 10.39 Highlights of lunar chronology.

the Moon suffered very heavy bombardment by meteorites and asteroids as late as 3.9 Ga. The clustering of ages around this time gives rise to the idea of a “late heavy bombardment” of the Moon.

One of the more important conclusions derived from study of the Moon is that it underwent very extensive melting just after its formation, perhaps forming a magma ocean 100 km or more deep. The anorthosite of the highlands is thought to have originated by plagioclase flotation in the magma ocean (i.e., anorthosite “icebergs” forming the lunar crust). The lunar crust seems to have been largely in place within 100 to 200 Ma after the Moon’s formation. Fractional crystallization of this magma ocean and flotation of plagioclase accounts for the general Eu depletion (Figure 10.38) observed in basalts derived from the lunar mantle. A basalt particularly enriched in incompatible elements, called KREEP (K-REE-P: potassium, rare earth

Table 10.7 Comparison of the compositions of the Earth and Moon.

	Bulk Earth	Bulk Moon	Silicate Earth	Silicate Moon
SiO ₂	30.38	43.4	45.0	44.4
TiO ₂	0.14	0.3	0.201	0.31
Al ₂ O ₃	3.00	6.0	4.45	6.14
FeO	5.43	10.7	8.05	10.9
MgO	25.52	32	37.8	32.7
CaO	2.40	4.5	3.55	4.6
Na ₂ O	0.24	0.09	0.36	0.09
K ₂ O	0.02	0.01	0.029	0.01
Fe	28.43	2.166		
Ni	1.75	0.134		
S	1.62			
Core %	32.5%	2.3%		
Mantle %	67.5%	97.7%		

Terrestrial composition from McDonough and Sun (1995), with permission from Elsevier, assuming the light element in the core is S. Lunar composition from Taylor (1992).

element, and phosphorus enriched), is thought to be derived from a part of the mantle that was the last part of the magma ocean to crystallize and was consequently strongly enriched in incompatible elements.

10.6.4.2 Composition of the Moon

Table 10.7 compares the composition of the Earth and the Moon. There are several notable differences. First, the Moon is depleted in moderately volatile elements (Na, K) compared with the Earth. Second, although the silicate part of the Moon (mantle + crust) is richer in iron than the silicate Earth, the bulk lunar composition is poorer in iron than the Earth. The latter reflects the small size of the lunar core, which has a diameter of only 340 ± 90 km, representing only about 2% of the mass of the planet. The lunar depletion in Fe compared with the Earth extends to all siderophile and chalcophile elements as well. The Moon is also much more depleted in the volatile elements. Conversely, the Moon is enriched the refractory lithophile elements, having nearly twice the Al and Ca as the Earth. Despite these compositional differences, the similarity of terrestrial and lunar oxygen isotope compositions strongly suggests they formed from the same part of the solar nebula.

10.6.5 The giant impact hypothesis and formation of the Earth and the Moon

The Earth and the Moon are physically closely associated, so one might suspect their origins are related. As we saw, there is evidence that the Moon was extensively melted very early in its history. We saw that numerical simulations of planetary growth within the solar nebula predict collisions between very large bodies in the final stages of accretion. We have also seen that the Moon and the Earth share an identical oxygen isotopic composition. In addition, the Earth–Moon system has an anomalously large amount of angular momentum compared with the other planets, and most of this angular momentum resides in the Moon. Finally, the Earth’s rotational axis is tilted 23° with respect to its orbital plane. These observations led Hartmann and Davis (1975) to propose that the Earth collided with a very large body in the final stages of its accretion. A substantial fraction, perhaps 2%, of the mass of the two bodies was blasted into orbit around the Earth, with much of this debris later coalescing to form the Moon. This idea has become known as the *giant impact hypothesis* and is now widely accepted (even if some of the details are still debated). The energy released in this impact would have been sufficient to melt the entire mantle of the Earth and to vaporize some of it, though this depends on how efficiently the impact energy is dissipated.

Detailed models of the impact have been produced (e.g., Newsom and Taylor, 1989). The model can be best summarized as follows. During the latest stages of accretion, the Earth was struck at low angle, and relatively low velocity (5 km/s), by a body slightly larger than Mars. The impactor, called *Theia* by some, presumably shared the depletion in highly and moderately volatile elements that characterizes the terrestrial planets. Metallic cores had already formed in both the Earth and the impactor. The collision completely disrupted the impactor and partially disrupted the Earth’s mantle, with much of the resulting debris having gone into orbit around the Earth. Most of the disrupted core of the impactor quickly (a matter of hours) accreted to the Earth and the silicate material in orbit more slowly coalesced to form the Moon. About 85% of the material forming the Moon was derived from the impactor, the

remainder from the Earth’s mantle. A small core then segregated from the largely molten Moon. Most of the core of the impactor quickly sank through the disrupted mantle of the Earth to coalesce with the existing terrestrial core. The depletion of siderophile elements in the Moon is thus a result of its formation from the silicate portions of the impactor and Earth, in which the siderophiles were already depleted by core formation. The volatile element depletion of the Moon is a consequence of the evaporation of these elements during impact.

10.6.6 Tungsten isotopes and the age of the Earth

If the giant impact hypothesis is correct, then the impact marks the time of final segregation of the metal and silicate in the Earth (and the Moon) as well as the completion (or nearly so) of accretion of the Earth. If we can date this event, we can determine the age of the Earth. Looking at Table 10.4, we see that the ^{182}Hf – ^{182}W decay pair is just the ticket: Hf is lithophile and should be concentrated in the mantle while W is siderophile and should be concentrated in the core. Furthermore, the half-life of ^{182}Hf is 9 Ma, just right for examining events in the first few tens of millions of years in the solar system. Furthermore, both are highly refractory elements, which has the advantage that one can reasonably assume that bodies such as the Earth should have an approximately chondritic Hf/W ratio. Over the last dozen years this system has provided very interesting insights into the early history of Mars, the Moon, and the Earth. Among other things, the story of Hf–W illustrates the importance of the fundamental dictum in science that results need to be independently replicated before they are accepted.

Because the variations in $^{182}\text{W}/^{183}\text{W}$ ratio are quite small, they are generally presented and discussed in the same ϵ notation used for Nd and Hf isotope ratios (Chapter 8). However, ϵ_{W} is the deviation in parts per 10,000 from a terrestrial tungsten standard, and $f_{\text{Hf/W}}$ is the fractional deviation of the Hf/W ratio from the chondritic value. Assuming that the silicate Earth has a uniform W isotope composition identical to that of the standard (an assumption supported by the observation that, with one exception,* all terrestrial samples analyzed

thus far have this composition), then the silicate Earth has ϵ_W of 0 by definition. The basic question can be posed this way: if the $^{182}\text{W}/^{183}\text{W}$ ratio in the silicate Earth is higher than in chondrites, it would mean that much of the Earth's tungsten had been sequestered in the Earth's core before ^{182}Hf had entirely decayed. Since the half-life of ^{182}Hf is 9 Ma and using our rule of thumb that a radioactive nuclide is fully decayed in 5 to 10 half-lives, this would mean the core must have formed within 45 to 90 million years of the time chondritic meteorites formed. If, on the other hand, the $^{182}\text{W}/^{183}\text{W}$ ratio in the silicate Earth was the same as in chondrites, which never underwent silicate-metal fractionation, this would mean that at least 45 to 90 million years must have elapsed (enough time for ^{182}Hf to fully decay) between the formation of chondrites and the formation of the Earth's core.

"Anomalous" W isotopic compositions were first found in the IA iron *Toluca* by

Harper *et al.* (1991). They found the $^{182}\text{W}/^{183}\text{W}$ ratio in the meteorite was 2.5 epsilon units lower (and subsequently revised to 3.9 epsilon units lower) than terrestrial W. Lee and Halliday (1995) reported W isotope ratios for two carbonaceous chondrites (*Allende* and *Murchison*), two additional iron meteorites, and a lunar basalt. They found the iron meteorites showed similar ^{182}W depletions as observed in *Toluca*. The chondrites, however, had ϵ_W values that were analytically indistinguishable from "terrestrial" W, as was the lunar basalt. Based on this similarity of isotopic compositions of chondritic and terrestrial W, Lee and Halliday (1995) concluded that the minimum time required for formation of the Earth's core was 62 million years. Subsequently, Yin *et al.* (2002) and Kleine *et al.* (2002) reported W isotope ratios in carbonaceous chondrites that were 1.9 to 2.6 epsilon units lower than the terrestrial standard (Figure 10.40). Furthermore, Kleine *et al.* (2002) analyzed a variety of terrestrial materials and found they

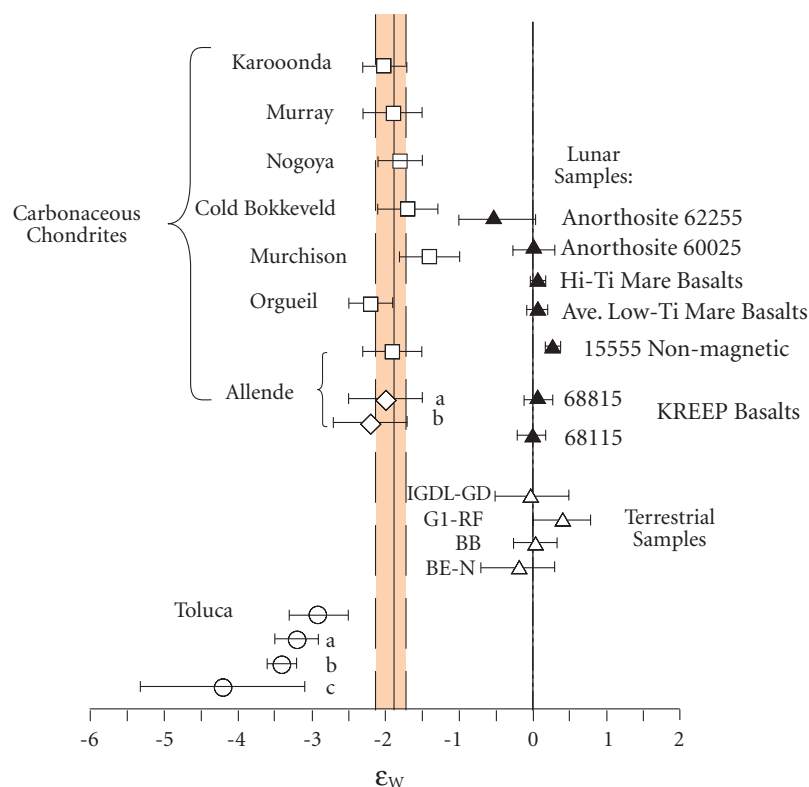


Figure 10.40 W isotope ratios measured in meteorites and terrestrial and lunar materials. Data from Kleine *et al.* (2002), Yin *et al.* (2002), and Touboul *et al.* (2007, 2009).

*Touboul *et al.* (2012) reported that the 2.8 billion year old Kostomuksha komatiites have a ^{182}W excess of 15 ppm. The most likely explanation is that the mantle had not been completely homogenized by that time.

all had identical W isotopic compositions that are some 2 epsilon units higher than those of the chondrites (Figure 10.40). It thus appears that the original measurements of Lee and Halliday (1995) were wrong. The measurement error most likely relates to what was at the time an entirely new kind of instrument, namely the multi-collector ICP-MS. Touboul *et al.* (2007) showed that the Moon has an isotopically homogenous W composition that is identical to that of the silicate Earth – about 2 epsilon units higher than chondrites. This episode illustrates why science places such importance on *independent replication* of observations (Chapter 1).

What the high $^{182}\text{W}/^{183}\text{W}$ ratio of the silicate Earth relative to chondrites implies for the timing of formation of the Earth's core depends on how the core formed. In the simplest model, the core segregated instantaneously from homogeneously accreted metal and silicate. The ^{182}W excess of the Earth's mantle of $\epsilon_{\text{W}} = +1.9$ and a Hf/W ratio of 17 then indicates this happened ~ 30 Ma after the time of CAI formation. This model is unlikely, however, since we know that metal cores segregated from even relatively small planetary embryos only a few million years after the start of the solar system. Large planets such as the Earth formed by collisions between and accretion of these already-

differentiated planetary embryos. Thus most studies have considered continuous core formation models. In such models, there is no single “age” for core formation; rather, the reported ages correspond to either the mean time of core formation (formation of 63% of the core) or the time of the Moon-forming impact, which most likely determined the termination of the major stage of Earth's accretion and the last time the core and mantle could equilibrate. In continuous formation models, W isotopic evolution will depend on the degree to which tungsten in the silicate and metal parts of colliding bodies equilibrate. Thus tungsten isotope systematics have been variously interpreted as implying core formation ages anywhere from 30 Ma to >100 Ma after CAI formation. However, the identical $^{182}\text{W}/^{184}\text{W}$ ratios of the Moon and silicate Earth, despite the two bodies having different Hf/W ratios, indicate that the Moon-forming giant impact occurred after extinction of ^{182}Hf , or more than 60 Ma after CAI formation. The minimum age of the Moon is constrained by Sm-Nd ages of the lunar anorthositic crust. This age is 4.456 ± 0.040 Ga, or about 100 Ma after the beginning of the solar system. Touboul *et al.* (2007) thus estimated an age for the Moon of $62 \pm 90/-10$ Ma after the start of the solar system (Figure 10.41).

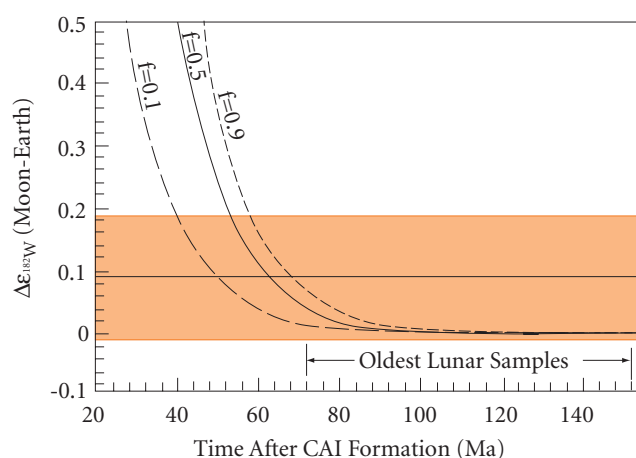


Figure 10.41 Difference between ϵ_{W} of Earth and Moon (Δ) versus the age of the giant impact (and presumably final core formation of the Earth). f is the relative difference in Hf/W ratios between the silicate part of the Earth and the silicate part of the Moon. The gray area shows the best value of $\Delta\epsilon_{\text{W}}$ (0.09 ± 0.10). The Hf/W ratio of the Earth's mantle is estimated at 18 and that of the Moon is 26.5, corresponding to an f of 0.47 (solid line). Reprinted by permission from McMillan Publishers Ltd: Touboul *et al.* (2007).

We will see in the following chapter that the Earth has a composition that is different from that of the Sun and of chondrites, particularly with respect to volatile and moderately volatile elements. This distinctive composition reflects its formation in the warm

inner region of the solar nebula, but perhaps also the cataclysmic events in the final stages of its formation. The final energetic cataclysm, the giant impact, may be partly responsible for the volatile-poor nature of the Earth as well as the Moon.

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