August 2004

The Physics of Radiation

(required supplement to textbook readings)

The subject of remote sensing naturally requires a basic knowledge of the physics of radiation. Especially important are the methods for describing the angular and spectral fields of radiation and the way that radiation is influenced by emission, absorption and reflection (or scattering) from solid and liquid surfaces, small particles and gases.

1. The spectral properties of radiation

It is essential that we distinguish between the different wavelengths of light, as the interaction of light with solids, particles and gases is strongly sensitive to wavelength. As a result, the relationship between intensity and wavelength generally carries much of the information about the objects which the radiation has touched. Because the wavelength can vary over many orders of magnitude, we use various notations for describing it. The most commonly used length unit for wavelength is the micrometer (i.e. micron) defined as one millionth of a meter. For very short wavelengths, the nanometer is useful (i.e. one trillionth of a meter). For longer waves, the millimeter, centimeter or meter can be used. By convention, the full spectrum of electromagnetic radiation is subdivided in to the following ranges: gamma rays, x-rays, ultraviolet, visible, infrared, microwave and radiowaves. These ranges can be further subdivided; for example, the ultraviolet (.12 to .4 microns) can be divided into the far UV (.2 to .3 microns) and the near UV (.3 to.4 microns). The visible range (.4 to .7 microns), that range to which the human eye is sensitive, can be crudely divided into blue (.4 to .5 microns), green (.5 to .6) and the red (.6 to .7). A much finer subdivision by wavelength in the visible range is used in the study of color. The infrared (.7 to about 400 microns) is divided in to the near infrared (i.e. NIR; .7 to 2 microns), middle infrared (2 to 5), the "thermal" infrared (5 to 100) and the far IR. Microwaves (1mm to 1 meter) are often divided into "bands"; such as X-band, L-band etc., based on wavelength. Radio wave subdivisions are referred to by their wavelengths directly (e.g. 2-meter, 6-meter) by HAM radio operators.

In describing how much radiation is present with a particular wavelength, care must be taken to use precise language. Except for certain theoretical studies, or perhaps for laser light, a precisely specified wavelength (e.g. 0.755555555…microns) would have no radiation. Only a RANGE of wavelengths (e.g. 0.75 to 0.76 microns) would have finite radiative energy. If the specified range is sufficiently small, the amount of radiation is proportional to the range. That is, the amount of radiation between wavelengths 0.75 and 0.76 microns (range 0.01) would be ten times that falling between 0.755 and 0.756 (range 0.001). Mathematical functions describing the field of radiation by wavelength must carry the extra unit (inverse wavelength) so that they can be multiplied by the incremental wavelength range to get the amount of radiation (see the discussion of the Planck function).

Two alternate indicators of spectral position are used, the wavenumber and the frequency. The wavenumber (i.e. the number of waves in a given unit of distance) is defined as the inverse of the wavelength. It could have units of inverse centimeters for example. The frequency is the number of waves passing a location per unit time. It is related to wavelength by frequency = c/wavelength, where c is the speed of light. Frequency is usually given in hertz, kilohertz, megahertz, etc. One 'hertz' is one cycle per second.

For those of us who reside on planet earth, the absorptive property of the earth's atmosphere is an important factor in remote sensing. Unfortunately, the earth's atmosphere is opaque to large portions of the electromagnetic spectrum. Radiation reaching the observer in these wavelengths will have been absorbed and re-emitted by the atmosphere and will not contain information about the object we are trying to observe. Luckily, there are several discrete portions of the spectrum for which the atmosphere is transparent. We call these spectral segments "atmospheric windows" or just "windows". The most important windows are:1) The visible/NIR window (0.4 to about 1 micron), 2) The IR window (in the thermal IR between 8 and 12 microns) and 3) the microwave/radio window (wavelengths longer than about 1 cm.). Almost all remote sensing of "objects" uses these windows. The non-window wavelengths are being used for remote sensing of the atmosphere itself (i.e. mapping out its temperature, density and chemical composition). The human eye has evolved to use the visible window. The location of the windows is determined by the gases existing in the atmosphere. For earth, these include the air (N_2, O_2 and Argon) and the greenhouse gases (H_2O,CO_2, O_3, N_2O, etc.)

On occasion, it is useful to describe the total amount of radiation, without regard for its spectral distribution. An example of such a quantity is the Solar Constant; the total radiant energy, per unit time and area, received from the sun, at the position of the earth's orbit. This quantity has a value of S=1380 Watts/m2. A spectrally resolved solar constant, describing how much each wavelength contributes to the sun's radiant energy would have units of Watts/m2/micron.

2. The angular distribution of radiation.

There are two common ways of describing the amount of radiation present in a certain environment: angularly resolved and unresolved. The term intensity (or radiance) refers to the angularly resolved description of radiation; i.e. the information needed to create an image. The human eye, and a video camera, are devices to detect radiance. Such devices have an optical focusing system and an array of detectors on the focal plane. Hundreds or thousands of brightness values at each wavelength are needed to describe the radiance.

When describing the angular distribution of radiation (i.e. the radiance), we must use precise language. If we ask how much radiation is approaching the sample point from a precisely specified direction (e.g. azimuth= 26.5555… degrees and elevation

=85.33333…degrees) the correct answer would usually be zero. Only some finite range of angles would contain a finite amount of radiation. (Recall that the proper unit for a range of angles in space (not in a plane) is the steradian. The "solid angle" Ω of an object of area A, at a distance D, is Ω=A/D^2. For example, an object with area 1 m^2 at a distance of 100 meters would subtend a solid angle of Ω=1/10000=0.0001 steradians.)

The term flux (or irradiance) refers to the angularly unresolved radiation; i.e. the total radiation reaching a surface without regard to the direction from which it comes. The human eye, with frosted glasses, would detect this quantity. Only a single brightness value at each wavelength is needed to describe the flux.

The measurement or calculation of flux is always done with reference to an oriented surface. Rays approaching a surface at right angles make a full contribution to the flux while slanting rays make a smaller contribution. This reduction in contribution is described by the cosine of the angle between the "surface normal vector" and the incoming ray. As an example, consider a horizontal section of the earth's surface. The point directly above this surface, its "zenith point", serves as reference for the "zenith angle" of the incoming radiation. If the sun's beam has a zenith angle of 30 degrees (i.e. the sun is 30 degrees away from the zenith point) the flux striking the surface will be F=S*cos(zenith angle)=1380*cosine(30)= 1195 watts/m^2. At sunrise and sunset, the solar zenith angle is 90 degrees and the flux hitting a horizontal surface is zero.

The clear and correct use of radiation quantities is made easier if the proper units are always included. To review:

a) total irradiance	Watts/m^2
b) spectrally resolved irradiance	Watts/m^2/micron
c) total radiance	Watts/m^2/steradian
d) spectrally resolved radiance	Watts/m^2/steradian/micron

It is the last of these quantities, spectrally resolved radiance (or just "radiance"), that is most widely used in satellite remote sensing. Let's consider an example; a satellite sensor looking down at a uniform region of desert. If the radiance at the satellite, reflected from the desert scene, is 1 watt/m^2/steradian/micron, then the radiant energy approaching the satellite within a wavelength range of 0.1 micron and a solid angle range of 0.0001 steradian would be $1*10^{-5}$ watt/m^2. If the aperture of the sensor has an area of .01 m^2, the radiant energy received by the sensor would be $1*10^{-7}$ watts.

3. Interaction of radiation with matter

The interaction of radiation and a solid surface involves two processes: emission and reflection. Typically, the radiation beaming away from any surface is the sum of emitted and reflected light. For most surfaces and wavelengths, the emission of radiation is close to the Black Body approximation described by the Planck Function:

$$B_\lambda(T) = \frac{2hc^2\lambda^{-5}}{(e^{hc/k\lambda T} - 1)}$$

where $h = 6.6262 \times 10^{-27}$ erg sec (Planck's constant)
$k = 1.3806 \times 10^{-16}$ erg deg^{-1} (Boltzman's constant)
$c = 2.99793 \times 10^{10}$ cm/sec (speed of light)

When evaluating this function, T is the object temperature in Kelvins and λ is the wavelength in centimeters. The Planck can be written in several ways. The particular form given above is based on wavelength λ, and the constants are given in the cgs system of units. If the constants were given in the SI (or mks) system, then the input and output quantities should also be in the SI system.

This formula, one of the triumphs of statistical physics, is well treated in advanced texts. We only mention a few interesting aspects. First, B gives the radiance emitted from a surface, that is, power per unit area per unit solid angle per unit wavelength. For example, B could have the (SI) units of watts/m^2/steradian/m (or some equivalent units such as watts/m^3/steradian). To obtain a quantity with units of power (i.e. Watts), increments of emitting area (dA), conical solid angle ($d\Omega$) and wavelength range ($d\lambda$) must be chosen. Then, the power emitted by that area into the specified cone, within the specified wavelength range is

$$B_\lambda(T)dA\, d\Omega\, d\lambda$$

When plotted against wavelength (λ), B has a bell shape with a peak at λ_{max}; the wavelength being most profusely emitted by the object. This special wavelength is given by Wien's Law; λ_{max}=C/T; where the constant C=2898 microns* $Kelvins^{-1}$. Thus, hotter objects tend to emit at shorter wavelengths. As an example, a body with a temperature of 10C (283K) will emit most strongly at a wavelength of λ_{max}= 2898/283= 10.2 microns (i.e. in the thermal infrared). The sun, with a surface temperature of 6000K, will emit most strongly at λ_{max}= 2898/6000= 0.483 microns (i.e. in the middle of the visible range)

Note that the chemical or physical nature of the emitting object does not influence B; it depends on temperature only. The Black Body radiance described by B is also predicted to be isotropic; independent of angle. If the Black Body radiance is integrated over all wavelengths and angles, the total irradiance is found to be

$$F=\sigma*T^4 \quad \text{with units Watts/m}^2$$

This is the Stephan-Boltzman Law, and σ is the Stephan-Boltzman constant. The temperature in this formula must always be given in kelvins.

Under certain circumstances, thermal emission can be significantly less than predicted by the Black Body Laws. To describe this behaviour we define the emissivity (ε) as the ratio of the actual emission to the Black Body prediction. If $\varepsilon = 1$, the emission is following the Black Body rule. If $\varepsilon < 1$, the body emits less than the Black Body Law predicts. The emissivity varies with wavelength. In the thermal infrared part of the spectrum, ε seldom drops below 0.9 while in the distant microwave spectrum ε can be as small as 0.2 or 0.3.

The other essential radiation process occurring at solid surfaces is reflection. There are two type of reflection: specular and diffuse. Specular reflection is dominant from a smooth, shiny surface such as a mirror, a waxed car fender or a glassy smooth lake. The reflected fraction of the incoming beam has an angle equal to the incident angle. If the incident light is a single narrow beam then the reflected light will also be beam-like. If the incident light is isotropic, like sky-light, the reflected light will also be isotropic. Little or no color shift occurs during specular reflection; the spectral content of the reflected light is the same as the incident light. The reflected light contains no information about the chemical composition or temperature of the object.

Diffuse reflection occurs on rough surfaces such as paper, cloth, grass, matte-finished paint, snow, soil, etc. To a good approximation, diffuse reflection is isotropic, even if the incident light is in a tight beam. Most important, diffuse reflection usually includes a color shift due to partial absorption. If the incident light is white (i.e. an equal mixture of red, green and blue) a red reflection, for example, will occur by absorption of green and blue. In this way, the reflected light carries away information about the chemical composition of the object. As the object temperature plays little or no role in reflection, the reflected light contains no temperature information. Clearly, it is the process of diffuse reflection that underlies much of remote sensing practice, as well as human color vision.

To characterize the reflective properties of a surface, we define the reflectivity (or reflectance) R.

R= reflected light/incident light

Since the reflected light can never exceed the incident light (note: We are neglecting emission here.), the quantity R can never exceed unity. Note that the reflectivity may be a strong function of wavelength. A particular surface may absorb some wavelengths while reflecting others. In practice, there are three ways to measure reflectivity. Each one requires some measurement of incident and reflected radiation, either a flux (F) or a radiance (I).

a) ratio of fluxes

If it is possible to measure the incident and reflected fluxes (Fi and Fr), then the reflectivity is R=Fr/Fi.

b) incident flux and reflected radiance

If it is possible to measure the incident flux (Fi) and the reflected radiance (Ir), assumed isotropic, then the reflectivity is $R=\pi*Ir/Fi$. This formula follows from the formula for flux in a situation with isotropic radiance over a hemisphere; $F=\pi*I$.

c) radiances and a reference block

If a perfect diffuse reflector is available, the radiance reflected from the test object (It) and the perfect reference block (Iref) can measured. The reflectivity is then R=It/Iref.

No single surface is more important in remote sensing, or more interesting physically, than a simple green leaf. Its reflectivity $R(\lambda)$ is a strong function of wavelength. In the blue and red wavelengths, the reflectivity is very low (about 0.1 or 10%). In the green, the reflectivity is much higher, but still modest (about 0.2 or 20%). This pattern of wavelength-dependent reflectivity over the visible part of the spectrum gives the leaf its green appearance to the human eye. At longer wavelengths, just beyond the red (λ> 0.7 microns) the reflectivity rises sharply. In the NIR band, R is typically 0.5 to 0.8. Other species (or aliens) with eyes sensitive in the NIR would perceive this "dark" vegetation as "bright". In the field of remote sensing, the ratio of reflectivity in the NIR and visible is taken as an indication of green plant biomass.

A widely used measure of the ratio of NIR to visible reflectance is the Normalized Difference Vegetation Index (NDVI). If we adopt the terms R4 and R3* for the reflectance in the NIR and Visible range respectively,

$$NDVI = (R4-R3)/(R4+R3)$$

Several properties of NDVI are evident. Obviously, its value is a measure of how much R4 exceeds R3. The word "normalized" refers to the dividing of (R4-R3) by the sum (R4+R3). This normalization gives the NDVI two nice properties: a simple range from –1 to +1 and a insensitivity to the overall magnitude of the reflectance values. That is, if R4 and R3 are multiplied by a constant value, that factor will cancel out of the formula leaving NDVI unchanged. Green plants typically have NDVI in the 0.5 to 0.8 range. Soils have NDVI near to zero. Snow, clouds and lakes can have negative NDVI values as they reflect more strongly in the Visible than the NIR wavelengths. Occasionally, NDVI is computed directly from satellite-received radiance values in the NIR and visible channels, rather than from reflectance values. The resulting NDVI will not agree quantitatively with that computed from reflectance. (* This choice of symbols for NIR and visible wavelengths comes from the popular Landsat TM instrument. The TM sensor has 7 channels. Channels 3 and 4 are the red and NIR bands respectively.)

The assumption is often made that reflected light from a rough scene, like a forest, would be purely diffuse and isotropic (i.e. independent of angle) even though most of the illumination is the pencil-thin direct solar beam. Recent work has shown that this assumption can be inaccurate. In addition to a nearly isotropic field of reflected light, two preferred angles of reflection are seen. The first of these angles is at the specular angle. The increased brightness at the specular angle is due to specular reflection from small horizontal surfaces such as pools of water, or with some tree species, glare from horizontally oriented leaves. The second preferred angle is the so-called "hot-spot", directly back towards the sun. This is a characteristic of plants with a large Leaf Area Index (LAI); the ratio of total leaf area to footprint. When observed from most directions, most of the leaves of a tree are shadowed by other leaves, reducing the apparent reflected intensity. When viewed from the sun's position in the sky however, every leaf seen by the observer is sunlit. This gives a brighter reflection. It has been suggested that the observation of enhanced "hot spot" reflection be used as an indication of LAI.

4. An earth-like scene

Having reviewed the fundamentals of radiation, we are ready to consider a typical daytime or night time scene on earth. Imagine a region with trees, grass, soil, lakes and clouds illuminated (or not) by the sun. A satellite in orbit observes the scene and measures the radiance pixel-by-pixel at different wavelengths.

At night, the situation is simple. With no illuminating radiation from the sun, the only radiation reaching the satellite is that emitted from the objects in the scene. If all the surface objects (e.g. trees, grass, soil, lake) are at the same temperature, they will all emit with the same intensity. The Thermal IR sensors on the satellite will detect this emission, but will be unable to distinguish between all these objects. If the lake is cooler or warmer than the land, it will be distinguishable. The clouds, occurring at different levels in the atmosphere, will have different temperatures. The higher clouds will be colder and will emit less radiation. The satellite sensors will detect this difference. As the Planck function $B(\lambda,T)$ gives a small value when evaluated for earthlike temperatures and shorter wavelengths, the visible and NIR sensors on the satellite will not detect significant emitted radiation.

The daytime situation is more complex. First, as the sun's illumination in thermal IR wavelengths is rather small, the thermal radiation received by the satellite from the earth's surface in daylight will not differ substantially from the nighttime received values. An exception to this might occur if the surface warms up during the day and begins to emit more strongly. In the shorter wavelengths however, the day/night difference is profound. With strong illumination, the reflected visible and NIR will reach the satellite in proportion to the reflectivity of the objects in the scene. The snow and clouds will appear bright in all these wavelengths. The grass and trees will appear dark in the visible wavelengths but bright in the NIR. The soil will be of medium brightness at all the Vis/NIR wavelengths while the lake may appear very dark.

Probably the most complex aspect of the daytime situation is the brightness of the scene in the middle IR part of the spectrum (e.g. near 3 microns). At these wavelengths, the outgoing radiation would be a mixture of reflected sunlight and emitted radiation. The intensity of that radiation will depend on both the reflectivity and the temperature of the objects. A quantitative interpretation of a daytime satellite image in the middle IR would be difficult.

5. Atmospheric effects

In the brief description above, we neglected the radiative effect of the atmosphere. First, unless we are in a perfect spectral "window", gaseous absorption of Vis/NIR radiation will occur as the sunlight penetrates the atmosphere to the surface and as the reflected radiation beams upward to the satellite. In the Thermal IR wavelengths, gaseous emission may occur as well. Even if the window is perfect (i.e. no gaseous emission or absorption), scattering of light by molecules or small aerosol in the atmosphere may be important. As a result of all those atmospheric processes, the radiance received at the satellite will not be a perfect measure of the reflectivity of the objects on the ground that we are trying to identify. Several methods to correct satellite observations for atmospheric effects have been proposed, but a comprehensive treatment of these goes beyond the scope of the current discussion. In the paragraph below, we discuss one of the most important corrections; that for path radiance.

The subject of path radiance is evidently important to any one who has observed a natural scene. Consider what you see when looking up, while standing in an open field. No object subtends your line of sight and the background of outer space is dark. Yet, your eyes detect considerable radiance which has been scattered from the solar beam into your line of sight. This "sky-light" is blue (on a clear day) because the scattering is done by very small sub-micron particles which scatter short wavelengths more strongly (i.e. Rayleigh scattering proportional to λ^{-4}). On a hazy or cloudy day, the scattered light will be whiter, closer to the sunlight's spectral composition, as the particles are larger. These particles scatter all wavelengths equally (Mie scattering).

Consider next, a horizontal view of a distant mountain range, covered with dark forests. The nearest hills appear quite dark. The more distant hills are progressively brighter. Why is this so? With a dark forest background, most of the light received by the observer will be path radiance. Sunlight passing through the line of sight will be scattered into the direction towards the observer. The longer the line of sight, the greater will be the perceived radiance, until a limit is reached (the limit of visibility) where the rate of scattering into the line-of-sight equals the scattering out.

Now consider a satellite looking down at a dark spot on the surface of the earth (e.g. a shadow, a lake, a forest, a dark moist soil, etc.). The dark object will reflect little of the sun's radiance upwards to the satellite. Along the path between the object and the satellite however, sunlight will be scattered upward unto the sensor aperture. To determine the actual reflectivity of the surface object, we must subtract off the path radiance. An estimate of the path radiance can be made by finding the darkest object in the scene and assuming that its satellite-received-radiance is entirely due to path radiance. Path radiance is an especially serious problem in the shorter wavelengths. On a hazy day, it is not unusual in the blue channel for path radiance to contribute more than half of the satellite-received radiance.

Clouds are a common problem in remote sensing. Clouds are composed of liquid or ice particles with diameters varying all the way from 10microns to 2 mm. If the cloud has a large density of particles and is thick, it will completely scatter or absorb radiation. No information about the surface of the earth will reach the satellite. Only microwave radar could penetrate such a cloud. Occasionally, with thin cirrus clouds, surface features can be clearly discerned from satellite through the cloud, but the quantitative interpretation of the radiance values is compromised. Cirrus clouds will increase the visible and NIR radiation reaching the satellite by scattering sunlight upwards, while decreasing the thermal infra-red radiation (TIR). Cirrus cloud are very cold (typically T=-60C) and thus their emission of TIR is small while their absorption of earth-emitted TIR is significant.

6. New terms (self-examination)

The student should understand and define the following terms:

Aerosol
Aperture
Absorption
Atmospheric window
Azimuth
Black Body radiation
Blue light
Cirrus cloud
Cosine
Diffuse reflection
Elevation angle
Emission
Emissivity
Flux (irradiance)
Frequency
Green light
Greenhouse gases
Haze
Infrared radiation
Intensity (radiance)
Irradiance
Isotropic
Landsat TM
Leaf Area Index (LAI)
Micron
Microwave
Mie scattering
Millimeter
Nadir point
Nanometer
NDVI

Near IR
Path radiance
Pi (geometric ratio:
3.14159)
Pixel
Planck function
Power
Radian
Radiance
Radiowave
Rayleigh scattering
Red light
Reflectance
Skylight
Solar Constant
Solid Angle
Spectrometer
Specular reflection
Stephan-Boltzman Law
Steradian
Temperature
Thermal infrared
radiation
Ultra-violet radiation
Visible radiation
Watts
Wavelength
Wavenumber
White light
Window (spectral)
Wien's Law
Zenith angle
Zenith point

7. Further reading

High school and college level physics books usually include chapters on electromagnetic radiation. For a more advanced treatment, books on radiative transfer, atmospheric radiation, color perception or remote sensing are useful. Chapter 1 in Mather's book "Computer processing of remotely-sensed images" offers a compact introduction as well.

RBS, August, 2004