years we will lie in its equatorial plane (Figure 2–4B).

Mercury and Venus have low rotation rates, which may be explained by a **spin–orbit coupling** and **resonance.** The synodic rotation period of Venus (its rotation period with respect to the Earth) is 146 days, which is one-quarter of its synodic orbital period. The sidereal rotation period of Mercury is exactly two-thirds its sidereal orbital period (Table A3–2). Hence, Venus is in a roughly 4:1 synodic resonance with the Earth, and Mercury exhibits a 3:2 sidereal lock with the Sun. Both planets are slightly nonspherical. Solar tidal forces interact with these deformations, slowing each planet's direct rotation until a resonance is achieved. Mercury's highly eccentric orbit resulted in the Sun–Mercury resonance (Figure 2–5): Mercury points alternate longitudes toward the Sun at each perihelion, rotating three times in two sidereal orbital periods.

(B) INTERIORS

The planets naturally divide into two groups (Table A3–3): (1) the small, solid **terrestrial** planets (Mercury, Venus, Earth, Moon, and Mars), with masses no greater than the Earth's, and (2) the large, liquid **Jovian** planets (Jupiter, Saturn, Uranus, and Neptune), which range in mass from $15M_{\oplus}$ to $318M_{\oplus}$. (Pluto, a binary planet, does not fall cleanly into either class.) The split between terrestrial and Jovian reflects fundamental differences in composition between the two groups.

Planetary masses are determined (1) by applying Kepler's third law to satellite orbits and, when there are no natural satellites, (2) by observing a planet's gravitational perturbations on the orbits of other planets, asteroids, comets, and artificial space probes. Planetary radii are deduced (1) by measuring the apparent optical size of the planetary disk, (2) by accurately timing the **occultations** (when the planet passes in front of an object) of stars, the planet's moons, and space probes, and, for the closer planets, (3) by precisely timing radar pulses reflected from various points on the planet's surface.

By dividing a planet's mass M by its total volume $4\pi R^3/3$, we may define its **average density** $\langle\rho\rangle$ (SI units = $\mathrm{kg/m^3}$):

$$\langle\rho\rangle = M/(4\pi R^3/3) \qquad \textbf{(2–1)}$$

The density of pure water is 1000 $\mathrm{kg/m^3}$. The high densities of the terrestrial planets, in the range of 3400 to 5500 $\mathrm{kg/m^3}$, reflect the fact that they are composed of heavy, nonvolatile elements such as iron, silicon, and magnesium. The very low densities of the Jovian planets (Saturn could *float* on water!) imply a composition similar to that of our Sun, with hydrogen and helium dominating.

The internal structure of a planet depends up the distribution of its chemical composition, de sity, temperature, and pressure. In general, t pressure increases closer to the center of a plan and the temperature also rises as a result of t greater pressure. Different materials are stal under different conditions, so that the chemi composition is radially layered—the interiors a **differentiated,** rather than homogeneous. T Earth has a metallic core above which lie light silicates; we expect Venus to exhibit a similar in rior because its mass and composition are simi to the Earth's. The Jovian planets have thick interiors of mostly hydrogen and helium, probably with rocky cores.

Using the observed average density, composition, and oblateness of a planet, we may construct physically consistent models of its interior. Usually a range of models accommodate our data, so that only improved theory and observation can lead us to a unique picture.

(C) SURFACES

Whereas the Jovian planets and Venus show only their upper cloudy atmospheres, the other terrestrial planets and larger moons reveal surface markings. The most important general data on planetary surfaces include color, albedo, and temperature.

A planet's **color** relates to the composition of its surface and atmosphere. The oceans and continents give the Earth a blue color mottled with green, brown, and orange; large areas of cloud or snow cover appear white. The basaltic surface of the Moon looks dark gray with some tan, whereas the deserts of Mars give its characteristic brown-orange color. The surface of Io (a large moon of Jupiter) has a yellowish cast because of outflows from sulfur volcanoes.

The **albedo** of an object is the fraction of the incident sunlight reflected by it. Astronomers usually write this reflectivity as

$$A = \text{amount reflected/amount incident}$$

For planets with little or no atmosphere (Mercury, our Moon, and Mars), the albedo is very low because basaltic rocks are poor reflectors. Icy surfaces, as on most of Saturn's moons, have moderate albedos. The high reflectivity of clouds leads to the high albedos of the Jovian planets and Venus. The Earth's albedo is variable since it depends upon the season and upon snow and cloud cover; it averages about 0.35. Note that albedos are a function of wavelength; a rocky surface has different reflectivities at optical and infrared wavelengths, for example.

An important characteristic of planetary surfaces is **surface temperature.** Let's see how surface temperatures are observed and computed for **blackbody radiators.** Chapter 8 contains a complete discussion about blackbody radiators. Here we will introduce you to a few basics of these hypothetical objects. You can think of blackbodies in two ways: by their absorption properties and by their emissive ones. As the name implies, a blackbody completely absorbs all forms of electromagnetic radiation striking it; none is reflected, and so its albedo is zero. When a blackbody heats up to some temperature, its spectrum of emitted light (Figure 2–6) has a characteristic shape, called a **Planck curve,** with one maximum. A blackbody's emission peaks at the wavelength

$$\lambda_{max} = (0.002898 \text{ m})/T = (2898\ \mu\text{m})/T \qquad \textbf{(2–2)}$$

where T is the absolute temperature (kelvins) of the blackbody; one micrometer (μm) equals 10^{-6} m. This relation is **Wien's law,** which tells us that solar radiation ($T \approx 6000$ K) peaks at 0.5 μm; at normal room temperature ($T \approx 290$ K), the peak is at the longer infrared wavelength $\lambda_{max} = 10\ \mu$m (Figure 2–6). Hence, we can determine a planet's surface temperature by observing its peak of emission.

Radiation of different wavelengths comes from different parts of a planet's surface. Therefore, we can probe the temperature and composition of varying levels of a planet's surface by observing the amount of radiation at different wavelengths. Because the Moon and Mercury have no atmosphere, infrared comes from the top few millimeters of the visible surfaces, and longer-wavelength radiation originates several centimeters beneath the surface. We can use these thermal emissions to probe subsurface properties and conditions.

A planet that radiates more energy per second than it receives from the Sun must have an *internal heat source.* Delicate measurements at the Earth's surface reveal heat flowing from the hot interior; the same occurs with the Moon. The other planets known (from infrared observations) to produce excess heat are Jupiter, Saturn, and Neptune.

Now we will show how planetary blackbody temperatures are calculated (Table A3–3). The basic idea is to find the temperature at which a small blackbody (a planet) must radiate to balance the energy input from the Sun. We use **Stefan's law** of blackbody radiation (Chapter 8). This law relates the **energy flux** E (energy radiated per unit area per unit time, or W/m^2) to the temperature T of a blackbody:

$$E = \sigma T^4 \text{ W/m}^2 \qquad \textbf{(2–3)}$$

where the constant is $\sigma = 5.67 \times 10^{-8}$ W $\cdot$ m^{-2} $\cdot$ K^{-4}. Because the area of a spherical surface of radius R is $4\pi R^2$, our Sun radiates energy like a blackbody at the rate ($T_\odot = 5800$ K)

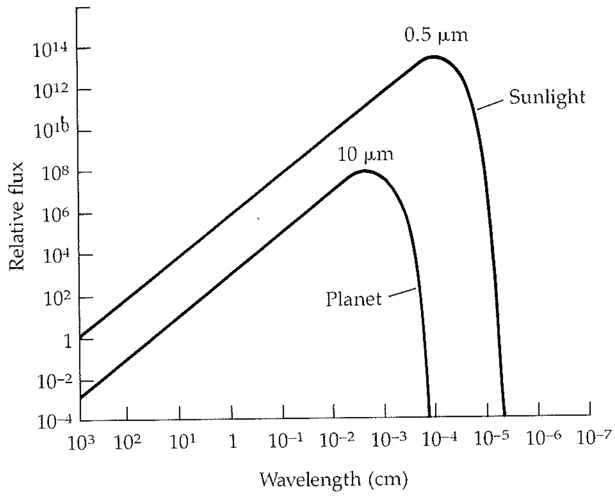

FIGURE 2–6 Continuous emission of reflected sunlight and a planet. Radiating like a blackbody, the Sun's emission peaks near 0.5 μm. For a planet at 290 K, the peak occurs at 10 μm.

$$4\pi R_{\odot}^2 E_{\odot} = 4\pi R_{\odot}^2 \sigma T_{\odot}^4 = 3.8 \times 10^{26}\ \mathrm{W}$$

which is the Sun's luminosity ($L_{\odot}$)—its total radiative power. This same energy flows through a sphere of area $4\pi r_p{}^2$ at the Sun–planet distance r_p, so that the energy flux there is

$$E_p = 4\pi R_{\odot}^2 E_{\odot} / 4\pi r_p{}^2 = (R_{\odot}/r_p)^2 E_{\odot}$$

If 1 m^2 of blackbody material intercepts this energy head-on with the Sun overhead, it will be raised to the **subsolar temperature** (in kelvins):

$$T_{ss} = (R_{\odot}/r_p)^{1/2} T_{\odot} \approx 394(r_p)^{-1/2} \qquad \textbf{(2–4)}$$

since $E_p = \sigma T_{ss}{}^4$. In the last equality, we have inserted the appropriate solar values and expressed r_p in AU. The subsolar temperatures are essentially the equilibrium **noontime** temperatures.

Although subsolar temperatures apply to the local noon on the surfaces of very slowly rotating planets (Mercury, the Moon, and Pluto), they are not appropriate for planets with atmospheres or for planets in rapid rotation. For these, assume that the effective absorbing area is the cross section $\pi R_p{}^2$ and that the effective radiating area is the total surface area $4\pi R_p{}^2$, where R_p is the radius of the planet. The albedo A is the fraction of incident solar radiation that is reflected, so that only the fraction $1 - A$ is absorbed. Therefore, energy absorbed per second is

$$(1 - A)\pi R_p{}^2 E_p = (1 - A)\pi R_p{}^2 (R_{\odot}/r_p)^2 \sigma T_{\odot}^4$$

while the planet radiates the power $4\pi R_p{}^2 \sigma T_p{}^4$. We again use the concept of equilibrium: that the rate of absorption of energy equals the rate at which it is radiated away fixes an equilibrium temperature. (Otherwise the temperature would either rise or fall.) Equating these and solving for the temperature, we have (in kelvins)

$$T_p = (1 - A)^{1/4} (R_{\odot}/2r_p)^{1/2} T_{\odot} \approx 279(1 - A)^{1/4}(r_p)^{-1/2} \qquad \textbf{(2–5a)}$$

with r_p expressed in AU. The equilibrium blackbody temperatures (Table A3–3) follow when $A = 0$. For the subsolar temperature, we get

$$T_{ss} \approx 394(1 - A)^{1/4}(r_p)^{-1/2} \qquad \textbf{(2–5b)}$$

By including the albedo effect, we may more closely approximate the observed planetary temperatures, but remember that these estimates neglect important complications, such as the circulation of planetary atmospheres, their heat retention, internal heat sources, and the variation of A with wavelength.

(D) ATMOSPHERES

Of the terrestrial planets, Mercury and the Moon have essentially no atmosphere, Venus and Mars possess a carbon dioxide (CO_2) atmosphere, and the Earth's atmosphere is primarily molecular nitrogen (N_2) and oxygen (O_2). The principal constituents of the atmospheres of the Jovian planets are molecular hydrogen (H_2) and helium (He). In general, planetary atmospheres are densest near the planet's surface and thin rapidly with increasing altitude. The composition of an atmosphere may be stratified, with the densest gases residing closest to the surface of the planet, but turbulent mixing and winds can lead to regions of homogeneous composition. Far from the planet's surface, incoming solar ultraviolet rays and X-rays usually ionize atmospheric atoms or dissociate molecules to form the layered **ionosphere** (Chapter 4).

To gain some understanding of planetary atmospheres, consider a simple model for their retention. To a first approximation, an atmosphere behaves like a **perfect gas,** that is, as particles that interact only through elastic collisions. Such a gas obeys a special relationship between pressure,

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is the number density
absolute temperature
nstant, equal to 1.38 ×
continuous collisions,

the particles of the gas achieve, at a given temperature, an equilibrium distribution of velocities so that

$$\ldots{}^2/kT)v^2\,dv$$

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