Electromagnetic Radiation

2.1. Introduction

With the exception of objects at absolute zero, all objects emit electromagnetic radiation. Objects also reflect radiation that has been emitted by other objects. By recording emitted or reflected radiation, and applying a knowledge of its behavior as it passes through the earth's atmosphere and interacts with objects, remote sensing analysts develop a knowledge of the character of features such as vegetation, structures, soils, rock, or water bodies on the earth's surface. Interpretation of remote sensing imagery depends on a sound understanding of electromagnetic radiation and its interaction with surfaces and the atmosphere. The discussion of electromagnetic radiation in this chapter builds a foundation to permit development in subsequent chapters of the many other important topics within the field of remote sensing.

The most familiar form of electromagnetic radiation is visible light, which forms only a small (but very important) portion of the full electromagnetic spectrum. The large segments of this spectrum that lie outside the range of human vision require our special attention because they may behave in ways that are quite foreign to our everyday experience with visible radiation.

2.2. The Electromagnetic Spectrum

Electromagnetic energy is generated by several mechanisms, including changes in the energy levels of electrons, acceleration of electrical charges, decay of radioactive substances, and the thermal motion of atoms and molecules. Nuclear reactions within the sun produce a full spectrum of electromagnetic radiation, which is transmitted through space without experiencing major changes. As this radiation approaches the earth, it passes through the atmosphere before reaching the earth's surface. Some is reflected upward from the earth's surface; it is this radiation that forms the basis for photographs and similar images. Other solar radiation is absorbed at the surface of the earth, and is then reradiated as thermal energy. This thermal energy can also be used to form remotely sensed images, although they differ greatly from the aerial photographs formed from reflected energy. Finally, man-made radiation, such as that generated by imaging radars, is also used for remote sensing.

Electromagnetic radiation consists of an electrical field (E) that varies in magnitude in a direction perpendicular to the direction of propagation (Figure 2.1). In addition, a magnetic field (H) oriented at right angles to the electrical field is propagated in phase with the electrical field.

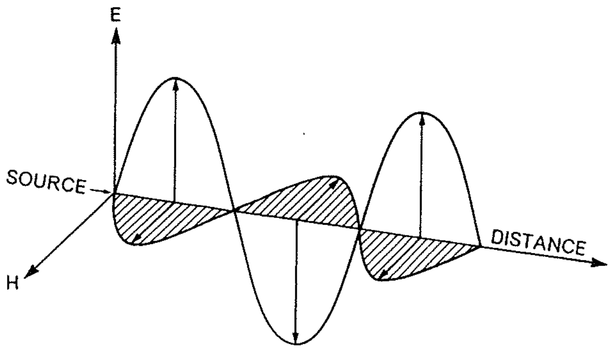


FIGURE 2.1. Electric (E) and magnetic (H) components of electromagnetic radiation. The electric and magnetic components are oriented at right angles to one another, and vary along an axis perpendicular to the axis of propagation.

Electromagnetic energy displays three properties (Figure 2.2):

1. *Wavelength* is the distance from one wave crest to the next. Wavelength can be measured in everyday units of length, although very short wavelengths have such small distances between wave crests that extremely short (and therefore less familiar) measurement units are required (Table 2.1).
2. *Frequency* is measured as the number of crests passing a fixed point in a given period of time. Frequency is often measured in *hertz,* units each equivalent to one cycle per second (Table 2.2), and multiples of the hertz.
3. *Amplitude* is equivalent to the height of each peak (see Figure 2.2). Amplitude is often measured as energy levels (formally known as *spectral irradiance*), expressed as watts per square meter per micrometer (i.e., as energy level per wavelength interval).

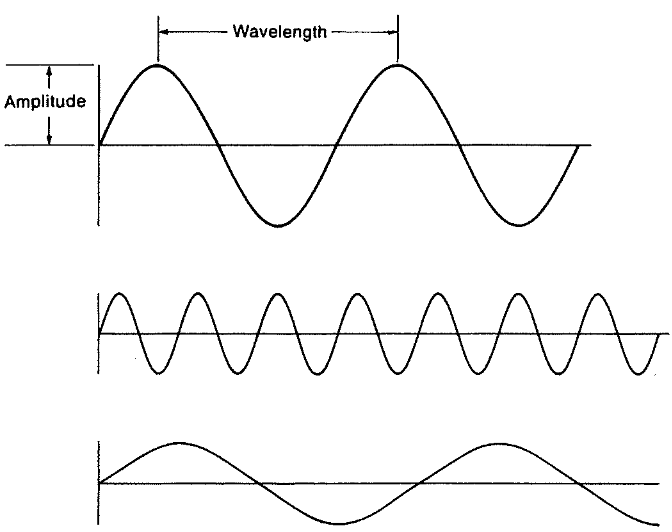


FIGURE 2.2. Amplitude, frequency, and wavelength. The center diagram represents high frequency, short wavelength; the bottom diagram shows low frequency, long wavelength.

TABLE 2.1. Units of Length Used in Remote Sensing

Unit	Distance	
Kilometer (km)	1,000 m	
Meter (m)	1.0 m	
Centimeter (cm)	0.01 m	$= 10^{-2}$ m
Millimeter (mm)	0.001 m	$= 10^{-3}$ m
Micrometer (μm)[a]	0.000001 m	$= 10^{-6}$ m
Nanometer (nm)		10^{-9} m
Ångstrom unit (Å)		10^{-10} m

[a]Formerly called the "micron" (μ); the term "micrometer" is now used by agreement of the General Conference on Weights and Measures.

The speed of electromagnetic energy (c) is constant at 299,893 kilometers (km) per second. Frequency (ν) and wavelength (λ) are related:

$$c = \lambda \nu \tag{Eq. 2.1}$$

Therefore, characteristics of electromagnetic energy can be specified using either frequency or wavelength. Varied disciplines, and varied applications, follow different conventions for describing electromagnetic radiation, using either wavelength (measured in Angstrom units [Å], microns, micrometers, nanometers, millimeters, etc., as appropriate), or frequency (using hertz, kilohertz, megahertz, etc., as appropriate). Although there is no authoritative standard, a common practice in the field of remote sensing is to define regions of the spectrum on the basis of wavelength, often using micrometers (each equal to one one-millionth of a meter, symbolized as μm), millimeters (mm), and meters (m) as units of length. Departures from this practice are common; for example, electrical engineers who work with microwave radiation traditionally use frequency to designate subdivisions of the spectrum. This text will usually employ wavelength designations. The student should, however, be prepared to encounter different usages in scientific journals and in references.

2.3. Major Divisions of the Electromagnetic Spectrum

Major divisions of the electromagnetic spectrum (Table 2.3) are, in essence, arbitrarily defined. In a full spectrum of solar energy there are no sharp breaks at the divisions indicated graphically in Figure 2.3. Subdivisions are established for convenience and by traditions within different disciplines, so do not be surprised to find different definitions in other sources or in references pertaining to other disciplines.

TABLE 2.2. Frequencies Used in Remote Sensing

Unit	Frequency (cycles per second)
Hertz (Hz)	1
Kilohertz (kHz)	10^3 (= 1,000)
Megahertz (MHz)	10^6 (= 1,000,000)
Gigahertz (GHz)	10^9 (= 1,000,000,000)

TABLE 2.3. Principal Divisions of the Electromagnetic Spectrum

Division	Limits
Gamma rays	<0.03 nm
X-rays	0.03–300 nm
Ultraviolet radiation	0.30–0.38 μm
Visible light	0.38–0.72 μm
Infrared radiation	
Near infrared	0.72–1.30 μm
Mid-infrared	1.30–3.00 μm
Far infrared	7.0–1,000 μm (1 mm)
Microwave radiation	1 mm–30 cm
Radio	≥ 30 cm

Two important categories are not shown in Table 2.3. The *optical spectrum,* from 0.30 to 15 μm, defines those wavelengths that can be reflected and refracted with lenses and mirrors. The *reflective spectrum* extends from about 0.38 to 3.0 μm; it defines that portion of the solar spectrum used directly for remote sensing.

The Ultraviolet Spectrum

For practical purposes, radiation of significance for remote sensing can be said to begin with the ultraviolet region, a zone of short-wavelength radiation that lies between the X-ray region and the limit of human vision. Often the ultraviolet region is subdivided into the *near ultraviolet* (sometimes known as *UV-A*; 0.32 to 0.40 μm), the *far ultraviolet* (*UV-B*; 0.32 to 0.28 μm), and the *extreme ultraviolet* (*UV-C*; below 0.28 μm). The ultraviolet region was discovered in 1801 by the German scientist Johann Wilhelm Ritter (1776–1810). Literally, "ultraviolet" means "beyond the violet," designating it as the region just outside the violet region, the shortest wavelengths visible to humans. Near ultraviolet radiation is known for its ability to induce *fluorescence,* emission of visible radiation, in some materials; it has significance for a specialized form of remote sensing (see p. 43). However, ultraviolet radiation is easily scattered by the earth's atmosphere, so it is not generally used for remote sensing of earth materials.

The Visible Spectrum

Although the visible spectrum constitutes a very small portion of the spectrum, it has obvious significance in remote sensing. Limits of the visible spectrum are defined by the sensitivity of

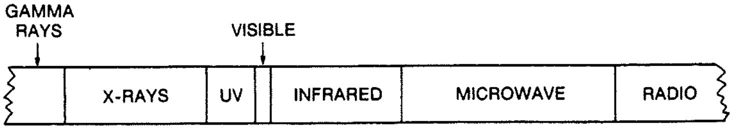


FIGURE 2.3. Major divisions of the electromagnetic spectrum. This diagram gives only a schematic representation—sizes of divisions are not shown in correct proportions. (See Table 2.3.)

the human visual system. Optical properties of visible radiation were first investigated by Isaac Newton (1641–1727), who during 1665 and 1666 conducted experiments that revealed that visible light can be divided (using prisms, or, in our time, diffraction gratings) into three segments. Today we know these segments as the *additive primaries,* defined approximately from 0.4 to 0.5 μm (blue), 0.5 to 0.6 μm (green), and 0.6 to 0.7 μm (red) (Figure 2.4). *Primary colors* are defined such that no single primary can be formed from a mixture of the other two, and that all other colors can be formed by mixing the three primaries in appropriate proportions. Equal proportions of the three additive primaries combine to form white light.

The color of an object is defined by the color of the light that it reflects (Figure 2.4). Thus a "blue" object is "blue" because it reflects blue light. Intermediate colors are formed when an object reflects two or more of the additive primaries, which combine to create the sensation of "yellow" (red and green), "purple" (red and blue), or other colors. The additive primaries are significant whenever we consider the colors of light, as, for example, in the exposure of photographic films.

In contrast, *representations* of colors in films, paintings, and similar images are formed by combinations of the three *subtractive primaries* that define the colors of pigments and dyes. Each of the three subtractive primaries absorbs a third of the visible spectrum (Figure 2.4). *Yellow* absorbs blue light (and reflects red and green); *cyan* (a greenish-blue) absorbs red light (and reflects blue and green); and *magenta* (a bluish red) absorbs green light (and reflects red and blue light). A mixture of equal proportions of pigments of the three subtractive primaries yields black (complete absorption of the visible spectrum). The additive primaries are of interest in matters concerning radiant energy, whereas the subtractive primaries specify colors of the pigments and dyes used in reproducing colors on films, photographic prints, and other images.

The Infrared Spectrum

Wavelengths longer than the red portion of the visible spectrum are designated as the infrared region, discovered in 1800 by the British astronomer William Herschel (1738–1822). This segment of the spectrum is very large relative to the visible region, as it extends from 0.72 to 15

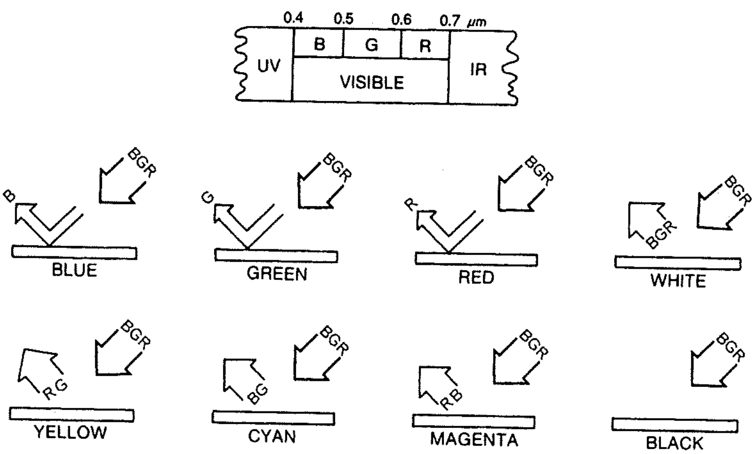


FIGURE 2.4. Colors.

μm—making it more than 40 times as wide as the visible light spectrum. Because of its broad range, it encompasses radiation with varied properties. Two important categories can be recognized here. The first consists of *near infrared* and *mid-infrared* radiation—defined as those regions of the infrared spectrum closest to the visible. Radiation in the near infrared region behaves, with respect to optical systems, in a manner analogous to radiation in the visible spectrum. Therefore, remote sensing in the near infrared region can use films, filters, and cameras with designs similar to those intended for use with visible light.

The second category of infrared radiation consists of the *far infrared* region, consisting of wavelengths well beyond the visible, extending into regions that border the microwave region (Table 2.3). This radiation is fundamentally different from that in the visible and the near infrared regions. Whereas near infrared radiation is essentially solar radiation reflected from the earth's surface, far infrared radiation is emitted by the earth. In everyday language, the far infrared consists of "heat," or "thermal energy." Sometimes this portion of the spectrum is referred to as the *emitted infrared.*

Microwave Energy

The longest wavelengths commonly used in remote sensing are those from about 1 mm to 1 m in wavelength. The shortest wavelengths in this range have much in common with the thermal energy of the far infrared. The longer wavelengths merge into the radio wavelengths used for commercial broadcasts. Our knowledge of the microwave region originates from the work of the Scottish physicist James Clerk Maxwell (1831–1879) and the German physicist Heinrich Hertz (1857–1894).

2.4. Radiation Laws

The propagation of electromagnetic energy follows certain physical laws. In the interests of conciseness, some of these laws are outlined in abbreviated form because our interest here is the basic relationships they express rather than the formal derivations that are available to the student in more comprehensive sources.

Isaac Newton was among the first to recognize the dual nature of light (and by extension, all forms of electromagnetic radiation), which simultaneously displays behaviors associated with both discrete and continuous phenomena. Newton maintained that light is a stream of minuscule particles ("corpuscles") that travel in straight lines. This notion is consistent with the modern theories of Max Planck (1858–1947) and Albert Einstein (1879–1955). Planck discovered that electromagnetic energy is absorbed and emitted in discrete units called *quanta,* or *photons.* The size of each unit is directly proportional to the frequency of the energy's radiation. Planck defined a constant (h) to relate frequency (v) to radiant energy (Q):

$$Q = hv \qquad \text{(Eq. 2.2)}$$

His model explains the photoelectric effect, the generation of electric currents by the exposure of certain substances to light, as the effect of the impact of these discrete units of energy (quanta) upon surfaces of certain metals, causing the emission of electrons.

Newton knew of other phenomena, such as the refraction of light by prisms, that are best explained by assuming that electromagnetic energy travels in a wave-like manner. James Clerk Maxwell was the first to formally define the wave model of electromagnetic radiation. His mathematical definitions of the behavior of electromagnetic energy are based upon the assumption from classical (mechanical) physics that light and other forms of electromagnetic energy propagate as a series of waves. The wave model best explains some aspects of the observed behavior of electromagnetic energy (e.g., refraction by lenses and prisms, and diffraction), whereas quantum theory provides explanations of other phenomena (notably, the photoelectric effect).

The rate at which photons (quanta) strike a surface is the *radiant flux* (ϕ_e), measured in watts (W); this measure specifies energy delivered to a surface in a unit of time. We also need to specify a unit of area; the irradiance (E_e) is defined as radiant flux per unit area (usually measured as watts per square meter). Irradiance measures radiation that strikes a surface, whereas the term *radiant exitance* (M_e) defines the rate at which radiation is emitted from a unit area (also measured in watts per square meter).

All objects with temperatures above absolute zero have temperature and emit energy. The amount of energy and the wavelengths at which it is emitted depend upon the temperature of the object. As the temperature of an object increases, the total amount of energy emitted also increases, and the wavelength of maximum (peak) emission becomes shorter. These relationships can be expressed formally using the concept of the *blackbody*. A blackbody is a hypothetical source of energy that behaves in an idealized manner. It absorbs all incident radiation; none is reflected. A blackbody emits energy with perfect efficiency; its effectiveness as a radiator of energy varies only as temperature varies.

The blackbody is a hypothetical entity because in nature all objects reflect at least a small proportion of the radiation that strikes them, and thus do not act as perfect reradiators of absorbed energy. Although truly perfect blackbodies cannot exist, their behavior can be approximated using laboratory instruments. Such instruments have formed the basis for the scientific research that has defined relationships between the temperatures of objects and the radiation they emit. *Kirchhoff's law* states that the ratio of emitted radiation to absorbed radiation flux is the same for all blackbodies at the same temperature. This law forms the basis for the definition of *emissivity* (ε), the ratio between the emittance of a given object (M) and that of blackbody at the same temperature (M_b):

$$\varepsilon = M/M_b \qquad \text{Eq. 2.3}$$

The emissivity of a true blackbody is 1, and that of a perfect reflector (a *whitebody*) would be 0. Blackbodies and whitebodies are hypothetical concepts, approximated in the laboratory under contrived conditions. In nature, all objects have emissivities that fall between these extremes (*graybodies*). For these objects, emissivity is a useful measure of their effectiveness as radiators of electromagnetic energy. Those objects that tend to absorb high proportions of incident radiation and then to reradiate this energy will have high emissivities. Those that are less effective as absorbers and radiators of energy have low emissivities (i.e., they return much more of the energy that reaches them). (In Chapter 8, further discussion of emissivity explains that emissivity of an object can vary with its temperature.)

The *Stefan–Boltzmann law* defines the relationship between the total emitted radiation (W) (often expressed in watts·cm^{-2}) and temperature (T) (absolute temperature, K):

$$W = \sigma T^4 \quad \text{(Eq. 2.4)}$$

Total radiation emitted from a blackbody is proportional to the fourth power of its absolute temperature. The constant (σ) is the Stefan–Boltzmann constant (5.6697×10^{-8}) (watts $\cdot$ m^{-2} $\cdot$ K^{-4}), which defines unit time and unit area. In essence, the Stefan–Boltzmann law states that hot blackbodies emit more energy per unit area than do cool blackbodies.

Wien's displacement law specifies the relationship between the wavelength of radiation emitted and the temperature of a blackbody:

$$\lambda = 2{,}897.8/T \quad \text{(Eq. 2.5)}$$

where λ is the wavelength at which radiance is at a maximum, and T is the absolute temperature (K). As blackbodies become hotter, the wavelength of maximum emittance shifts to shorter wavelengths (Figure 2.5).

All three of these radiation laws are important for understanding electromagnetic radiation. They have special significance later in discussions of detection of radiation in the far infrared spectrum (Chapter 8).

2.5. Interactions with the Atmosphere

All radiation used for remote sensing must pass through the earth's atmosphere. If the sensor is carried by a low-flying aircraft, effects of the atmosphere upon image quality may be negligible. In contrast, energy that reaches sensors carried by earth satellites (Chapter 6) must pass through the *entire depth* of the earth's atmosphere. Under these conditions, atmospheric effects may have substantial impact upon the quality of images and data that the sensors generate. Therefore, the practice of remote sensing requires knowledge of interactions of electromagnetic energy with the atmosphere.

In cities we often are acutely aware of the visual effects of dust, smoke, haze, and other atmospheric impurities due to their high concentrations. We easily appreciate their effects upon brightnesses and colors we see. But even in clear air, visual effects of the atmosphere are

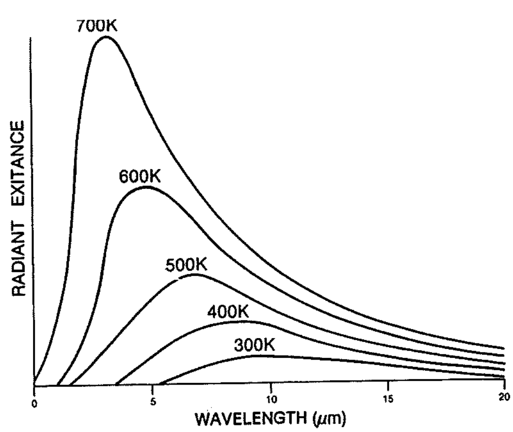


FIGURE 2.5. Wien's displacement law. For blackbodies at high temperatures, maximum radiation emission occurs at short wavelengths. Blackbodies at low temperatures emit maximum radiation at longer wavelengths.

numerous, although so commonplace that we may not recognize their significance. In both settings, as solar energy passes through the earth's atmosphere, it is subject to modification by several physical processes, including (1) scattering, (2) absorption, and (3) refraction.

Scattering

Scattering is the redirection of electromagnetic energy by particles suspended in the atmosphere or by large molecules of atmospheric gases (Figure 2.6). The amount of scattering that occurs depends upon the sizes of these particles, their abundance, the wavelength of the radiation, and the depth of the atmosphere through which the energy is traveling. The effect of scattering is to redirect radiation, so that a portion of the incoming solar beam is directed back toward space, as well as toward the earth's surface.

A common form of scattering was discovered by the British scientist Lord J. W. S. Rayleigh (1824–1919) in the late 1890s. He demonstrated that a perfectly clean atmosphere, consisting

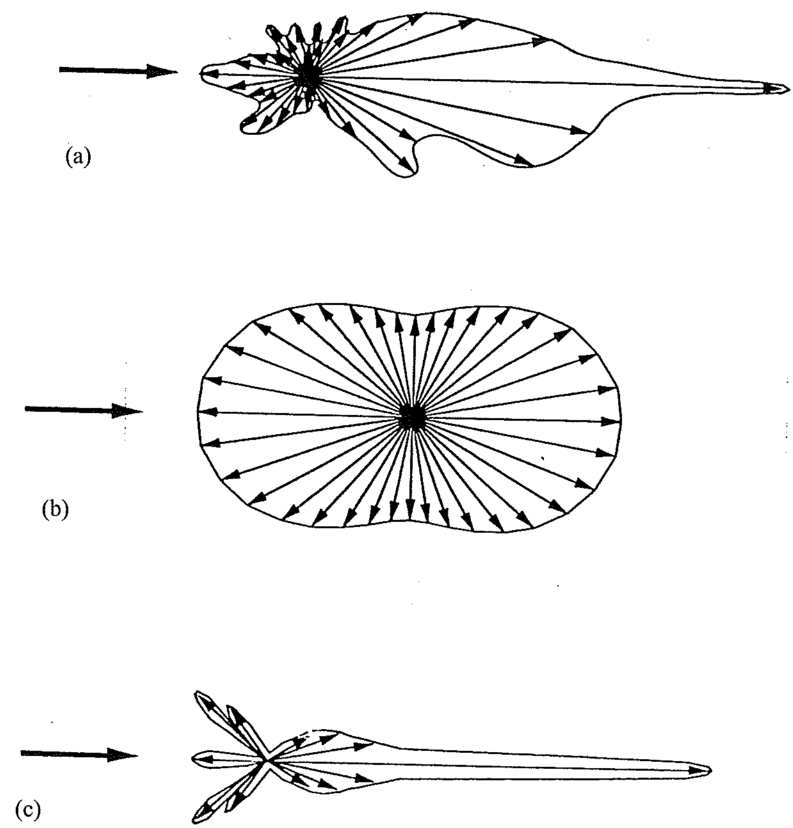


FIGURE 2.6. Scattering behaviors of three classes of atmospheric particles. (*a*) Atmospheric dust and smoke form rather large irregular particles that create a strong forward-scattering peak, with a smaller degree of backscattering. (*b*) Atmospheric molecules are more nearly symmetric in shape, creating a pattern characterized by preferential forward- and backscattering, but without the pronounced peaks observed in the first example. (*c*) Large water droplets create a pronounced forward-scattering peak, with smaller backscattering peaks. From Lynch and Livingston (1995). Reprinted with the permission of Cambridge University Press.

only of atmospheric gases, causes scattering of light in a manner such that the amount of scattering increases greatly as wavelength becomes shorter. *Rayleigh scattering* occurs when atmospheric particles have diameters that are very small relative to the wavelength of the radiation. Typically, such particles could be very small specks of dust, or some of the larger molecules of atmospheric gases, such as nitrogen (N_2) and oxygen (O_2). These particles have diameters that are much smaller than the wavelength of visible and near infrared radiation (on the order of diameters less than λ).

Because Rayleigh scattering can occur in the absence of atmospheric impurities, it is sometimes referred to as *clear atmosphere scattering.* It is the dominant scattering process high in the atmosphere, up to altitudes of 9 to 10 km, the upper limit for atmospheric scattering. Rayleigh scattering is *wavelength-dependent,* meaning that the amount of scattering changes greatly as one examines different regions of the spectrum (Figure 2.7). Blue light is scattered about four times as much as is red light, and ultraviolet light is scattered almost 16 times as much as is red light. *Rayleigh's law* states that this form of scattering is in proportion to the inverse of the fourth power of the wavelength.

Rayleigh scattering is the cause both for the blue color of the sky and for the brilliant red and orange colors often seen at sunset. At midday, when the sun is high in the sky, the atmospheric path of the solar beam is relatively short and direct, so an observer at the earth's surface sees mainly the blue light preferentially redirected by Rayleigh scatter. At sunset, observers on the earth's surface see only those wavelengths that pass through the longer atmospheric path caused by the low solar elevation; because only the longer wavelengths penetrate this distance without attenuation by scattering, we see only the reddish component of the solar beam. Variations of concentrations of fine atmospheric dust or of tiny water droplets in the atmosphere may contribute to variations in atmospheric clarity, and therefore to variations in colors of sunsets.

Although Rayliegh scattering forms an important component of our understanding of atmospheric effects upon transmission of radiation in and near the visible spectrum, it applies only to a rather specific class of atmospheric interactions. In 1906 the German physicist Gustav Mie (1868–1957) published an analysis that describes atmospheric scattering involving a broader range of atmospheric particles. *Mie scattering* is caused by large atmospheric particles, including dust, pollen, smoke, and water droplets. Such particles may seem to be very small by the standards of everyday experience, but they are many times larger than those responsible for

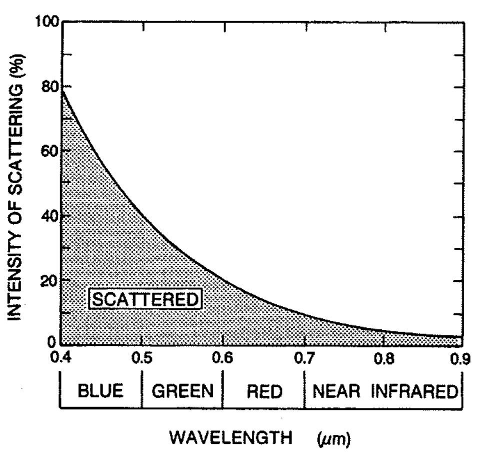


FIGURE 2.7. Rayleigh scattering. Scattering is much higher at shorter wavelengths.

Rayleigh scattering. Those particles that cause Mie scattering have diameters that are roughly equivalent to the wavelength of the scattered radiation. Mie scattering can influence a broad range of wavelengths in and near the visible spectrum; Mie's analysis accounts for variations in the size, shape, and composition of such particles. Mie scattering is wavelength-dependent, but not in the simple manner of Rayleigh scattering; it tends to be greatest in the lower atmosphere (0 to 5 km), where larger particles are abundant.

Nonselective scattering is caused by particles that are much larger than the wavelength of the scattered radiation. For radiation in and near the visible spectrum, such particles might be larger water droplets or large particles of airborne dust. "Nonselective" means that scattering is not wavelength-dependent, so we observe it as a whitish or grayish haze—all visible wavelengths are scattered equally.

Effects of Scattering

Scattering causes the atmosphere to have a brightness of its own. In the visible portion of the spectrum, shadows are not jet-black (as they would be in the absence of scattering), but are merely dark; we can see objects in shadows because of light redirected by particles in the path of the solar beam. The effects of scattering are also easily observed in vistas of landscapes—colors and brightnesses of objects are altered as they are positioned at locations more distant from the observer. Landscape artists take advantage of this effect, called *atmospheric perspective,* to create the illusion of depth by painting more distant features in subdued colors and those in the foreground in brighter, more vivid colors.

For remote sensing, scattering has several important consequences. Because of the wavelength dependency of Rayleigh scattering, radiation in the blue and ultraviolet regions of the spectrum (which is most strongly affected by scattering) is usually not considered useful for remote sensing. Images that record these portions of the spectrum tend to record the brightness of the atmosphere rather than the brightness of the scene itself. For this reason, remote sensing instruments often exclude short-wave radiation (blue and ultraviolet wavelengths) by use of filters or by decreasing sensitivities of films to these wavelengths. (However, some specialized applications of remote sensing, not discussed here, do use ultraviolet radiation.) Scattering also directs energy from outside the sensor's field of view toward the sensor's aperture, thereby decreasing the spatial detail recorded by the sensor. Furthermore, scattering tends to make dark objects appear brighter than they would otherwise be and bright objects appear darker, thereby decreasing the *contrast* recorded by a sensor (Chapter 3). Because "good" images preserve the range of brightnesses present in a scene, scattering degrades the quality of an image.

Some of these effects are illustrated in Figure 2.8. Observed radiance at the sensor, I, is the sum of I_S, radiance reflected from the earth's surface, conveying information about surface reflectance; I_O, radiation scattered from the solar beam directly to the sensor without reaching the earth's surface, and I_D, diffuse radiation, directed first to the ground, then to the atmosphere, before reaching the sensor. Effects of these components are additive within a given spectral band (Kaufman, 1984):

$$I = I_S + I_O + I_D \tag{Eq. 2.6}$$

I_S varies with differing surface materials, topographic slopes and orientation, and angles of illumination and observation. I_O is often assumed to be more or less constant over large areas,

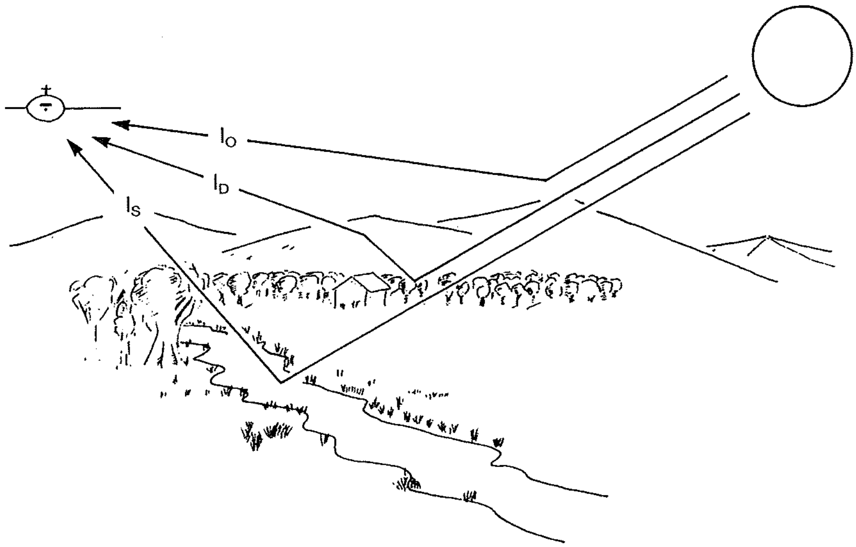


FIGURE 2.8. Principal components of observed brightness. I_S represents radiation reflected from the ground surface, I_O is energy scattered by the atmosphere directly to the sensor, and I_D represents diffuse light, directed to the ground, then to the atmosphere, before reaching the sensor. This diagram describes behavior of radiation in and near the visible region of the spectrum. From Campbell and Ran (1993). Copyright 1993 by Elsevier Science Ltd. Reproduced by permission.

although most satellite images represent areas large enough to encompass atmospheric differences sufficient to create variations in I_O. Diffuse radiation, I_D, is expected to be small relative to other factors, but varies from one land surface type to another, so in practice would be difficult to estimate. We should note the special case presented by shadows, in which $I_S = 0$, because the surface receives no direct solar radiation. However, shadows have their own brightness, derived from I_D, and their own spectral patterns, derived from the influence of local land cover upon diffuse radiation. Remote sensing is devoted to the examination of I_S at different wavelengths to derive information about the earth's surface. Figure 2.9 illustrates how I_D, I_S, and I_O vary with wavelength for surfaces of differing brightness.

Refraction

Refraction is the bending of light rays at the contact area between two media that transmit light. Familiar examples of refraction are the lenses of cameras or magnifying glasses (Chapter 3), which bend light rays to project or enlarge images, and the apparent displacement of objects submerged in clear water. Refraction also occurs in the atmosphere as light passes through atmospheric layers of varied clarity, humidity, and temperature. These variations influence the density of atmospheric layers, which in turn causes a bending of light rays as they pass from one layer to another. Everyday examples are the shimmering appearances on hot summer days of objects viewed in the distance as light passes through hot air near the surface of heated highways, runways, and parking lots.

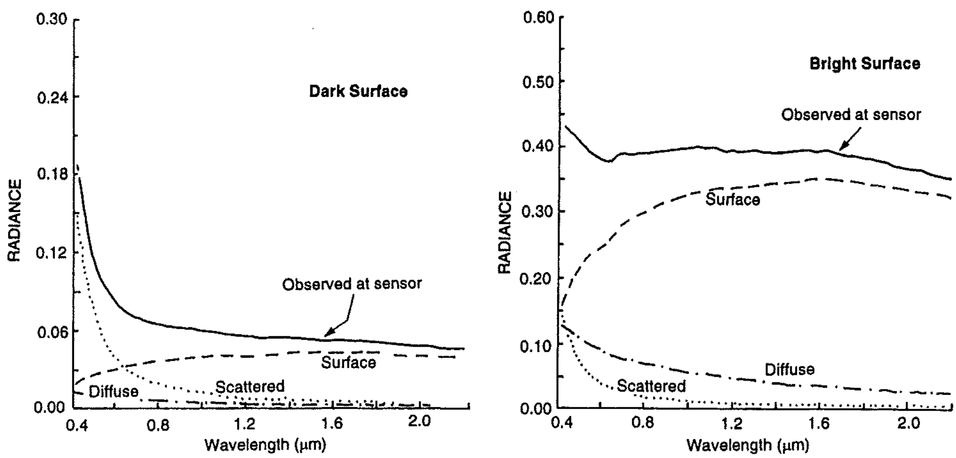


FIGURE 2.9. Changes in reflected, diffuse, scattered, and observed radiation over wavelength for dark (left) and bright (right) surfaces. This diagram shows the magnitude of the components illustrated in Figure 2.8. Atmospheric effects constitute a larger proportion of observed brightness for dark objects than for bright objects, especially at short wavelengths. Radiance has been normalized; note also differences in the scaling of the vertical axes for the two diagrams. Redrawn from Kaufman (1984). Reproduced by permission of the author and the Society of Photo-Optical Instrumentation Engineers.

The *index of refraction* (n) is defined as the ratio between the velocity of light in a vacuum (c) to its velocity in the medium (c_n):

$$n = c/c_n \qquad \text{(Eq. 2.7)}$$

Assuming uniform media, as the light passes into a denser medium it is deflected *toward* the *surface normal,* a line perpendicular to the surface at the point when the light ray enters the denser medium, as represented by the solid line in Figure 2.10. The angle θ' that defines the path of the refracted ray is given by *Snell's law:*

$$n \sin \theta = n' \sin \theta' \qquad \text{(Eq. 2.8)}$$

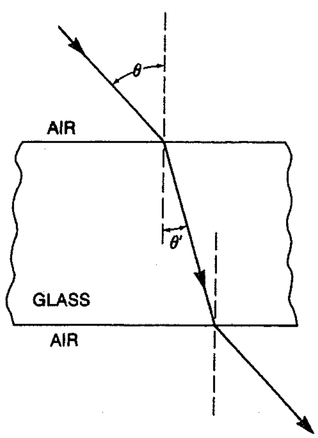


FIGURE 2.10. Refraction. The path of a ray of light as it passes from one medium (air) to another (glass), and then again as it passes back to the first.

where n and n' are the indices of refraction of the first and second media, respectively, and θ and θ' are angles measured with respect to the surface normal, as defined in Figure 2.10.

Absorption

Absorption of radiation occurs when the atmosphere prevents, or strongly attenuates, transmission of radiation or its energy through the atmosphere. (Energy acquired by the atmosphere is subsequently reradiated at longer wavelengths.) Three gases are responsible for most absorption of solar radiation. Ozone (O_3) is formed by the interaction of high-energy ultraviolet radiation with oxygen molecules (O_2) high in the atmosphere (maximum concentrations of ozone are found at altitudes of about 20 to 30 km in the stratosphere). Although naturally occurring concentrations of ozone are quite low (perhaps 0.07 parts per million at ground level, 0.1 to 0.2 parts per million in the stratosphere), ozone plays an important role in the earth's energy balance. Absorption of the high-energy, short-wavelength, portions of the ultraviolet spectrum (mainly λ less than 0.24 μm) prevents transmission of this radiation to the lower atmosphere.

Carbon dioxide (CO_2) also occurs in low concentrations (about 0.03% by volume of a dry atmosphere), mainly in the lower atmosphere. Aside from local variations caused by volcanic eruptions and mankind's activities, the distribution of CO_2 in the lower atmosphere is probably relatively uniform (although human activities that burn fossil fuels have apparently contributed to increases during the past 100 years or so). Carbon dioxide is important in remote sensing because it is effective in absorbing radiation in the mid- and far infrared regions of the spectrum. Its strongest absorption occurs in the region from about 13 to 17.5 μm, in the mid infrared.

Finally, water vapor (H_2O) is commonly present in the lower atmosphere (below about 100 km) in amounts that vary from 0 to about 3% by volume. (Note the distinction between *water vapor,* discussed here, and droplets of *liquid* water, mentioned previously.) From everyday experience we know that the abundance of water vapor varies greatly from time to time and from place to place. Consequently, the role of atmospheric water vapor, unlike those of ozone and carbon dioxide, varies greatly with time and location. It may be almost insignificant in a desert setting or in a dry air mass, but may be highly significant in humid climates and in moist air masses. Furthermore, water vapor is several times more effective in absorbing radiation than are all other atmospheric gases combined. Two of the most important regions of absorption are in several bands between 5.5 and 7.0 μm, and above 27.0 μm; absorption in these regions can exceed 80% if the atmosphere contains appreciable amounts of water vapor.

Atmospheric Windows

Thus the earth's atmosphere is by no means completely transparent to electromagnetic radiation because these gases together form important barriers to transmission of electromagnetic radiation through the atmosphere. It selectively transmits energy of certain wavelengths; those wavelengths that are relatively easily transmitted through the atmosphere are referred to as *atmospheric windows* (Figure 2.11). Positions, extents, and effectiveness of atmospheric windows are determined by the absorption spectra of atmospheric gases. Atmospheric windows are of obvious significance for remote sensing—they define those wavelengths that can be

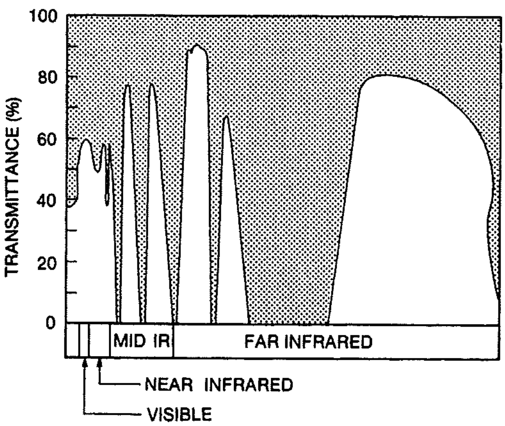


FIGURE 2.11. Atmospheric windows. This is a schematic representation that can depict only a few of the most important windows. The shaded area represents absorption of electromagnetic radiation.

used for forming images. Energy at other wavelengths, not within the windows, is severely attenuated by the atmosphere, and therefore cannot be effective for remote sensing. In the far infrared region, the two most important windows extend from 3.5 to 4.1 μm, and from 10.5 to 12.5 μm. The latter is especially important because it corresponds approximately to wavelengths of peak emission from the earth's surface. A few of the most important atmospheric windows are tabulated in Table 2.4; other smaller windows are not given here, but are listed in reference books.

Overview of Energy Interactions in the Atmosphere

Remote sensing is conducted in the context of all the atmospheric processes discussed thus far, so it is useful to summarize some of the most important points by outlining a perspective that integrates much of the preceding material. Figure 2.12 is an idealized diagram of interactions of shortwave solar radiation with the atmosphere; values are based upon typical, or average, values derived from many places and many seasons, so they are by no means representative of values that might be observed at a particular time and place. This diagram represents only the behavior of "shortwave" radiation (defined loosely here to include radiation with wavelengths less than 4.0 μm). It is true that the sun emits a broad spectrum of radiation, but the maximum intensity is

TABLE 2.4. Major Atmospheric Windows

Ultraviolet and visible	0.30–0.75 μm 0.77–0.91 μm
Near infrared	1.55–1.75 μm 2.05–2.4 μm
Thermal infrared	8.0–9.2 μm 10.2–12.4 μm
Microwave	7.5–11.5 mm 20.0+ mm

Note. Data selected from Fraser and Curran (1976, p. 35). Reproduced by permission of Addison-Wesley Publishing Co., Inc.

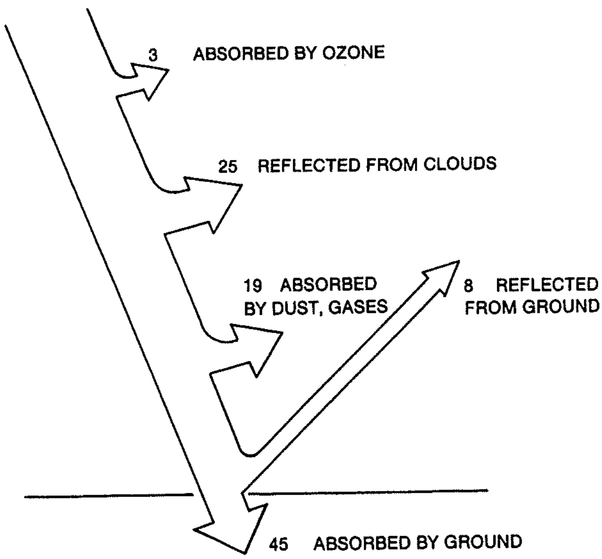


FIGURE 2.12. Incoming solar radiation. This diagram represents radiation at relatively short wavelengths, in and near the visible region. Values represent approximate magnitudes for the earth as a whole—conditions at any specific place and time would differ from those given here.

emitted at approximately 0.5 μm within this region, and little solar radiation at longer wavelengths reaches the ground surface.

Of 100 units of shortwave radiation that reach the outer edge of the earth's atmosphere, about 3 units are absorbed in the stratosphere as ultraviolet radiation interacts with O_2 to form O_3. Of the remaining 97 units, about 25 are reflected from clouds, and about 19 are absorbed by dust and gases in the lower atmosphere. About 8 units are reflected from the ground surface (this value varies greatly with different surface materials), and about 45 units ("about 50%") are ultimately absorbed at the earth's surface. For remote sensing in the visible spectrum, it is the portion reflected from the earth's surface that is of primary interest (see Figure 1.4), although knowledge of the quantity scattered is also important.

The 45 units that are absorbed are then reradiated by the earth's surface. From Wien's displacement law (Eq. 2.5), we know that the earth, being much cooler than the sun, must emit radiation at much longer wavelengths than does the sun. The sun, at 6,000 K, has its maximum intensity at 0.5 μm (in the green portion of the visible spectrum); the earth, at 300 K, emits with maximum intensity near 10 μm, in the far infrared spectrum.

Terrestrial radiation, with wavelengths longer than 10 μm, is represented in Figure 2.13. There is little, if any, overlap between the wavelengths of solar radiation, depicted in Figure 2.12, and the terrestrial radiation, shown in Figure 2.13. This diagram depicts the transfer of 143 units of long-wave radiation (as defined above) from the ground surface to the atmosphere, in three separate categories.

About 8 units are transferred from the ground surface to the atmosphere by "turbulent transfer" (heating of the lower atmosphere by the ground surface, which causes upward movement of air, then movement of cooler air to replace the original air). About 22 units are lost to the atmosphere by evaporation of moisture in the soil, water bodies, and vegetation (this energy is

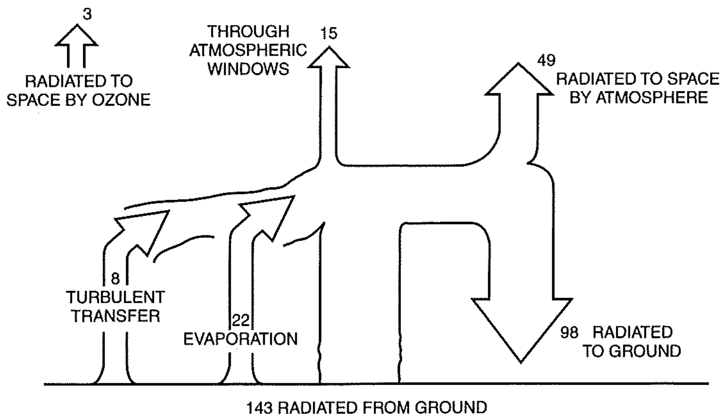


FIGURE 2.13. Outgoing terrestrial radiation. This diagram represents radiation at relatively long wavelengths—what we think of as sensible heat, or thermal radiation. Because the earth's atmosphere absorbs much of the radiation emitted by the earth, only those wavelengths that can pass through the atmospheric windows can be used for remote sensing.

transferred as the latent heat of evaporation). Finally, about 113 units are radiated directly to the atmosphere.

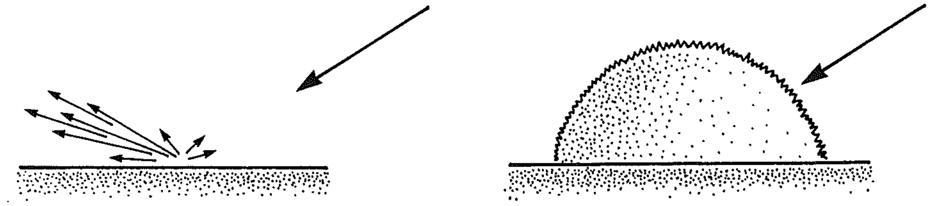
Because atmospheric gases are very effective in absorbing this long-wave (far infrared) radiation, much of the energy that the earth radiates is retained (temporarily) by the atmosphere. About 15 units pass directly through the atmosphere to space; this is energy emitted at wavelengths that correspond to atmospheric windows (chiefly 8 to 13 μm). Energy absorbed by the atmosphere is ultimately reradiated to space (49 units) and back to the earth (98 units). For meteorology, it is these reradiated units that are of interest because they are the source of energy for heating of the earth's atmosphere. For remote sensing, it is the 15 units that pass through the atmospheric windows that are of significance, as it is this radiation that conveys information concerning the radiometric properties of features on the earth's surface.

2.6. Interactions with Surfaces

As electromagnetic energy reaches the earth's surface, it must be reflected, absorbed, or transmitted. The proportions accounted for by each process depend upon the nature of the surface, the wavelength of the energy, and the angle of illumination.

Reflection

Reflection occurs when a ray of light is redirected as it strikes a nontransparent surface. The nature of the reflection depends upon sizes of surface irregularities (roughness or smoothness) in relation to the wavelength of the radiation considered. If the surface is smooth relative to wavelength, *specular* reflection occurs (Figure 2.14a). Specular reflection redirects all, or almost all, of the incident radiation in a single direction. For such surfaces, the angle of incidence is equal to the angle of reflection (i.e., in Eq. 2.8, the two media are identical, so $n = n'$,



(a) (b)

FIGURE 2.14. Specular (*a*) and diffuse (*b*) reflection. Specular reflection occurs when a smooth surface tends to direct incident radiation in a single direction. Diffuse reflection occurs when a rough surface tends to scatter energy more or less equally in all directions.

and therefore $\theta = \theta'$). For visible radiation, specular reflection can occur with surfaces such as a mirror, smooth metal, or a calm water body.

If a surface is rough relative to wavelength, it acts as a *diffuse,* or *isotropic,* reflector. Energy is scattered more or less equally in all directions. For visible radiation, many natural surfaces might behave as diffuse reflectors, including, for example, uniform grassy surfaces. A perfectly diffuse reflector (known as a *Lambertian surface*) would have equal brightnesses when observed from any angle (Figure 2.14b).

The idealized concept of a perfectly diffuse reflecting surface is derived from the work of Johann H. Lambert (1728–1777), who conducted many experiments designed to describe the behavior of light. One of Lambert's laws of illumination states that the perceived brightness (radiance) of a perfectly diffuse surface does not change with the angle of view. This is Lambert's cosine law, which states that that the observed brightness (I') of such a surface is proportional to the cosine of the incidence angle (θ), where I is the brightness of the incident radiation as observed at zero incidence:

$$I' = I/\cos\theta \qquad \text{(Eq. 2.9)}$$

This relationship is often combined with the equally important inverse square law, which states that observed brightness decreases according to the square of the distance from the observer to the source:

$$I' = \frac{I}{D^2}(\cos\theta) \qquad \text{(Eq. 2.10)}$$

(Both the cosine law and the inverse square law are depicted in Figure 2.15.)

Bidirectional Reflectance Distribution Function

Because of its simplicity and directness, the concept of a Lambertian surface is frequently used as an approximation of the optical behavior of objects observed in remote sensing. However, the Lambertian model does not hold precisely for many, if not most, natural surfaces. Actual surfaces exhibit complex patterns of reflection determined by details of surface geometry (e.g., the sizes, shapes, and orientations of plant leaves). Some surfaces may approximate Lambertian

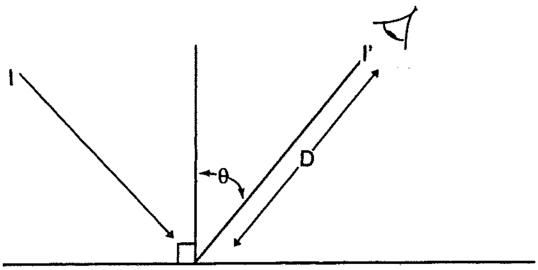


FIGURE 2.15. Inverse square law and Lambert's cosine law.

behavior at some incidence angles, but exhibit clearly non-Lambertian properties at other angles.

Reflection characteristics of a surface are described by the *bidirectional reflectance distribution function* (BRDF). The BRDF is a mathematical description of the optical behavior of a surface with respect to angles of illumination and observation, given that it has been illuminated with a parallel beam of light at a specified azimuth and elevation. (The function is "bidirectional" in the sense that it accounts both for the angle of illumination and the angle of observation.) The BRDF for a Lambertian surface has the shape depicted in Figure 2.14b, with even brightnesses as the surface is observed from any angle. Actual surfaces have more complex behavior. Description of BRDFs for actual, rather than idealized, surfaces permits assessment of the degrees to which they approach the ideals of specular and diffuse surfaces (Figure 2.16).

Transmission

Transmission of radiation occurs when radiation passes through a substance without significant attenuation (Figure 2.17). From a given thickness, or depth, of a substance, the ability of a medium to transmit energy is measured as the transmittance (t):

$$t = \frac{\text{Transmitted radiation}}{\text{Incident radiation}} \qquad \text{(Eq. 2.11)}$$

In the field of remote sensing, the transmittance of films and filters is often important. With respect to naturally occurring materials we often think only of water bodies as capable of transmitting significant amounts of radiation. However, the transmittance of many materials varies greatly with wavelengths, so our direct observations in the visible spectrum do not transfer to other parts of the spectrum. For example, plant leaves are generally opaque to visible radiation but transmit significant amounts of radiation in the infrared.

Fluorescence

Fluorescence occurs when an object illuminated with radiation of one wavelength emits radiation at a different wavelength. The most familiar examples are some sulfide minerals, which emit visible radiation when illuminated with ultraviolet radiation. Other objects also fluoresce, although observation of fluorescence requires very accurate and detailed measurements, not now routinely available for most applications. Figure 2.18 illustrates the fluorescence of healthy

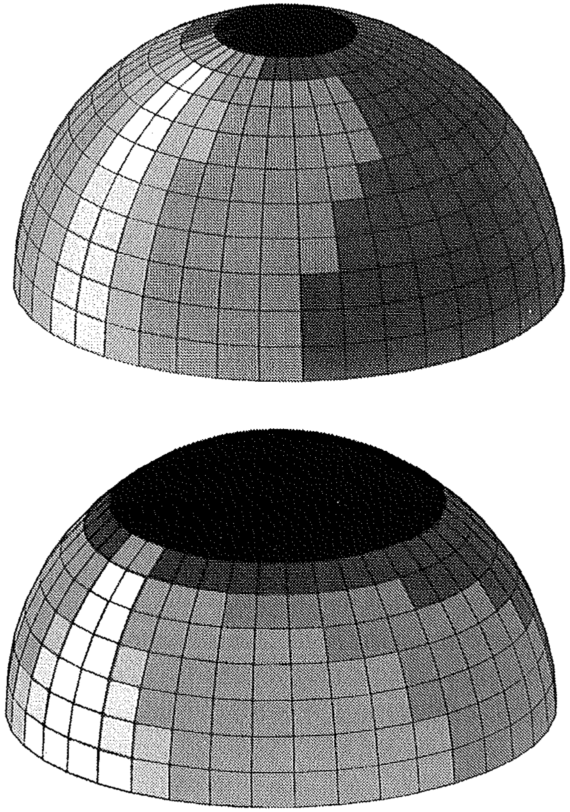


FIGURE 2.16. BRDFs for two surfaces. The varied shading represents differing intensities of observed radiation. (Calculated by Pierre Villeneuve.)

and senescent leaves, using one axis to describe the spectral distribution of the illumination and the other to show the spectra of the emitted energy. These contrasting surfaces illustrate the effectiveness of fluorescence in revealing differences between healthy and stressed leaves.

Spectral Properties of Objects

Remote sensing consists of the study of radiation emitted and reflected from features at the earth's surface. In the instance of emitted (far infrared) radiation, the object itself is the im-

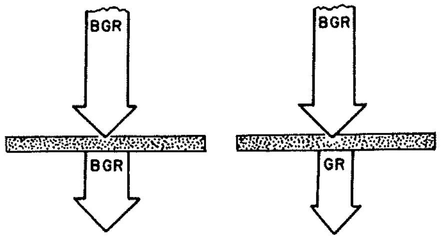


FIGURE 2.17. Transmission. Incident radiation passes through an object without significant attenuation (left), or may be selectively transmitted (right). The object on the right would act as a yellow ("minus blue") filter, as it would transmit all visible radiation except for blue light.

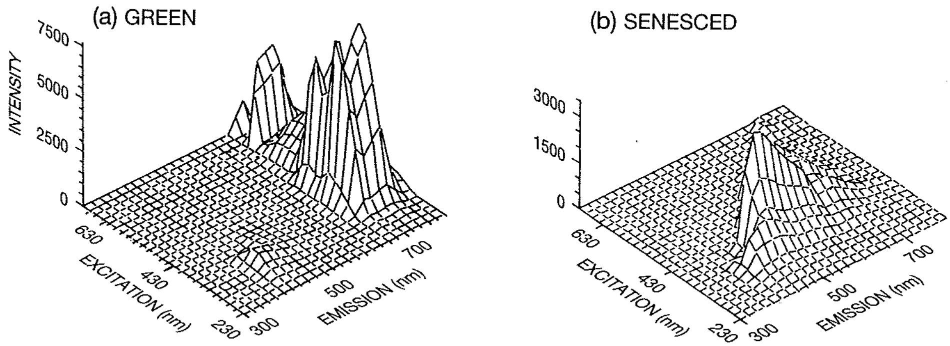


FIGURE 2.18. Fluorescence. Exitation and emission are shown along the two horizontal axes (with wavelengths given in nanometers). The vertical axes show strength of fluorescence, with the two examples illustrating the contrast in fluorescence between healthy (*a*) and senescent (*b*) leaves. From Rinker (1994).

mediate source of radiation. For reflected radiation, the source may be the sun, the atmosphere (by means of scattering of solar radiation), or man-made radiation (chiefly imaging radars).

A fundamental premise in remote sensing is that we can learn about objects and features on the earth's surface by studying the radiation reflected and/or emitted by these features. Using cameras and other remote sensing instruments, we can observe the brightnesses of objects over a range of wavelengths, so that there are numerous points of comparison between brightnesses of separate objects. A set of such observations or measurements constitute a spectral response pattern, sometimes called the *spectral signature* of an object (Figure 2.19). In the ideal, detailed knowledge of a spectral response pattern might permit identification of features of interest, such as separate kinds of crops, forests, or minerals. This idea has been expressed as follows:

> Everything in nature has its own unique distribution of reflected, emitted, and absorbed radiation. These spectral characteristics can—if ingeniously exploited—be used to distinguish one thing from another or to obtain information about shape, size, and other physical and chemical properties. (Parker and Wolff, 1965, p. 21)

This statement expresses the fundamental concept of the spectral signature, the notion that features display unique spectral responses that would permit clear identification, from spectral information alone, of individual crops, soils, and so on, from remotely sensed images. In practice, it is now recognized that spectra of features change both over time (e.g., as a cornfield grows during a season) and over distance (e.g., as proportions of specific tree species in a forest change from place to place).

Nonetheless, the study of the spectral properties of objects forms an important part of remote sensing. Some research has been focused upon examination of spectral properties of different classes of features. Thus, although it may be difficult to define unique signatures for specific kinds of vegetation, we can recognize distinctive spectral patterns for vegetated and nonvegetated areas, and for certain classes of vegetation, and we can sometimes detect the existence of diseased or stressed vegetation. In other instances, we may be able to define spectral patterns that are useful within restricted geographic and temporal limits as a means of studying the distributions of certain plant and soil characteristics. Chapter 14 describes how

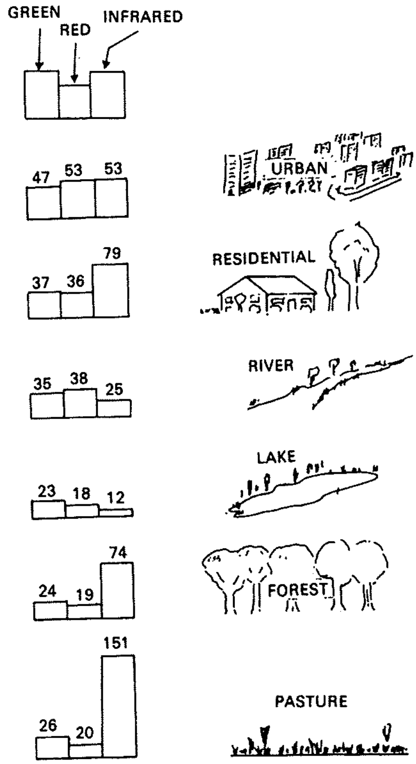


FIGURE 2.19. Spectral signatures.

very detailed spectral measurements permit application of some aspects of the concept of the spectral signature.

2.7. Summary: Three Models for Remote Sensing

Remote sensing typically takes one of three basic forms depending on the wavelengths of energy detected and on the purposes of the study. In the simplest form one records the reflection of solar radiation from the earth's surface (Figure 2.20). This is the kind of remote sensing most nearly similar to everyday experience. For example, film in a camera records radiation from the sun after it is reflected from the objects of interest, regardless of whether one uses a simple hand-held camera to photograph a family scene or a complex aerial camera to photograph a large area of the earth's surface. This form of remote sensing mainly uses energy in the visible and near infrared portions of the spectrum. Key variables include atmospheric clarity, spectral properties of objects, angle and intensity of the solar beam, choices of films and filters, and others explained in Chapter 3.

A second strategy for remote sensing is to record radiation *emitted* from (rather than *reflected* from) the earth's surface. Because emitted energy is strongest in the far infrared spectrum, this kind of remote sensing requires special instruments designed to record these wavelengths.

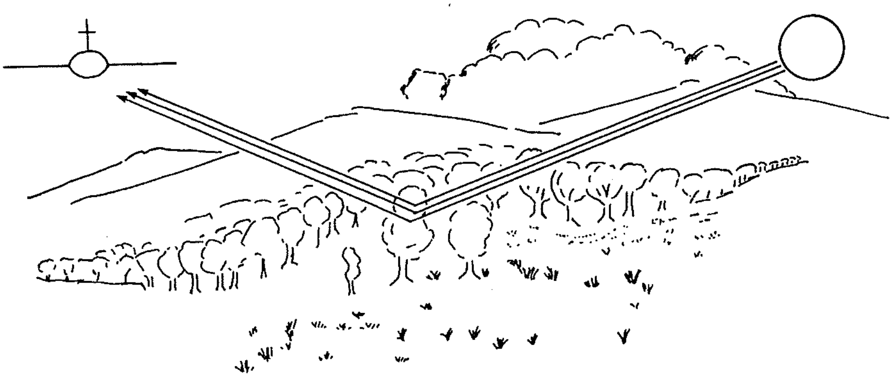


FIGURE 2.20. Remote sensing using reflected solar radiation. The sensor detects solar radiation that has been reflected from features at the earth's surface. (See Figure 2.12.)

(There is no direct analogue to everyday experience for this kind of remote sensing.) Emitted energy from the earth's surface is mainly derived from short-wave energy from the sun that has been absorbed, then reradiated at longer wavelengths (Figure 2.21). Emitted radiation from the earth's surface reveals information concerning thermal properties of materials, which can be interpreted to suggest patterns of moisture, vegetation, surface materials, and man-made structures. Other sources of emitted radiation (of secondary significance here, but often of primary significance elsewhere) include geothermal energy, heat from steam pipes, power plants, buildings, and forest fires. This example also represents "passive" remote sensing, because it employs instruments designed to sense energy emitted by the earth, not energy generated by a sensor.

Finally, a third class of remote sensing instruments generate their own energy, then record the reflection of that energy from the earth's surface (Figure 2.22). These are "active" sensors—"active" in the sense that they provide their own energy, so they are independent of solar and terrestrial radiation. As an everyday analogy, a camera with a flash attachment can be considered to be an active sensor. In practice, active sensors are best represented by imaging radars

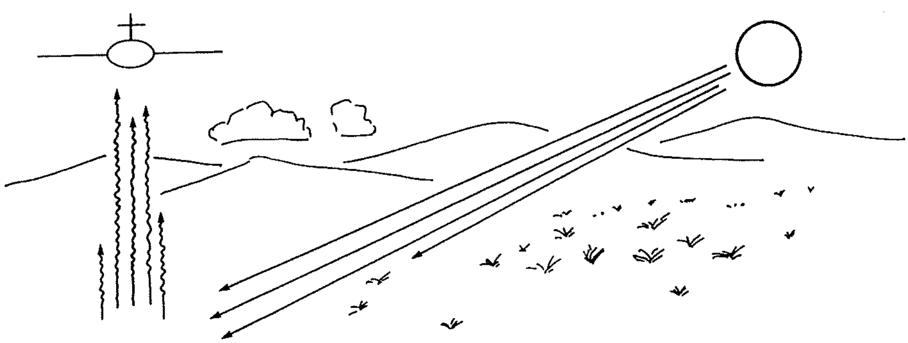


FIGURE 2.21. Remote sensing using emitted terrestrial radiation. The sensor records solar radiation that has been absorbed by the earth, then reemitted as thermal infrared radiation. (See Figures 2.12 and 2.13.)

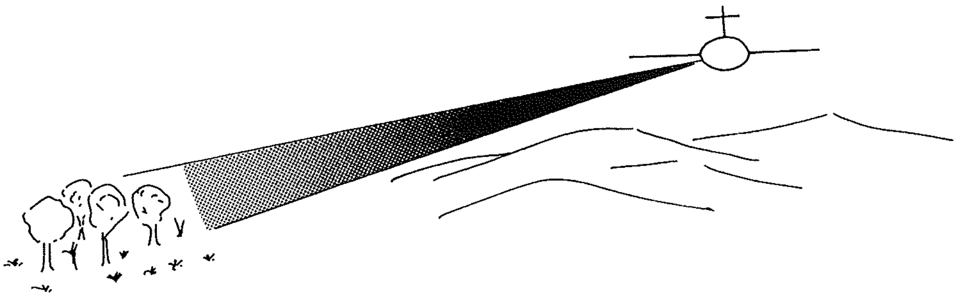


FIGURE 2.22. Active remote sensing. The sensor illuminates the terrain with its own energy, then records the reflected energy as it has been altered by the earth's surface.

and lidars (Chapter 7), which transmit energy toward the earth's surface from an aircraft or satellite, then receive the reflected energy to form an image. Because they sense energy provided directly by the sensor itself, such instruments have the capability to operate at night and during cloudy weather.

Review Questions

1. Using books provided by your instructor or available through your library, examine reproductions of landscape paintings to identify artistic use of atmospheric perspective. Perhaps some of your own photographs of landscapes illustrate the optical effects of atmospheric haze.
2. Some streetlights are deliberately manufactured to provide illumination with a reddish color. From material presented in this chapter, can you suggest why?
3. Although this chapter has largely dismissed ultraviolet radiation as an important aspect of remote sensing, there may well be instances where it might be effective, despite the problems associated with its use. Under what conditions might it prove practical to use ultraviolet radiation for remote sensing?
4. The human visual system is most nearly similar to which model for remote sensing as described in the last sections of this chapter?
5. Can you identify analogues from the animal kingdom for each of the models for remote sensing discussed in Section 2.7?
6. Examine Figures 2.12 and 2.13, which show the radiation balance of the earth's atmosphere. Explain how it can be that 100 units of solar radiation enter at the outer edge of the earth's atmosphere, yet 113 units are emitted by the atmosphere.
7. Examine Figures 2.12 and 2.13 again. Discuss how the values in this figure might change in different environments, including (a) desert, (b) the arctic, (c) an equatorial climate. How might these differences influence our ability to conduct remote sensing in each region?
8. Special signatures can be illustrated using values indicating the brightness in several spectral regions.

	UV	Blue	Green	Red	IR
Forest	28	29	36	27	56
Water	22	23	19	13	8
Corn	53	58	59	60	71
Pasture	40	39	42	32	62

Assume for now that these are "pure" signatures, not influenced by effects of the atmosphere. Can all categories be reliably separated, based upon these spectral values? Which bands are most useful for distinguishing between these classes?

9. Describe ideal atmospheric conditions for remote sensing.

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