4. Radiation Through the Atmosphere

Radiation that travels through the atmosphere will be attenuated and distorted because atmospheric gases, particles and other aerosols scatter and absorb radiant energy. Moreover, the atmosphere may add radiation ("noise") to the radiation ("signal") sensed from an object by scattering incident energy or re-radiating absorbed energy. If a photon (e.g., traveling from the object to the sensor) encounters one of these particles, the photon may change direction with no loss of energy (elastic scattering), it may change direction and lose part of its energy to the particle (inelastic scattering and absorption, or it may be absorbed and lose all of its energy to the particle (absorption).

4.1 Absorption

Absorption usually results in a reduced signal, however, the effects can be difficult to predict. The energy exchange (or lack of exchange) in absorption is related to the frequency of the photon \( (Q = hf) \) and to the possible energy states of the absorbing particle (Figure 4.3). If a photon is absorbed, the energy of the absorbing particle will be raised to a higher energy state ("excitation"). In dropping from the excited state to a lower energy state, the particle releases (emits) a photon, whose energy (frequency) corresponds to the magnitude of the energy transition. If the transition from the excited state is to some interim energy state before dropping to the original state, the energy of the emitted photons will be less than the energy of the absorbed photon. Moreover, if the absorbed photon energy is sufficiently high, it may produce a free electron ("ionization"). The free electron could also absorb a photon, thereby going to a higher level of kinetic energy, or it could emit a photon and drop to a lower free energy.

Figure 4.1: Absorption. A photon interacts with a particle by imparting its energy to an electron which moves to a higher orbit. The electron eventually returns to the lower energy state, but may do so in one or more steps, emitting photons at the same or longer wavelengths than that of the absorbed photon.

Clearly, absorption is going to be very wavelength dependent and can often provide a characteristic signature for a particular substance. In a purely absorbing medium (no scattering), the change in irradiance can be described by Beer's Law:

\[
E = E_0 \exp(-ar)
\]

(4.1)

where \( E_0 \) is the initial irradiance, \( a \) is the absorption coefficient (with units of inverse length) and \( r \) is the distance between the points where \( E \) and \( E_0 \) are measured.

Absorption in the atmosphere is primarily due to oxygen (O\(_2\)), carbon dioxide (CO\(_2\)), ozone (O\(_3\)), and water (H\(_2\)O). Absorption by these molecules ranges from narrow, shallow absorption bands to largely opaque wavelength ranges in the EM spectrum. As an example,
consider the transmission spectrum of water, illustrated in the graphs below. Figure 4.2 shows the Atmospheric transmission in the wavelength range 0.9 – 2.7µm above Mauna Kea for air mass = 1.0 and water vapor column of 1.6mm. These data were computed using the ATRAN atmospheric model. Because of the altitude, the aerosol content is minimal and absorption is the primary mechanism affecting transmission over this wavelength range. There are a multitude of fine absorption lines that are not usually significant for typical earth remote sensing applications. However, the broader absorption bands due to water at 0.9µ, 1.15µ, 1.4µ, and 1.9µ are all apparent in spectral data. Data collected in the 1.4µ and 1.9µ bands are generally not usable for earth remote sensing applications.

Figure 4.2: Atmospheric transmission in the wavelength range 0.9 – 2.7µ above Mauna Kea for air mass = 1.0 and water vapor column of 1.6mm. Several of the major water absorption bands are labeled with the component responsible for the absorption. Note that, because of the altitude (~4200 m) there is little effect from scattering; the transmission is near 1.0 outside of the absorption bands.

Atmospheric absorption in the wavelength range 2.7-5.6 µ is shown in Figure 4.3. The atmosphere is essentially opaque from 2.5-3.0µ due to water vapor and CO2 absorption, and between 4.2-4.5µ, due to CO2, N2O and CO. On the other hand the atmosphere is largely transmissive in the 3.0-4.1µ and 4.6-5.3µ, in spite of the numerous fine absorption features in these regions.

1 Figure 4.2, Figure 4.3, Figure 4.4, and Figure 4.5 have been taken from the Gemini Observatory web site. site http://www.gemini.edu/sciops/ObsProcess/obsConstraints/ocTransSpectra.html#0.9-2.7um. The graphs were computed using the ATRAN atmospheric modeling program (Lord, S.D. 1992, NASA TM. 103957)
Figure 4.3: Atmospheric transmission in the wavelength range 2.7 – 5.6 μm above Mauna Kea for air mass = 1.0 and water vapour column of 1.6mm.

Figure 4.4: Atmospheric transmission in the wavelength range 8-14μm.

The atmosphere is again opaque from 5.0-8.0μ because of strong absorption by water vapor, but becomes highly transmissive between 8.0-9.5μ and 10.-13μ (Figure 4.4). Absorption from 9.5-10μ is predominantly by O₃. From 14μ~0.1mm the atmosphere becomes virtually opaque due to CO₂ and H₂O absorption. The only exception is between 16-28μ where the atmosphere becomes partially and intermittently transmissive. Transmission in this region is very sensitive to water vapor content and is not typically used for earth remote sensing.
Figure 4.5: Atmospheric transmission in the wavelength range 16-28μm.

Figure 4.6: Atmospheric transparency/absorption for the dominant molecules in the atmosphere from 1-16μm. Absorption ranges from zero (transparent – white in the figure) to 1 (opaque – colored dark in the figure).
4.2 Scattering

Scattering is the term used to describe any process by which radiation changes direction. Although the process can result in a change in the wavelength (inelastic scattering), most atmospheric scattering is elastic, i.e., there is no loss of energy in the process and the wavelength of the scattered radiation is identical to that of the incident radiation. Only the direction of propagation changes. Elastic scattering is essentially instantaneous, occurring on time scales of $\sim 10^{-14}$ seconds and is generally characterized as either Rayleigh or Mie scattering, depending on the size of the scattering particles relative to the wavelength.

**Figure 4.7:** Elastic scattering. The scattered radiation has the same energy, and therefore the same wavelength, as the incident radiation. Only the direction changes.

Rayleigh scattering results from interaction of radiation with particles which are much smaller than the scattered wavelength, $\lambda$, and the amount of scattering is proportional to $1/\lambda^4$. For visible radiation, Rayleigh scattering is predominantly due to molecules (N$_2$, O$_2$, etc.) and blue light is much more likely to be scattered than red light. Rayleigh scattering of sunlight is, of course, the origin of the blue sky on a clear day.

**Figure 4.8:** With Rayleigh scattering the scattered radiation is symmetric in the forward/backward direction as well as being symmetric about the incident direction. The scattered radiation is also polarized at $\theta=90^\circ$ from the incident direction and unpolarized in the forward and backward directions ($\theta=0^\circ$ and $\theta=180^\circ$).

Rayleigh scatter also has a unique scattering pattern (phase function) that is symmetric both about the direction of incident radiation and in the forward/backward direction. Defining the scattering angle, $\theta$, as the angle between the incident direction and the scattered direction (Figure 4.8), the flux density scattered in the direction $\theta$ is given by:

$$E = E_0 \left(1 - \cos^2 \theta \right) / \lambda^4$$  \hspace{1cm} (4.2)
This distribution arises from differences in polarization and Rayleigh's model which represents molecules as dipoles (positive and negative charge pairs). In effect, both parallel and perpendicular polarized radiation (relative to the axis of the dipole) will be scattered equally in the forward and backward direction, but only radiation polarized parallel to the axis of the dipole will be scattered at 90°. The effect can be observed on a clear day with the sun at a zenith angle between about 30° and 60°. The blue skylight will be least intense and most highly polarized at about 90° from the sun in the solar plane (e.g., the plane defined by the observer, the sun, and the local vertical). Because atmospheric scattering is not purely Rayleigh scattering, the location of the dark point is generally less than 90° from the sun and the polarization is far from complete, but the effect is still quite noticeable.

A second form of scattering that is important for understanding atmospheric transmission is Mie scattering, a form of scattering that is caused by particles that are as large as or larger than the scattered wavelength. When caused by particles whose diameters are approximately equal to the scattered wavelength, the amount of scattering is proportional to $1/\lambda^2$; if the particle diameters are equal to 1.5 times the scattered wavelength, the amount of scattering is proportional to $1/\lambda$; but if the particles are at least twice the scattered wavelength, the amount of scattering is independent of $\lambda$ (Figure 4.9). Mie scattering is sometimes referred to as "particle scattering" and is often treated in discussions as if there were no wavelength dependence at all. This is strictly true only for particles that are much larger than the wavelength of the incident radiation.

\[
\begin{align*}
\text{scattering} & \propto 1/\lambda^4 \text{ when } r \ll \lambda, \\
\text{(Rayleigh limit)} \\
\text{scattering} & \propto 1/\lambda^2 \text{ when } r \approx \lambda, \\
\text{scattering} & \propto 1/\lambda \text{ when } r \approx 3\lambda / 2, \\
\text{scattering independent of } \lambda \text{ when } r \gg \lambda.
\end{align*}
\]

**Figure 4.9:** Wavelength dependence of atmospheric scattering ranges from the $\lambda^{-4}$ dependence of molecular (Rayleigh) scattering to the largely wavelength independent scattering from particles that are large relative to the optical wavelength. The specific dependence will depend on the particle size distribution at a particular time and a particular site.

The Mie scattering phase function is much more complex and variable than that for Rayleigh scattering. For our purposes we may describe the distribution as being simply very strongly dominated by forward scattering (Figure 4.10) and largely independent of wavelength.
Figure 4.10: The Mie scattering phase function is very strongly dominated by forward scattering and, for particles much smaller than the wavelength, it is largely wavelength independent.

Scattering by large particles (Mie scattering) tends to dominate at longer wavelengths while molecular scattering (Rayleigh scattering) is more important at shorter wavelengths. The exact spectral distribution of scattered radiation depends on the aerosol particle size distribution. At this altitude (~12m) scattering is significant even above 0.9μ although it is essentially negligible at 2.5μ. Scattering is most important in the visible where it also exhibits the strongest wavelength dependence. Conversely, absorption is almost negligible in the blue but is extremely important over most of the rest of the spectrum except at selected wavelengths.

Figure 4.11: Atmospheric transmission spectrum derived from AVIRIS data over Pocomoke City, MD, 5 Jul 1996. The spectrum represents conditions for air mass = 1.09, and water vapor column of 1.2cm. The dashed line indicates losses due to scattering. The difference between the dashed line and the solid line represents attenuation due to absorption. Compare this

4.3 Dominant Processes and Phenomena

Ultraviolet region -- Passive sensors of ultraviolet radiation make use of reflected solar radiation, while active sensors (lasers or lidars) supply their own radiation. As shown in Figure 4.12, wavelengths shorter than 0.28 μm are attenuated strongly by the atmosphere. Oxygen absorbs strongly from 0.13 to 0.20 μm; ozone has a strong absorption band from 0.20 to 0.28 μm
and a weaker band from 0.32 to 0.35 µm. Attenuation by scattering in the ultraviolet region is dominantly Rayleigh scattering.

![Figure 4.12: Dominant atmospheric interactions by wavelength region.](image)

**Visible region** -- In the absence of clouds or other hydrometeors, the visible region is characterized by high transparency or windows. Rayleigh and Mie scattering by atmospheric gases (argon, oxygen and nitrogen) and particulates (aerosols) are the dominant radiative processes that affect both passive sensing of reflected solar radiation and active sensing with laser or lidar systems.

**Infrared region** -- Passive sensing in the 0.70 to 4 µm portion of the infrared region is predominately of reflected solar radiation during daylight hours, while passive sensing in the 4 to 22 µm portion is dominated by thermal emission. Molecular absorption becomes significant in the infrared, with strong gaseous absorption bands associated with water and carbon dioxide, and to a lesser extent, with nitrous oxide and ozone. Rayleigh scattering is generally negligible compared to Mie scattering, and windows are found at several spectral bands (e.g., 8 to 14 µm).

**Microwave region** -- Depending on the sensor, microwave radiation is sensed passively, collecting long wavelength thermal emissions (microwave radiometer), or actively, collecting reflected (re-radiated) energy which originated at the sensor (radar). The atmosphere is nearly opaque between 22 and 300 µm (0.03 cm), but transmission is high for wavelengths longer than about 3 cm (10 GHz). The spectral bands of high absorption are also characterized by radiant energy emission. Molecular oxygen and water vapor are the principal causes of gaseous absorption, with oxygen absorption occurring at 0.25 and 0.50 cm, and water absorption at 0.17 and 1.35 cm.

**4.4 Simple Radiative Transfer Model**

In order to get a better sense of the effect of the atmosphere on remote observations, it is very useful to set up a simple radiative transfer model of the process. This will make the atmospheric effects explicit and will provide a guide to the advantages and disadvantages of different observation strategies.
The sun is our ultimate light source. Outside the atmosphere, the solar irradiance can be taken as a constant to a good approximation, $E_{\text{sun}}$. The solar irradiance (solar constant) including photons of all wavelengths is approximately:

$$E_{\text{sun}} = 1367 \text{ W m}^{-2} \quad (4.3)$$

This value varies by only a fraction of a percent over time scales of minutes to decades.

Another handy attribute of the solar irradiance is that, since the sun is so far away, the sun's rays are virtually parallel and the spreading of those rays over the distance from the top of the atmosphere to the earth's surface is also negligible. If the atmosphere were not in the way, we would measure essentially the same irradiance at the earth's surface as at the top of the atmosphere. However, the atmosphere alters the solar radiation by absorption and scattering. This results in a decrease in the solar irradiance at the earth's surface and gives rise to the radiation from other directions (skylight).

First, consider the attenuation of the solar irradiance. Absorption by a beam of collimated radiation may be described by the beam attenuation coefficient, $a$, which has units of inverse length. The amount of radiation absorbed over a path length $r$ over which the absorption coefficient is invariant (or can be replaced with an average value) is then given by Beer's Law:

$$E = E_0 e^{-ar} \quad (4.4)$$

Similarly, the amount of radiation scattered out of the beam (in any direction), is described by the total scattering coefficient, $b$. If the effects of scattering and absorption are independent, then the total attenuation, $c$, is then simply the sum of the losses due to absorption and scattering:

$$c = a + b \quad (4.5)$$

and the beam attenuation is then given by:

$$E = E_0 e^{-cr} \quad (4.6)$$

The path over which the beam is attenuated by $1/e$ is 1 attenuation length and is equal to $1/c$, and the product, $cr$, describes the optical depth (dimensionless) of the medium.

We may now describe the transmission of the solar irradiance through the atmosphere. If $c_{\text{atm}}$ represents the average attenuation coefficient for the atmosphere\(^2\), then we may express the irradiance on a surface perpendicular to the direction of light from the sun as:

$$E_{\perp} = E_{\text{sun}} \exp\left(-c_{\text{sun}} z \sec \theta_{\text{sun}}\right) \quad (4.7)$$

For the atmosphere we are concerned with the total transmission and the specific altitude is not really an issue. In that case the product $c_{\text{sun}}z$ is often replaced with an equivalent term, $\tau$, representing the optical depth of the atmosphere in the vertical direction. [Note that, for solar zenith angles greater than $\theta_{\text{sun}} \sim 60^\circ$, the correction term of $\sec \theta_{\text{sun}}$ is too inaccurate and must be replaced with a more accurate formula that accounts for earth curvature.]

\(^2\) The atmosphere is clearly not homogeneous. Even in the relatively simple case of a plane-parallel atmosphere in which all of the variability is in the vertical dimension, the attenuation coefficient is a function of altitude. Thus, $c_{\text{atm}} = c(z)$ and the optical depth is given by:

$$\int_0^h c(z) \, dz$$
The downwelling irradiance on a horizontal surface is a combination of the solar irradiance projected on the surface and the sky irradiance, $E_{\text{sky}}$:

$$E_d = E_{\text{sun}} \cos \theta_{\text{sun}} \tau_{\text{sun}} + E_{\text{sky}}$$

(4.8)

where: $\tau_{\text{sun}} = \exp(-c z \sec(\theta_{\text{sun}}))$

If we know the BRDF, $R(\theta_{\text{sun}}, \varphi_{\text{sun}})$, of the surface, we may describe the radiance leaving the surface in the direction $(\theta, \varphi)$ as:

$$L_0(\theta, \varphi) = R(\theta_{\text{sun}}, \varphi_{\text{sun}}) E_d = R(\theta_{\text{sun}}, \varphi_{\text{sun}}) (E_{\text{sun}} \cos \theta_{\text{sun}} \tau_{\text{sun}} + E_{\text{sky}})$$

(4.9)

If we further assume that the surface is Lambertian, then we may replace the BRDF with a simple reflection factor, $\rho$, that is independent of the direction of observation $(\theta_v, \varphi_v)$, and write:

$$L_0(\theta_v, \varphi_v) = \rho (E_{\text{sun}} \cos \theta_{\text{sun}} \tau_{\text{sun}} + E_{\text{sky}}) \cos(\theta_v)$$

(4.10)

The $\cos \theta_v$ term is present to adjust for the effective source area perpendicular to the viewing direction. This is necessary since we wish to view a constant area regardless of the viewing angle. The projected area is smaller than the horizontal area by $\cos \theta_v$.

The reflected radiance propagates through the atmosphere to the detector with a transmission factor:

$$\tau_v = \exp(-c z \sec(\theta_v))$$

(4.11)

There is also a contribution to the radiance at the detector from sunlight that is scattered into the sensor's field of view. This radiation, called path radiance, $L*$, either never reaches the ground or, if it does, is scattered into the viewing direction from outside the sensor's field of view. It is called "path radiance" because it is produced along the viewing path. Finally, we have a full expression for the radiance at the detector:

$$L_0(\theta_v, \varphi_v) = \rho (E_{\text{sun}} \cos \theta_{\text{sun}} \tau_{\text{sun}} + E_{\text{sky}}) \cos(\theta_v) \tau_v + L*$$

(4.12)

The process is illustrated in Figure 4.13.
minimize the atmospheric path for viewing since that will minimize both the path radiance and the transmission loss. This is easily done by restricting the viewing angle to a small range of angles around nadir, e.g., $\theta_v \approx 0$. Another approach is to use spectral ratios and differences to optimize the signal. If the reflectance of a ground target varies significantly in two bands for which the atmospheric transmission and path radiance are effectively constant, then the subtracting one band from another will minimize the effect of the path radiance and enhance the ground target.

4.5 Atmospheric Correction (in development)

Atmospheric correction for remote sensing generally means solving for the reflectance, $\rho$, in (4.12). Accurate retrieval of spectral surface reflectance from hyperspectral and multispectral imagery is typically important when discriminating subtle differences among targets or when comparing data collected at different times. For example, water quality measurements must discriminate between very slight absolute reflectance variations. Thus it is important for these applications to accurately correct for the contributions of the atmosphere to light reaching the remote sensor.

Models used to correct for atmospheric effects must account for scattering and absorption of light within the atmosphere. Both are dependent on wavelength, and both vary over space and time as atmospheric characteristics change. Some atmospheric gasses (notably O$_3$, CH$_4$, and most significantly water vapor) are not well mixed in the atmosphere, and airborne particulates are likewise variable. Spectral absorption features of these gasses are well catalogued. Scattering distributions (Mie, Rayleigh) for gasses and particulates can be calculated when atmospheric constituents are known. Ideally, one would have access to concurrent measurements of the atmosphere within the scene. Unfortunately, precise measurements of the atmosphere between the surface and the sensor are rarely available. For images collected in the past, the user is restricted to whatever meteorological or other atmospheric data may have been collected roughly concurrent with the observation for points near the area of interest.

In order to surmount this problem, algorithms have been developed which infer gaseous and particulate constituents from observed absorption and scattering features within the hyperspectral or multispectral dataset itself. These serve to provide estimates of the atmospheric constituents to the radiative transfer models. The two most widely-used radiative transfer models at present are the Moderate Spectral Resolution Atmospheric Transmission algorithm (MODTRAN) and the Second Simulation of the Satellite Signal in the Solar Spectrum (6S) (Berk et al. 1999, Vermote et al. 1997). These basic models, together with the spectral analysis algorithms, are incorporated in a number of computer programs designed to evaluate the effects of the atmosphere pixel-by-pixel in multispectral and hyperspectral images to give accurate surface reflectance values, without recourse to detailed atmosphere profile measurements.

When a full atmospheric correction is impractical or unnecessary, it may still be necessary to adjust for atmospheric effects at least partially, and there are several techniques designed to accomplish this. Simplistically, (4.12) reduces the atmospheric correction to a simple, linear problem: a multiplicative factor that accounts for the transmission of radiation and the illumination at the surface, and an additive term that accounts for the path radiance. Ultimately the problem reduces to finding these two coefficients for each waveband in the remote system and, for situations in which the atmosphere can be regarded as uniform over the
area of interest there are several simple, approximate approaches for estimating or normalizing these coefficients. A few of these will be discussed briefly here.


**Dark Object Subtraction (in development)**

The additive term in (4.12) is arguably the most troublesome atmospheric contribution. In terms of retrieving the reflectance it is strictly additive noise. While it is particularly severe in the visible portion of the spectrum, path radiance contributes throughout the reflective range and can easily skew comparisons of the same target in images collected at different locations or times.

**Spectral Ratios (in development)**

**Empirical Line Method (in development)**