8 Overview of the Solar System

Our aim over the next four chapters is to understand the basic characteristics of the solar system, and to integrate this information into a self-consistent picture of how the solar system formed and how it has evolved with time. Some basic facts should be kept in mind as we pursue our goal:

- The Sun contains 99.8% of the mass in the solar system.
- Most of the remaining 0.2% of the mass is confined to a flattened disk. Within this disk, all the planets revolve in the same direction, most (but not all) of the planets rotate in the same direction, and all objects have similar ages ($t \sim 4.6$ billion years, when measurable).

Figure 8.1 shows a plot of mass versus orbital semimajor axis for the largest objects known to be orbiting the Sun (in the case of the smaller objects, the mass is a rough estimate). The eight largest objects orbiting the Sun have been given the collective name **planets**. The rocky and metallic objects in the asteroid belt, lying primarily between the planets Mars and Jupiter, are called **asteroids**; the largest asteroid, Ceres, is also a **dwarf planet**, according to the definition approved by the International Astronomical Union.[1] The icy objects beyond the orbit of Neptune are called **trans-Neptunian objects** (TNOs); many of the known TNOs are in a region just beyond Neptune called the Kuiper belt. The largest known TNOs—Eris, Pluto, Haumea, and Makemake—are also dwarf planets.

8.1 ▪ TWO TYPES OF PLANETS

The planets can be divided into two major types, or families, each named after their largest member. The smaller **terrestrial** planets, named after the Earth (alias Terra), are closer to the Sun; the larger **Jovian** planets, named after Jupiter (alias Jove), are farther

[1] A dwarf planet must be large enough to be squeezed by its own self-gravity into a spherical shape, but not large enough to be gravitationally dominant in the region near its orbit. We discuss the definition of "planet" and "dwarf planet" in more detail in Section 11.2.

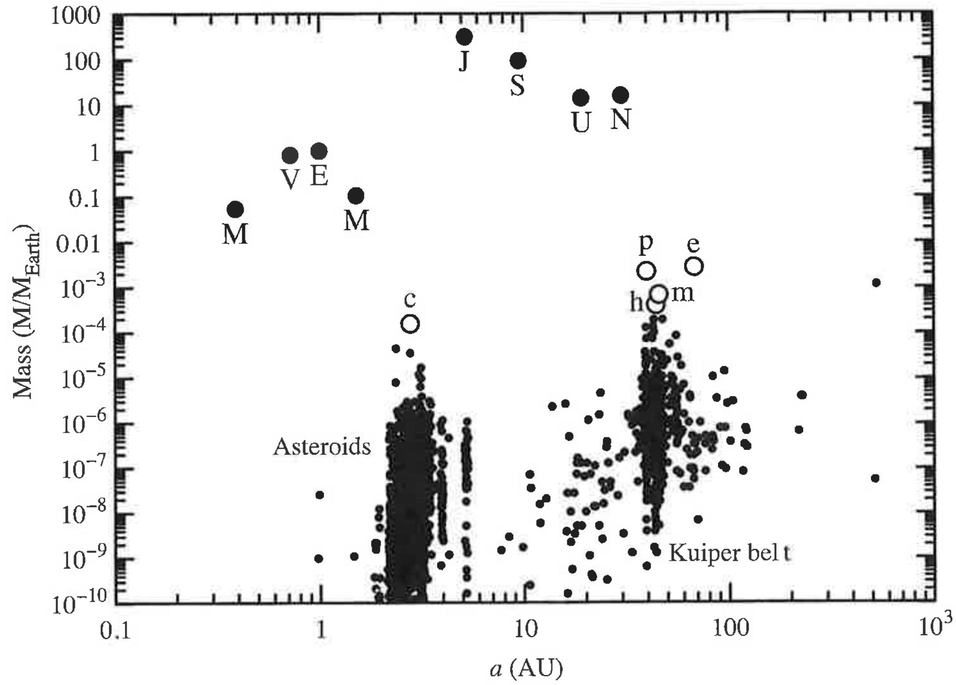

FIGURE 8.1 Estimated mass (in units of the Earth's mass) as a function of semimajor axis for bodies orbiting the Sun. Planets are labeled with their initials. (The dwarf planets Ceres, Pluto, Makemake, Haumea, and Eris are labeled with lowercase initials.)

from the Sun.[2] The characteristics of the two families are compared and contrasted in Table 8.1.

Any useful theory for the origin of the solar system must explain the observed differences between terrestrial and Jovian planets. In addition, the following questions should be addressed:

- Why are planetary orbits nearly circular?
- What is the nature and origin of the small "debris," such as comets and asteroids?
- What is the origin of the planetary satellites?
- Why are there differences in chemical composition among bodies in the solar system?

[2] The dwarf planets Ceres, Pluto, Makemake, Haumea, and Eris don't fit into this scheme. Ceres resembles the rocky satellites of the solar system, while Pluto resembles the icy satellites. The properties of Makemake, Haumea, and Eris are as yet poorly known, but their high albedo, or reflectivity, indicates that they have icy surfaces.

TABLE 8.1 Characteristics of Planetary Types

Characteristic	Terrestrial	Jovian
Mass	Low ($\leq 1\,M_\oplus$)	High ($> 10\,M_\oplus$)
Composition	Rocky/metallic	Gaseous/icy
	($\rho \gtrsim 3000\,\mathrm{kg\,m^{-3}}$)	($\rho \lesssim 2000\,\mathrm{kg\,m^{-3}}$)
Rotation	Slow ($P \geq 24$ hr)	Fast ($P < 18$ hr)
Satellites	Few	Many
Distance from Sun	$a < 2$ AU	$a > 5$ AU

- Why are there rings around the Jovian planets (and not around the terrestrial planets)?
- Why are the terrestrial planets chemically differentiated, with a rocky outer layer wrapped around a metallic core?

We will first describe the physical characteristics of the constituents of the solar system, introducing new physical concepts as necessary. We will then try to put these elements into a single coherent picture of the solar system.

8.2 ▪ PHYSICAL PROPERTIES OF PLANETS

The masses of astronomical objects are measured by looking at how they accelerate neighboring objects. It is particularly easy to measure the mass of a planet when it has a nearby small satellite; in that case, it is a simple matter of applying Kepler's third law (equation 3.54):

$$M_{\mathrm{planet}} \approx \frac{4\pi^2 a^3}{GP^2}, \tag{8.1}$$

where a and P are the semimajor axis and period of the satellite's orbit. Planets that do not have satellites (that is, Mercury and Venus) pose a more difficult problem. Before the advent of interplanetary space probes, the masses of Venus and Mercury were determined from their (tiny) perturbations to the orbits of other planets. Much more accurate masses are now available from space probes that have flown by them or orbited them.

The radius R of a planet is computed from its measured angular radius and its distance.[3] The mean density can then be computed from the known radius and mass:

$$\rho = \frac{3M}{4\pi R^3}. \tag{8.2}$$

Densities of $\rho \sim 700 \rightarrow 2000\,\mathrm{kg\,m^{-3}}$ are typical of Jovian planets, indicating a composition that is mostly gas or ice, where "ice," in the language of astronomers, can

[3] Distances within the solar system can be measured accurately using radar, as described in Section 13.1.

refer to frozen water, but also to frozen carbon dioxide ("dry ice"), frozen methane, and frozen ammonia. The lowest-density Jovian planet, Saturn, has a mean density $\rho \approx 710\,\mathrm{kg\,m^{-3}}$, which is less than that of water ($\rho = 1000\,\mathrm{kg\,m^{-3}}$).[4] Densities of $\rho \sim 3000 \rightarrow 5500\,\mathrm{kg\,m^{-3}}$ are typical of terrestrial planets, indicating a composition that is mostly rock and metal.

The surface temperature T of objects in the solar system depends on a number of factors. Since the Sun is an important source of energy, the temperature of an object depends on its distance from the Sun and on its **albedo**, or reflectivity.[5] However, the surface temperature also depends on whether the object has any internal heat sources, and whether it has an atmosphere that can act as an insulating blanket wrapped around its surface.

The Sun, to a rough approximation, radiates like a blackbody. The Sun's luminosity is thus

$$L_\odot = 4\pi R_\odot^2 \sigma_{\mathrm{SB}} T_\odot^4, \tag{8.3}$$

where the Sun's effective temperature is $T_\odot \approx 5800\,\mathrm{K}$. It was left as an exercise in Chapter 5 (Problem 5.6) to show that the peak of the Planck function I_λ occurs at a wavelength

$$\lambda_{\mathrm{p}} \approx 0.20 \frac{hc}{kT} \approx \frac{2.9 \times 10^7\,\text{Å K}}{T} \approx \frac{2900\,\mu\mathrm{m\,K}}{T}, \tag{8.4}$$

a relation that is known as **Wien's law**, after the physicist who discovered it empirically. For $T_\odot \approx 5800\,\mathrm{K}$, $\lambda_{\mathrm{p}} \approx 5000\,\text{Å}$, so the solar spectrum, expressed in terms of wavelength, peaks in the visible part of the spectrum.

The flux of energy received by a planet at a distance r from the Sun is

$$F(r) = \frac{L_\odot}{4\pi r^2}. \tag{8.5}$$

The energy that the planet *absorbs* per second is the flux F times the cross-section of the planet (πR^2) times the fraction of the incident light that is absorbed rather than reflected. Since the albedo A of the planet is the fraction of the light energy that is reflected, the fraction that is absorbed is $1 - A$. Thus, the rate at which the planet absorbs energy is

$$W_{\mathrm{p}} = \left(\frac{L_\odot}{4\pi r^2}\right)(\pi R^2)(1 - A). \tag{8.6}$$

The energy absorbed by the planet will raise its surface temperature to some value T_{p}. If we approximate the planet as a blackbody, it will radiate energy at a rate

$$L_{\mathrm{p}} = 4\pi R^2 \sigma_{\mathrm{SB}} T_{\mathrm{p}}^4, \tag{8.7}$$

assuming a uniform surface temperature T_{p} for the planet.

[4] The standard Bad Astronomical Joke is that Saturn would float in a large enough bathtub . . . but that it would leave a ring behind.

[5] The word "albedo" comes from the Latin word *albus*, meaning "white." A white object has a high albedo since it reflects most of the light that strikes it; a black object has a low albedo.

When the planet is in equilibrium, the rate at which energy is emitted by the planet, L_p, is equal to the rate at which energy is absorbed by the planet, W_p. In equilibrium, then,

$$4\pi R^2 \sigma_\mathrm{SB} T_\mathrm{p}^4 = \frac{L_\odot}{4\pi r^2} \pi R^2 (1 - A). \tag{8.8}$$

Solving for the equilibrium surface temperature T_p, we find

$$T_\mathrm{p} = \left(\frac{R_\odot}{r}\right)^{1/2} \left(\frac{1 - A}{4}\right)^{1/4} T_\odot. \tag{8.9}$$

Inserting numerical values for $R_\odot$ and $T_\odot$, and expressing the distance from the Sun in astronomical units, we find

$$T_\mathrm{p} \approx 279\ \mathrm{K}(1 - A)^{1/4} \left(\frac{r}{1\ \mathrm{AU}}\right)^{-1/2}. \tag{8.10}$$

If the object being heated by the Sun is a blackbody, then $A = 0$ and

$$T_\mathrm{bb} \approx 279\ \mathrm{K} \left(\frac{r}{1\ \mathrm{AU}}\right)^{-1/2}, \tag{8.11}$$

which is the **equilibrium blackbody temperature** for a spherical blackbody of uniform temperature at a distance r from the Sun.

The assumption of uniform surface temperature T_p is a good approximation for a planet that is rotating rapidly, like a chicken on a spit, or has efficient atmospheric circulation. However, we should also consider the case of a planet that is rotating slowly and is not a good conductor of heat. In that case, the energy that is absorbed by a small patch of area Σ must be re-radiated by the same patch. The equilibrium condition then becomes

$$\Sigma \sigma_\mathrm{SB} T_\mathrm{p}^4 = \frac{L_\odot}{4\pi r^2} \Sigma (1 - A), \tag{8.12}$$

where T_p is now the temperature of a small patch on the planet's surface for which the Sun is at the zenith. Solving for T_p, the equilibrium temperature of this patch, we find

$$T_\mathrm{p} = \left(\frac{R_\odot}{r}\right)^{1/2} (1 - A)^{1/4} T_\odot, \tag{8.13}$$

or

$$T_\mathrm{p} \approx 395\ \mathrm{K}(1 - A)^{1/4} \left(\frac{r}{1\ \mathrm{AU}}\right)^{-1/2}. \tag{8.14}$$

If the slowly rotating body is a blackbody, with $A = 0$, then the resulting temperature is the **subsolar blackbody temperature**

$$T_\mathrm{ss} \approx 395\ \mathrm{K} \left(\frac{r}{1\ \mathrm{AU}}\right)^{-1/2}, \tag{8.15}$$

TABLE 8.2 Planetary Albedos

Rocky Surfaces	Mercury	0.06
	Moon	0.07
	Mars	0.16
Complex Mix	Earth	0.40
Gases and Ices	Saturn	0.50
	Jupiter	0.51
	Neptune	0.62
	Uranus	0.66
	Venus	0.76

TABLE 8.3 Computed vs. Observed Temperatures

Planet	Albedo	r (AU)	Uniform T_p $T_{bb}(1-A)^{1/4}$	Slow Rotator $T_{ss}(1-A)^{1/4}$	T_{obs}
Venus	0.76	0.72	230 K	325 K	740 K
Earth	~ 0.4	1	246 K	347 K	290 K
Neptune	0.62	30.1	40 K	57 K	59 K

which represents the highest temperature a body can reach at a distance r from the Sun, if sunlight is the only source of heat.

Real objects within the solar system are neither perfectly white ($A = 1$) nor perfectly black ($A = 0$). Strictly speaking, the albedo is a function of wavelength; however, within the solar system, it's a reasonable approximation to use the albedo at visible wavelengths, since that's where most of the Sun's radiated energy is located. Table 8.2 summarizes the approximate albedos of several solar system objects, listed from darkest to brightest. The Earth, thanks to its changing cloud cover, has a highly variable albedo, ranging from a low of $A \approx 0.3$ to a high of $A \approx 0.5$.

In Table 8.3, we give the computed surface temperatures, T_p, for a sample of three planets; we give both the case of uniform temperature and that of slow rotation. For purposes of comparison, we also give the average observed temperature, T_{obs}, for each of these planets. Naïvely, we would expect all these planets to approximate the "uniform temperature" case: the Earth and Neptune are rapid rotators, and although Venus is a slow rotator, it has a dense atmosphere with global mixing. We see, however, that all the planets are warmer than expected. Neptune is warmer than expected because, like all the Jovian planets, it started its existence much hotter than it is today and has not had time to radiate away all its internal heat. Both the Earth and Venus are warmer than expected because of the **greenhouse effect**.

The greenhouse effect occurs when a planet's atmosphere is transparent at visible wavelengths but opaque at infrared wavelengths. To see why that influences temperature at the planet's surface, consider the energy radiated by the warm surface. For the Earth, with $T_p \approx 290$ K, the Planck spectrum peaks at $\lambda_p \approx 10\ \mu$m. For Venus, with $T_p \approx 740$ K, the peak is at $\lambda_p \approx 4\ \mu$m. These wavelengths are in the infrared range of the spectrum. If the atmosphere is opaque at infrared wavelengths, the radiated light will not freely escape from the atmosphere. Instead, it will be absorbed and re-radiated, until the light energy makes its way to the upper atmosphere, where the optical depth at infrared wavelengths drops to $\tau \sim 1$. Then, at last, the light can escape. Thus, the photosphere of the Earth and Venus—the layer from which their radiated infrared light escapes—is higher than, and cooler than, the solid surface of the planet. Naturally occurring greenhouse gases include water vapor (H_2O), carbon dioxide (CO_2), and methane (CH_4). The greenhouse effect is particularly strong on Venus because its dense atmosphere is roughly 95% carbon dioxide.

The flux of light we observe from a planet depends on wavelength. Consider, as an example, the planet Neptune, with a measured temperature of $T = 59$ K. Using Wien's law (equation 8.4), its spectrum should peak at a wavelength $\lambda_p = 50\ \mu$m, in the far infrared. The light we see from Neptune at visible wavelengths is reflected sunlight, modified by absorption in the atmosphere of Neptune. (Uranus and Neptune both appear bluer than the Sun at visible wavelengths, as shown in Color Figure 5, because their atmospheres contain methane, which strongly absorbs red wavelengths of light.) The spectrum of a planet can thus be roughly approximated as the sum of two blackbody spectra. One spectrum is that of reflected sunlight, containing both the Fraunhofer absorption lines of sunlight and any absorption lines contributed by the planet itself. The other spectrum is that of the planet's thermal emission. The reflected sunlight has an integrated energy proportional to the albedo A; the thermal emission has an energy proportional to $1 - A$ (plus any internal source of energy). For instance, Figure 8.2 shows the spectrum of Mars at UV, visible, and IR wavelengths. Iron oxide in the soil of Mars strongly absorbs blue light and thus makes Mars appear reddish at visible wavelengths (in other words, Mars looks red because it's rusty). The thermal emission peaks at $\lambda_p \sim 13\ \mu$m, corresponding to a temperature $T \sim 225\ \mathrm{K} \sim -48°\mathrm{C}$, about average for Mars.

An underlying reason for many of the observed differences between terrestrial and Jovian planets is that the Jovian planets have a chemical composition that closely resembles that of the Sun; that is, they are mostly hydrogen and helium. The terrestrial planets are deficient in hydrogen and helium because they are too hot to retain such light elements.

Let's consider the conditions under which planets can retain different atmospheric gases. The hotter a planet's atmosphere, the higher the random thermal speeds of the individual gas particles. In a dense atmosphere, though, any individual particle will travel only a short distance before colliding with another particle and changing its velocity. In the Earth's atmosphere at sea level, for instance, we computed the mean free path of a typical molecule to be only ~ 400 Å (see equation 5.67). However, at higher levels in the atmosphere, the density of molecules decreases, and the mean free path therefore increases. At some height in the atmosphere, the mean free path increases to the point

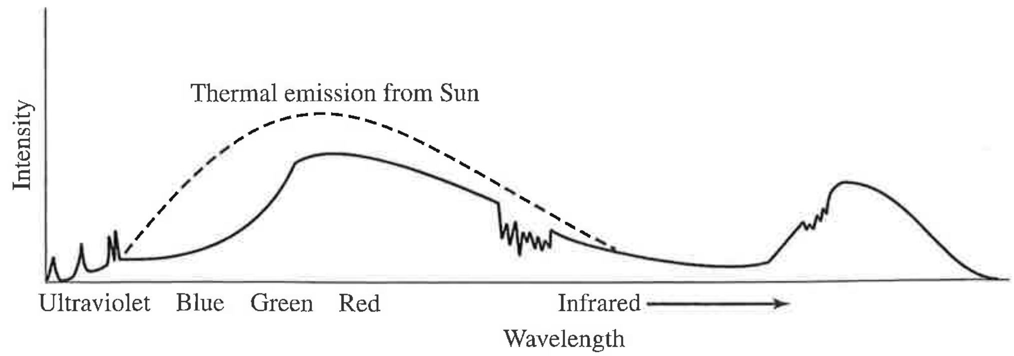

FIGURE 8.2 The spectrum of Mars at ultraviolet, visible, and infrared wavelengths. The peak on the left is reflected sunlight; the peak on the right is thermal emission at $T \sim 225$ K, the temperature of the surface of Mars.

where a gas particle moving upward faster than the escape speed is able to escape from the atmosphere before colliding with another particle. This height is called the **exobase**, and the layer of the atmosphere above the exobase is called the **exosphere**.

To compute where the exobase lies, let's assume that gas particles in the atmosphere have an average cross-section σ for collisions. (For small molecules, like N_2, O_2, and CO_2, we expect $\sigma \sim 10^{-18}\ \mathrm{m}^2$.) The number density of gas particles in the atmosphere is $n(z)$, where z is the height above the planet's surface, or above some convenient reference level, if the planet has no solid or liquid surface. The height z_{ex} of the exobase is given by the relation

$$\int_{z_{\mathrm{ex}}}^{\infty} \sigma n(z) dz = \sigma N(z_{\mathrm{ex}}) = 1. \tag{8.16}$$

Here $N(z_{\mathrm{ex}})$ is the column density of gas particles above the exobase. Thus, for typical molecule sizes, the exobase is located where the column density falls to $N \sim 1/\sigma \sim 10^{18}\ \mathrm{m}^{-2}$. In the Earth's atmosphere, the exobase is at a height $z_{\mathrm{ex}} \sim 500$ km above sea level.

At the exobase, the rms speed of gas particles with mass m will be

$$v_{\mathrm{rms}} = \left(\frac{3kT_{\mathrm{ex}}}{m} \right)^{1/2}, \tag{8.17}$$

where T_{ex} is the temperature at the exobase. We can compare this to the escape speed at the distance R_{ex} of the exobase from the planet's center:

$$v_{\mathrm{esc}} = \left(\frac{2GM}{R_{\mathrm{ex}}} \right)^{1/2}, \tag{8.18}$$

where M is the planet's mass. If the typical particle speed v_{rms} is comparable to the escape speed, the atmosphere quickly escapes. Even if $v_{\mathrm{rms}} < v_{\mathrm{esc}}$, there will be some fraction of particles in the high-speed exponential tail of the Maxwell–Boltzmann distribution (equation 5.40) that will exceed the escape speed. A useful rule of thumb[6] is that for a planet to retain a particular gas in its atmosphere for the age of the solar system, the particles of the gas must have $v_{\rm rms} \lesssim v_{\rm esc}/6$. Squaring both sides, we can write this condition as

$$\begin{aligned} v_{\rm rms}^2 &\lesssim \frac{1}{36} v_{\rm esc}^2 \\ \frac{3kT_{\rm ex}}{m} &\lesssim \frac{GM}{18R_{\rm ex}}. \end{aligned} \tag{8.19}$$

In other words, retaining gas particles of mass m requires a temperature

$$T_{\rm ex} \lesssim \frac{GMm}{54kR_{\rm ex}}. \tag{8.20}$$

It is useful to scale this result to the Earth's mass, $M_\odot = 5.97 \times 10^{24}$ kg, and the Earth's radius, $R_\odot = 6.37 \times 10^6$ m, while expressing the particle mass in terms of the molecular mass $\mu = m/m_p$. With this scaling,

$$T_{\rm ex} \lesssim \frac{m_p}{54k} \mu g R_{\rm ex} \tag{8.21}$$

$$\lesssim 140\ {\rm K} \left(\frac{M}{M_\oplus}\right) \left(\frac{R_{\rm ex}}{R_\oplus}\right)^{-1} \mu. \tag{8.22}$$

For a planet with exobase temperature $T_{\rm ex}$, we can write the condition for retaining a gas in the form

$$\mu \gtrsim \frac{54kT_{\rm ex}}{gR_{\rm ex}m_p}. \tag{8.23}$$

The Earth's exobase has a higher temperature than the atmosphere at sea level, because the upper atmosphere is heated by interactions with high-energy solar wind particles; during daytime, the exobase temperature is $T_{{\rm ex},\oplus} \approx 1000$ K. Scaling to the temperature of the Earth's exobase, we find that the condition for retaining a gas is

$$\mu \gtrsim 7.1 \left(\frac{T_{\rm ex}}{1000\ {\rm K}}\right) \left(\frac{M}{M_\oplus}\right)^{-1} \left(\frac{R_{\rm ex}}{R_\oplus}\right). \tag{8.24}$$

Thus, the Earth, where $R_{\rm ex} = 1.08 R_\oplus$, can retain gases with $\mu \gtrsim 8$; H_2 ($\mu = 2$) and He ($\mu = 4$) can escape, but N_2 ($\mu = 28$), O_2 ($\mu = 32$), and CO_2 ($\mu = 44$) are retained. However, let's now look at the planet Mercury, which has so few molecules in its atmosphere that its exobase is located at its solid surface. The mass and radius of Mercury are $M = 0.055 M_\oplus$ and radius $R = 0.38 R_\oplus$, and the daytime temperature is $T = 700$ K.

[6] The phrase "rule of thumb" apparently derives from brewers who estimated the temperature of their mash by sticking in their thumb. Hence, it means an approximate (as opposed to precise) guideline.

Thus, during daytime on Mercury, the retention condition for gas particles is

$$\mu_{\rm Merc} \gtrsim \frac{7.1 \cdot 0.7 \cdot 0.38}{0.055} \approx 34. \tag{8.25}$$

Thus, Mercury would be able to retain only highly massive molecules in its atmosphere.

8.3 ▪ FORMATION OF THE SOLAR SYSTEM

In this section, we give a broad overview of current ideas about how the solar system formed. We return to this topic in Chapter 12 and fill in some of the details.

The modern view of planetary formation is that it is a natural consequence of star formation. Put simply, formation of planets is a way for a collapsing gas cloud to rid itself of angular momentum so that it can shrink to the size of a star. Star formation occurs when a large gas cloud collapses under its own gravity; since this can happen only if the self-gravity exceeds the support provided by gas pressure, star formation occurs only in dense, cold ($T \lesssim 10$ K) regions of interstellar gas. If the collapsing gas cloud has a net angular momentum, it will collapse until it forms a rotationally supported disk. The dense central region of the disk ultimately becomes a star, while smaller condensations that occur within the disk ultimately grow into planets. Such rotationally supported, dusty, gaseous disks, called **protoplanetary disks**, are seen around young stars in our galaxy. For instance, Figure 8.3 shows a protoplanetary disk in the Orion Nebula. The edge-on protoplanetary disk resembles an edge-on hamburger: the dark "patty" represents a thin, dust-rich disk, and the bright "buns" consist of light from the central protostar scattered by a more diffuse distribution of dust.

The first step in planet formation is **condensation**, the formation of solid particles within the gaseous disk as it cools. Different materials condense into solids at different temperatures; metals condense at high temperatures, rocky materials at intermediate temperatures, and ices at low temperatures. Materials that condense into a solid (or liquid)

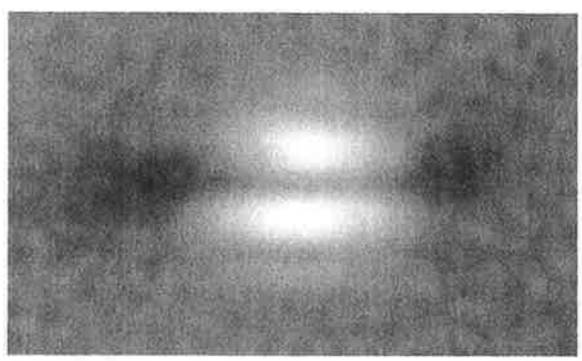

FIGURE 8.3 An image, taken at $\lambda \approx 2\ \mu$m, of the edge-on protoplanetary disk Orion 114-426. The size of the region shown is 1300 AU by 800 AU.

TABLE 8.4 Simplified Condensation Sequence

T (K)	Condensate	Planet
1500	Metal oxides	Mercury
1300	Fe, Ni	
1200	Silicates	
700	FeS (iron sulfide)	Venus
200	H_2O	Earth, Mars
150	NH_3	Jovian planets
120	CH_4	Pluto, Eris
65	Ar, Ne	

only at very low temperatures are called **volatile** materials.[7] The gaseous disk from which the planets ultimately form is hottest at the center, where the protosun is located, and becomes cooler with increasing distance. In the central parts of the protoplanetary disk, only the most refractory materials can condense to form solid particles; a **refractory** material is one that condenses into a solid (or liquid) at relatively high temperatures. In the outer parts of the disk, by contrast, the temperature is sufficiently low that even volatile materials can condense. Table 8.4 gives approximate condensation temperatures for different substances; the right-hand column lists the planets that will ultimately form at the given temperature.

Condensation, more specifically, is the process by which solids grow molecule by molecule, as individual molecules (or atoms) adhere to the solid body. As an example, snowflakes grow by condensation. Eventually, however, the major mode of growth switches over from condensation to **accretion**, in which solid condensates come together and are held together by weak electrical forces. For instance, individual snowflakes can accrete together to form a snowball.[8] In the early solar system, the collisions between individual condensates, or "snowflakes," are gentle, since they are on similar orbits. Objects grow by accretion until they are roughly 1 km across; these intermediate-sized bodies are known as **planetesimals**.[9] Near the Sun, planetesimals are made of the least volatile materials: metal oxides and metals. Farther from the Sun, where the temperatures are cooler, the planetesimals are made of a mix of materials with different degrees of volatility.

Eventually, planetesimals are drawn toward each other by gravity and merge to form larger bodies; this process is known as **coalescence**. Since only the least volatile materials can condense close to the Sun, the planets built up by coalescence within 1.5 AU of the Sun are rich in low-volatility elements, even though such elements are relatively rare

[7] The adjective "volatile" comes from the Latin word *volare*, meaning "to fly." A volatile solid is one in whose atoms fly apart to form a gas at a low temperature.

[8] In other subfields of astronomy, "accretion" can refer to a flow of gas onto a compact object such as a black hole. Such dual definitions are annoying, but it's hard to change entrenched usage.

[9] "Planetesimal" = "planet" + "infinitesimal." It's a portmanteau word, as Lewis Carroll would say.

TABLE 8.5 Uncompressed Densities of Terrestrial Planets

Planet	Density (kg m^{-3})
Mercury	5400
Venus	4200
Earth	4200
Mars	3300

in the universe at large. These metal-rich and silicate-rich bodies close to the Sun are the terrestrial planets. Since materials with low volatility tend to be dense, the terrestrial planets all have high densities. Table 8.5 lists the uncompressed densities of the terrestrial planets; in other words, this is the density the planet would have if it were not squeezed to a smaller volume and higher density by its own gravity.[10] Note that the uncompressed density of the terrestrial planets decreases with increasing distance from the Sun. (It is interesting, however, that the uncompressed density of the Moon is only 3350 kg m^{-3}, much less than the Earth's uncompressed density, despite the fact that they are at the same distance from the Sun. The unexpectedly low density of the Moon is a consequence of its unique formation history, discussed in Section 9.5.)

In the outer part of the protoplanetary disk, containing planetesimals of both high-volatility and low-volatility materials, larger bodies are able to form. Once these "protoplanets" reach a mass of $M \sim 15M_{\oplus}$, they are massive enough, given the lower temperatures in the outer disk, to retain hydrogen and helium. Since hydrogen and helium are the most abundant elements, being able to hang onto them permits the Jovian planets to grow to much larger masses than the terrestrial planets.

Protoplanets grow gradually by the coalescence of planetesimals until they reach the masses of the planets as we know them today. During and immediately following this early phase of coalescence, the interior and surface structure of terrestrial planets are changed by several processes, among them chemical differentiation, cratering, volcanic flooding, and slower processes of surface evolution.

Chemical differentiation is the process by which dense elements sink to the center of terrestrial planets while the lower-density elements float up to the surface. Since planets form by a jumbled coalescence of planetesimals of differing chemical composition, we might expect that planets would show the same jumble at their center as at their surface. However, study of the interiors of terrestrial planets shows that they are chemically differentiated; that is, their chemical composition varies with distance from the center. The cores of terrestrial planets are rich in dense elements like iron and nickel, with uncompressed densities of $8000 \rightarrow 9000$ kg m^{-3}. However, their outer layers are primarily made of silicate rocks, with typical uncompressed densities of ~ 3000 kg m^{-3}.

[10] The Earth is the most massive terrestrial planet and has the greatest gravitational compression. The measured mean density of the Earth is $\rho_{\oplus} = 5500$ kg m^{-3}, significantly higher than its uncompressed density of 4200 kg m^{-3}.

FIGURE 8.4 Mercury as imaged by the *MESSENGER* spacecraft.

Differentiation occurs naturally in a fluid body; the denser material sinks while the lower-density material rises to the top. The fact that terrestrial planets are differentiated is an indication that they were once hot enough to be fully molten. In the Earth, the radioactive decay of elements such as uranium, thorium, and potassium-40 releases enough heat to melt iron. The iron, as it sinks toward the Earth's center, converts gravitational potential energy into heat, leading to an "iron catastrophe," that is, a runaway heating process that doesn't cease until the interior is highly differentiated.

Cratering begins as the newly formed planets sweep up the remaining planetesimals. The process of cratering modifies the outer crust of the terrestrial planets. For instance, cratering has given Mercury (Figure 8.4) its characteristically pock-marked appearance. Cratering was an important process until about 3.3 billion years ago; at that time, the early "period of heavy bombardment" ended because the planetestimals and other small objects had been largely cleared out of the inner solar system.

Volcanic flooding of planetary surfaces occurred at the same time as cratering. Fracturing a planet's crust by large impacts leads to flooding of the surface by lava welling up from below. This obliterates the older surface features; craters are paved over by lava, just as old potholes in a road are paved over by a new layer of asphalt. Surface features on a terrestrial planet can also be worn away by slower processes, such as erosion by wind and water. At the same time, surface features can be built up by other slow processes, such as plate tectonics and volcanism. On a geologically active body such as the Earth, surface features are in a constant state of flux.

The surface of a terrestrial planet can change with time; so can its atmosphere. The primeval atmosphere of a planet is whatever is left behind by the formation process. The atmospheres of terrestrial planets evolve for several reasons:

1. Gases can escape from planets if the individual gas particles are moving sufficiently rapidly, as discussed in Section 8.2. Planets with cooler temperatures and higher surface gravity will retain more of their primeval atmosphere.

2. Outgassing from the planet's interior releases gases that were trapped in the interior during the formation process. These planetary belches occur as part of the differentiation process and continue through ongoing volcanic activity.

3. Chemical interactions between the atmosphere and the surface can alter the atmosphere's composition. For instance, the interaction of gaseous CO_2 with liquid water removed most of the CO_2 from the Earth's atmosphere and dissolved it in the Earth's oceans.

The mechanism by which planetary magnetic fields are generated is poorly understood. In general terms, it is thought to result from dynamo action in the planetary core; hot, partially ionized matter wells up by convection in planetary interiors and is deflected by the Coriolis effect. Our expectation, then, is that larger magnetic fields will be found to be associated with larger bodies (which will have larger liquid cores) and faster rotators (which will have a larger Coriolis effect). In fact, as we will see, the planet Jupiter, which is both the largest and the fastest rotating planet in the solar system, has the strongest magnetic field.

PROBLEMS

8.1 What is the mean mass density $\overline{\rho}$ of Saturn's largest satellite, Titan? What does this suggest about the composition of Titan?

8.2 Radioactive decay of elements in the Earth's interior results in a mean heat flux through the Earth's surface of 5×10^{-2} W m^{-2}. What is this flux expressed as a fraction of the energy flux due to thermal re-radiation of absorbed solar energy? If radioactive decay were the *only* heat source for the Earth, what would the Earth's surface temperature be?

8.3 Mercury has an orbit with semimajor axis $a = 0.387$ AU and eccentricity $e = 0.206$. Mercury is a slowly rotating planet with no atmosphere. What is the temperature of the subsolar point on Mercury at aphelion? What is the temperature of the subsolar point on Mercury at perihelion? (The "subsolar point" is the location on the planet's surface where the Sun is at the zenith.)

8.4 Pure, solid water ice has an albedo $A \approx 0.35$. What is the minimum distance from the Sun at which a rapidly rotating ice cube would remain frozen? Between which two planets does this distance lie?

8.5 Suppose that Uranus were moved to the location of Jupiter; would Uranus then retain its hydrogen-rich atmosphere?

8.6 Because Venus has a very feeble magnetic field, the solar wind collides with its atmosphere, instead of being deflected by magnetic forces. Suppose that if a solar wind particle strikes the atmosphere of Venus, all its kinetic energy will be absorbed.

(a) What is the rate, in watts, at which Venus absorbs energy from the solar wind? Assume that the energy density of the solar wind is

$$\rho v^2/2 = 2 \times 10^{-9}\,\mathrm{J\,m^{-3}},$$

and that the solar wind speed is $v = 400\,\mathrm{km\,s^{-1}}$.

(b) What is the rate, in watts, at which Venus absorbs energy from sunlight? Is the solar wind a significant heat source for Venus?

8.7 Jupiter's moon Callisto is slowly rotating and has a low albedo ($A \approx 0.2$). What is the temperature of Callisto's subsolar point? Would you expect Callisto to retain an atmosphere of N_2? What about an atmosphere of He? (Hint: you may assume that the exobase lies at the surface of Callisto.)