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Atmospheric Effects on Radiation Measurements

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Atmospheric Effects on
Radiation Measurements

by
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Purdue University, West Lafayette, Indiana

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ATMOSPHERIC EFFECTS ON RADIATION MEASUREMENTS

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ATMOSPHERIC EFFECTS ON RADIATION MEASUREMENTS

1. INTRODUCTION

Already during this course you have heard occasional references to the effect of the atmosphere upon radiation measurements obtained by remote sensing techniques. Usually the indication was that in some way the atmosphere acted to limit the quantity and quality of information available to the measuring system.

In certain respects this is a quite valid assessment of atmospheric processes. In fact, it is possible to envision the atmosphere as performing a dual role: 1) by acting as a generator it can introduce a spurious signal into the measured quantities, and 2) by acting as an unwanted filter the atmosphere can remove portions of the signal from the system. The effect of each of these actions is to disguise the information being sought and, as a result, techniques must be devised to remove such unwelcome effects.

Consider, for example, the photographer who places a filter in front of his normal lens system. His hope is to remove an excessive amount of scattered light brought about by hazy atmospheric conditions. This is an example of having to combat item 1 above. In addition, we might consider the effect of clouds upon the instrument system. Most certainly clouds act as a filter by reducing the overall level of the signal measured by the instrument. But more than that, clouds are not uniform in their radiative properties across the wavelength spectrum. Consequently, the presence of

clouds will alter the relationships between signals obtained in different spectral intervals. Until such time as information about the radiative properties of clouds -- and of various types of clouds -- can be included in a data analysis procedure, methods of remote identification of surface properties from spectral signatures are severely hampered. At the present time, aircraft measurement programs can avoid this problem by not flying on cloudy days -- an effective but unsatisfactory solution. Satellite programs simply accumulate large quantities of unusable data with this constraint.

We shall briefly consider the physical properties of the atmosphere which produce these effects. Firstly, let us develop a means of describing the transfer of radiative energy through a medium. Then, we shall apply this general treatment to cases of special interest to remote sensing technology.

2. BASIC CONCEPTS AND THE TRANSFER EQUATION

Essential to a discussion of the transfer of radiative energy are the concepts of solid angle and radiant intensity. In Figure 1 are presented both a pictorial representation and mathematical definition for solid angle, ω . We see that the differential solid angle, $d\omega$, is the projection of an area dA along the vector \hat{u} toward a point P , divided by the distance squared between dA and P . The integral of $d\omega$ over all space surrounding P yields a value of 4π . The concept of solid angle is of value when describing the directional properties of radiative energy flow.

Also, in Figure 1, radiant intensity, I , is defined as the radiant energy per unit time per unit area normal to the direction of propagation of energy per unit solid angle containing the radiant energy.

Figure 2 presents several steps in the development of the radiative transfer equation (Goody, 1964). This equation is a statement of the conservation of radiative energy, and can be obtained by considering a volume element $d\tau$. The volume element is illuminated by an intensity of radiation $I_\lambda(\omega, s)$ [the subscript λ indicates that the intensity may be wavelength dependent] within the solid angle $d\omega$. Upon emerging from the volume element

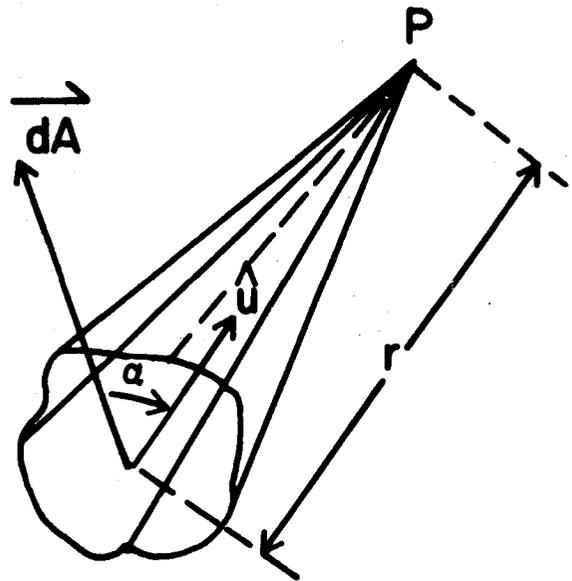
Basic Radiative Transfer Concepts

Solid Angle, ω

$$d\omega \equiv \frac{\vec{dA} \cdot \hat{u}}{r^2}$$

$$d\omega = \frac{dA \cos \alpha}{r^2},$$

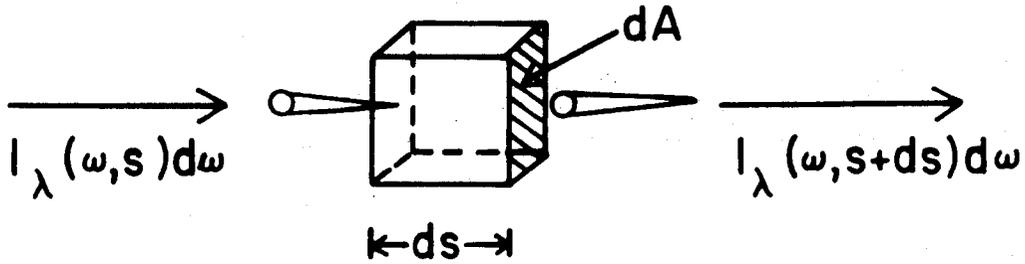
where $dA = |\vec{dA}|$



Radiant Intensity, I

The radiant energy per unit time per unit area normal to the direction of propagation of energy per unit solid angle containing the radiant energy.

Figure 1. Definitions of solid angle and radiant intensity.



The change in intensity I_λ within the volume ds dA and within the solid angle $d\omega$ is

$$\frac{dI_\lambda}{ds} dsdAd\omega = -k_{\lambda t} \rho I_\lambda ds dAd\omega$$

Attenuation of radiation

$$+ k_{\lambda t} \rho dsdAd\omega \int_0^{4\pi} p(\Theta) I'_\lambda d\omega'$$

Source of scattered radiation

$$+ E_\lambda \rho dsdAd\omega$$

Source of emitted radiation

Or,

$$\frac{1}{\rho k_{\lambda t}} \frac{dI_\lambda}{ds} = J_\lambda - I_\lambda$$

↳ Source function representing last 2 terms above.

Figure 2. Development of the radiative transfer equation.

the intensity has a new value $I_\lambda(\omega, s+ds)$. The rate of change with time of radiative energy within $dsdA$ and $d\omega$ is shown to result from several processes. Firstly, energy can be removed from the direction of propagation of I_λ through either absorption or scattering by the medium which has a density ρ . The sum of these two processes is called attenuation and is quantified by the mass attenuation coefficient $k_{\lambda t}$ having units of area per unit mass and representing the effective area intercepting the radiation field per unit mass of substance. Secondly, the scattering process can redirect energy into the direction of I_λ and thus serve as a source of energy. The integral of incident intensities [indicated as primed quantities] over all solid angles accounts for the total source of scattered energy. The incident intensities are weighted with the scattering phase function, $p(\theta)$, which gives the probability that an incident intensity I'_λ will be redirected by an angle θ , the scattering angle, into the direction of I_λ . Finally, the medium within $dsdA$ can emit energy into the direction of I_λ . This is represented by the quantity E_λ , the rate of emission of energy per unit time per unit mass. After cancelling common factors we obtain the last expression in Figure 2, a primitive form of the radiative transfer equation in which J_λ , the source function, replaces the two terms for scattering and emission sources for I_λ .

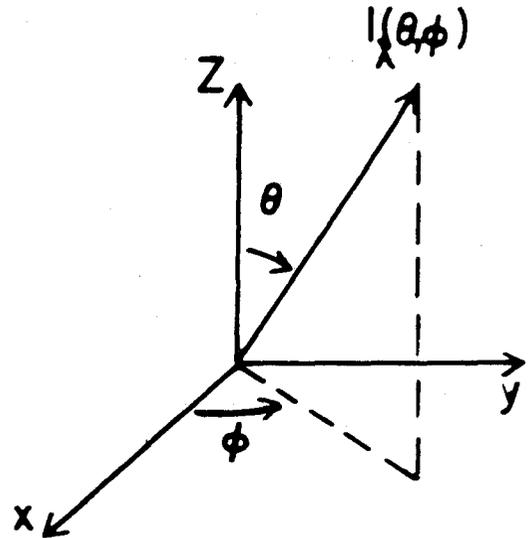
The form of the transfer equation can be modified using the z, θ, ϕ coordinate system shown in Figure 3. An often-used convention replaces the θ coordinate with $\mu \equiv \cos \theta$ and the z coordinate with optical thickness τ_λ , defined by $d\tau_\lambda \equiv -k_{\lambda t} \rho dz$. Optical thickness is zero at the top of the atmosphere and increases as height decreases to a total value $\tau_{\lambda T}$ at the earth surface. If we assume that the atmosphere is horizontally homogeneous and that time rates of change of I_λ are small compared to vertical changes, the transfer equation takes on the form shown.

Many remote sensing efforts involve the measurement of upward traveling radiation. Therefore, we present in Figure 3 the formal solution to the transfer equation for an upward-directed intensity, $I(\tau, \mu, \phi)$, at a level in the atmosphere where the optical thickness is τ and traveling in a direction defined by μ, ϕ . Note that the λ subscript has been deleted for convenience.

Define a coordinate system :

$$\text{Let } \mu \equiv |\cos \theta|$$

$$d\tau_\lambda = -k_\lambda \rho dz$$



Assume :

- 1) Horizontal homogeneity
- 2) Time rate of change of I_λ is small compared to space rate of change.

$$\boxed{\frac{-\mu dI_\lambda(\theta, \phi)}{d\tau_\lambda} = J_\lambda(\theta, \phi) - I_\lambda(\theta, \phi)}$$

The solution for an upward-travelling intensity (dropping λ subscript for simplicity),

$$-(\tau_T - \tau) / \mu$$

$$I(\tau, \mu, \phi) = I(\tau_T, \mu, \phi) e$$

$$+ \int_{\tau'=\tau}^{\tau_T} e^{-(\tau'-\tau)/\mu} J(\tau', \mu, \phi) \frac{d\tau'}{\mu}$$

Figure 3. The solution of the radiative transfer equation for an upward-traveling intensity.

If one recalls that the source function contains an integration over intensity, it is clear that the practical solution of the transfer equation entails an iterative numerical scheme of some type. The expression for $I(\tau, \mu, \phi)$ contains two terms. The first represents the intensity at the surface, $I(\tau_T, \mu, \phi)$, attenuated over the slant path optical thickness, $\tau_T - \tau/\mu$, between the surface and level τ . The second term represents the contribution to $I(\tau, \mu, \phi)$ from within the atmosphere. The source function $J(\tau', \mu, \phi)$ at every level below τ is attenuated by the appropriate optical thickness and the contributions from all levels are summed.

3. THE PURE SCATTERING CASE

Figure 4 (Fleagle and Businger, 1963) demonstrates a situation which allows considerable simplification of the relationships governing the transfer of radiant energy. There are shown curves representative of the radiative emission of the sun and of the earth-atmosphere system. These can be given in terms of the emission of black, or perfectly emitting, bodies at temperatures of 6000°K and 250°K , respectively. Of importance to note is the fact that these two emissions fall into distinctly different spectral intervals, usually referred to as short-wave and long-wave radiation. In view of the fact that scattering processes are usually of importance for wavelengths less than 1 or 2 micrometers, we see that only the solar or short-wave radiation need be considered since the emissions of the earth and atmosphere are negligibly small at these wavelengths. Conversely, when considering the emissions of the earth and atmosphere at wavelengths greater than a few micrometers, the scattering process can be neglected and solar radiation is of no importance. Thus, consideration of radiative energy transfer can be conveniently divided into two classes: 1) the scattering of solar or short-wave radiation and 2) the absorption and emission of long-wave earth and atmospheric radiation.

Let us consider first the situation in which the long-wave earth-atmosphere emissions can be neglected and only the scattering of solar radiation is of importance. Solar radiation is usually considered to originate from a single direction, and it is convenient to treat it separately from

BLACK-BODY EMISSION CURVES FOR
TEMPERATURES OF 6000°K A 250°K,
NORMALIZED TO EQUAL AREAS

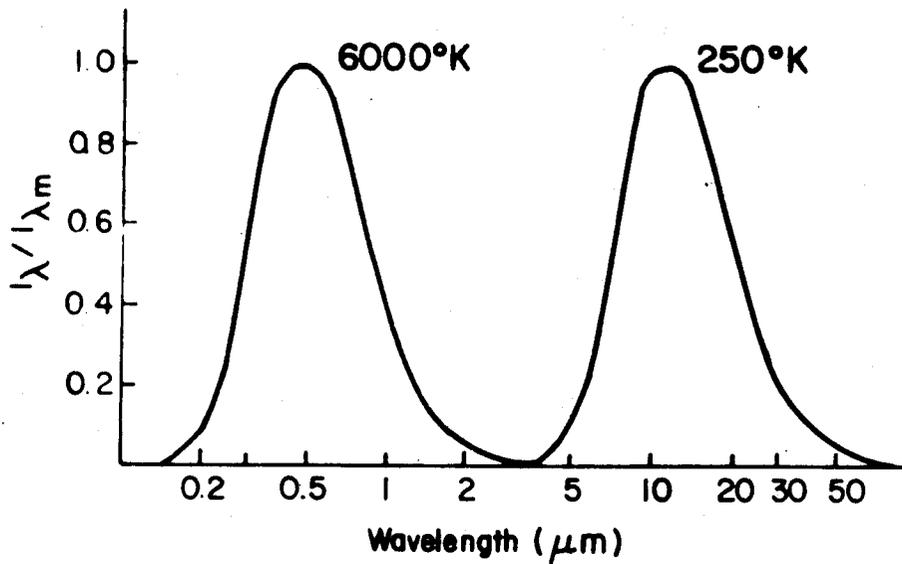


Figure 4. Black-body emission curves for temperatures of 6000°K and 250°K, normalized to contain equal radiative energies.

intensities of scattered radiation which can travel in any direction. In Figure 5 the source function J for a pure scattering case is given, broken into two parts to treat the solar flux, J_F , and other intensities, J_I , separately. The solar flux is defined to travel in the direction $-\mu_0, \phi_0$, where a negative sign is used to indicate downward propagating energy. The solar flux source term is the product of the solar flux and the scattering phase function for the direction μ, ϕ . The source function for other intensities is an integral over all possible incident intensities weighted by the source functions for scattering into the μ, ϕ direction.

An often-considered case is one when $J = 0$. The intensity $I(\tau, \mu, \phi)$ then consists only of the attenuated surface intensity, a relationship known as Beer's Law. Applying this assumption to the solar flux which is incident at the top of the atmosphere, travels to the surface and is reflected back to a sensor, we obtain the expression for $I(\tau, \mu, \phi)$ in Figure 6. We see that the surface reflectivity, r , the sought-after parameter in short-wave remote sensing efforts, can be directly obtained if we have knowledge of the atmospheric radiative properties.

A reasonable representation of the atmosphere might include the following components: molecular scattering, aerosol attenuation, ozone absorption, and cloud attenuation. These might well be specified in terms of the optical thicknesses corresponding to those processes. Considerable effort has been devoted to the determination of these properties. Molecular scattering and ozone absorption are very constant with time and can be described once and for all. But aerosol and cloud effects are highly variable and demand more attention. Indicated in Figure 7 are the total atmospheric optical thicknesses (Elterman, 1970) associated with molecular scattering, ozone absorption, and aerosol attenuation (the last for meteorological ranges, or visibilities, or 2 and 10 km). Ozone absorption is very strong in the ultraviolet spectrum, but becomes quite negligible for $\lambda > 0.4 \mu\text{m}$. Molecular scattering is rather strong in the blue end of the visible spectrum and decreases quickly (according to the familiar λ^{-4} law) with increasing wavelength. Two curves are shown for the aerosol to indicate the variability of this factor and also its dominance in the near infrared spectrum. No indication of cloud

Scattering only; parallel beam solar radiation incident at top of atmosphere, $F(0, -\mu_0, \phi_0)$.

Because parallel beam radiation cannot be treated as an intensity, let

$$\begin{aligned}
 J(\tau, \mu, \phi) &= J_F(\tau, \mu, \phi) + J_s(\tau, \mu, \phi) + \frac{E(\tau, \mu, \phi)}{k \lambda t} \\
 &= \rho(\mu, \phi, -\mu_0, \phi_0) F(\tau, \mu_0, \phi_0) \\
 &\quad + \int_0^{4\pi} \rho(\mu, \phi, \mu', \phi') I(\tau, \mu', \phi') d\omega'
 \end{aligned}$$

No source term: $J(\tau, \mu, \phi) = 0$

$$I(\tau, \mu, \phi) = I(\tau_T, \mu, \phi) e^{-(\tau_T - \tau)/\mu} \quad \text{Beer's Law}$$

A similar relation applies for the attenuation of downward traveling solar radiation.

Thus,

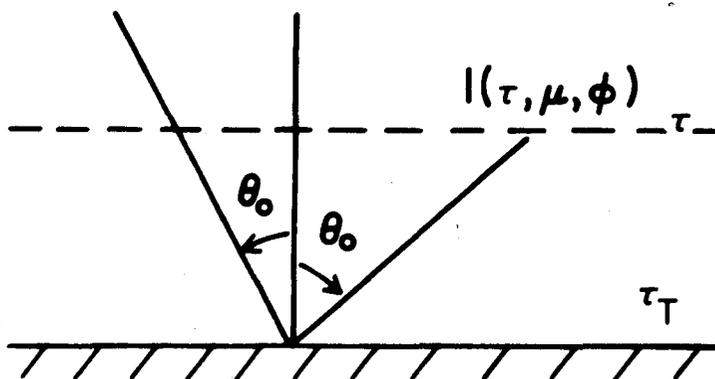


Figure 5. The source function for a pure scattering case, and the statement of Beer's Law.

$$F(\tau, -\mu_0, \phi_0) = F(0, -\mu_0, \phi_0) e^{-\tau/\mu_0}$$

So,

$$I(\tau_T, \mu, \phi) = F(\tau_T, -\mu_0, \phi_0) r(\mu, \phi, -\mu_0, \phi_0)$$

$$= F(0, -\mu_0, \phi_0) e^{-\tau_T/\mu_0} r(\mu, \phi, -\mu_0, \phi_0)$$

bi-directional reflectance

Thus,

$$I(\tau, \mu, \phi) = F(0, -\mu_0, \phi_0) e^{-\tau/\mu_0} r(\mu, \phi, -\mu_0, \phi_0) e^{-(\tau_T - \tau)/\mu}$$

Figure 6. The relation between measured upward-directed intensity and surface reflectivity for a zero source function case.

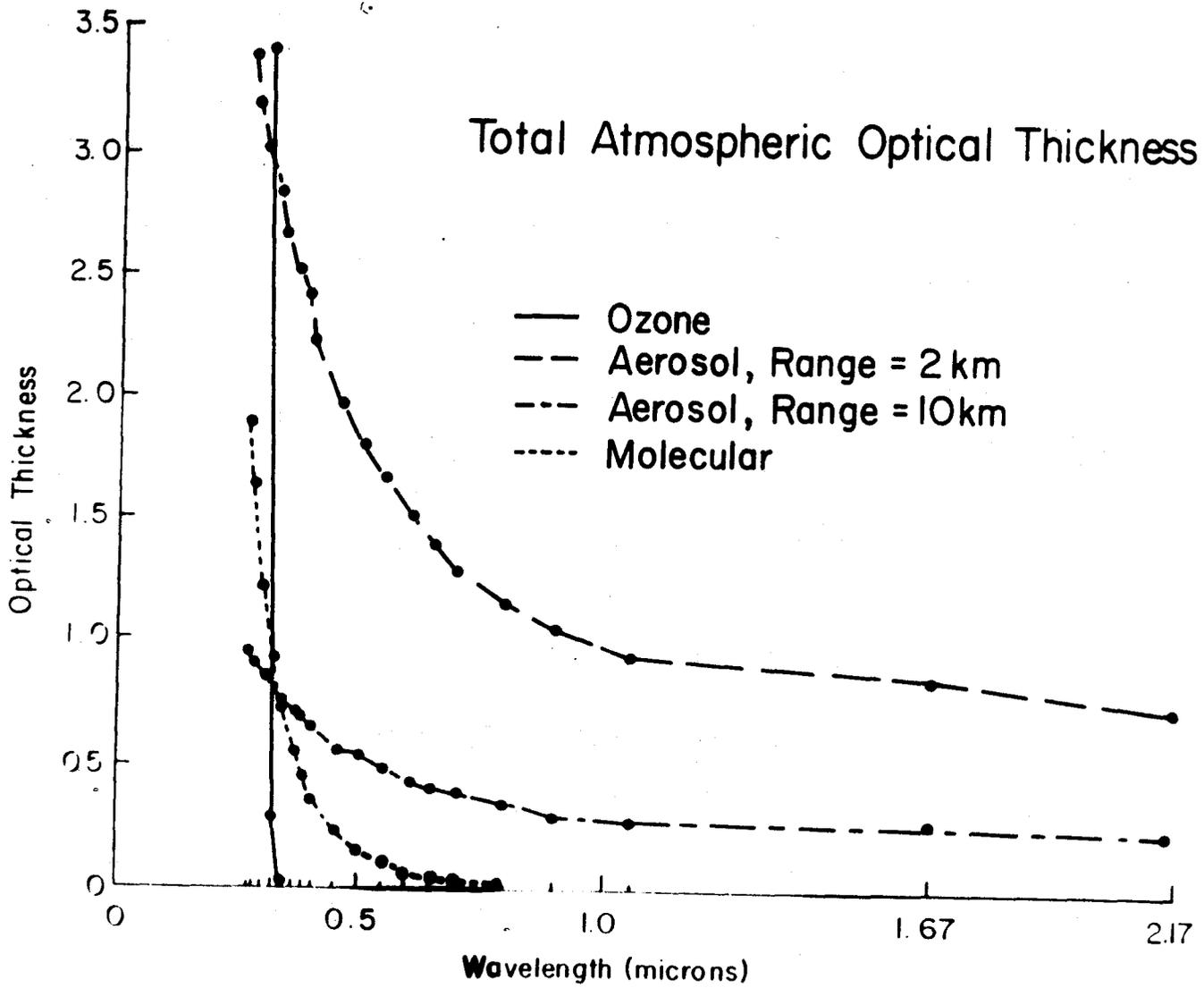


Figure 7. Total optical thicknesses of the molecular, ozone, and aerosol (for both 2 km and 10 km meteorological ranges) components of the atmosphere.

attenuation is shown because it is not well-known at this time. Determination of the spectral radiative properties of both aerosols and clouds constitutes a major area of research in the atmospheric sciences today.

The assumption of a zero source function is convenient, as we have seen, but not very realistic for most situations. The radiative energy removed from the incident solar flux will reappear in other directions and provide energy sources there. Figure 8 is a schematic attempt to represent the scattering process in the atmosphere. We see that the radiation incident upon the surface is now composed of both a direct component (as before) and a component of the scattered radiation from the atmosphere, usually called the diffuse component or sky radiation.

Thus, the surface irradiance is greater than before, and also has an altered spectral character. For example the λ^{-4} wavelength dependence in molecular scattering causes blue wavelengths to be scattered more than the red. The radiation scattered by aerosols and cloud droplets also bears a spectral identity which, in general, differs from that of the molecular component. The contributions to the total surface irradiance of several such components are presented in Figure 9 (Robinson, 1966). Of importance to note are the relatively large additions to the direct solar component made by molecular scattering at short wavelengths and by cloud droplet scattering at the longer wavelengths. As a result the radiation leaving the earth's surface will be representative not only of the spectral properties of the surface, but also of the spectral properties contributed by the atmosphere. The extraction of information concerning the surface now requires both knowledge of and accounting for these effects. In Figure 10 are given relationships which indicate the added complexity of extracting surface reflectivities from measured intensities under these more realistic assumptions.

In addition to the variations with wavelength which are introduced by scattering within the atmosphere, the size and composition of the scattering particles determine variations with angle of the scattered radiation. This is represented by the scattering phase function $p(\mu, \phi, \mu', \phi')$ or $p(\theta)$, where θ is the scattering angle. Several different forms for the

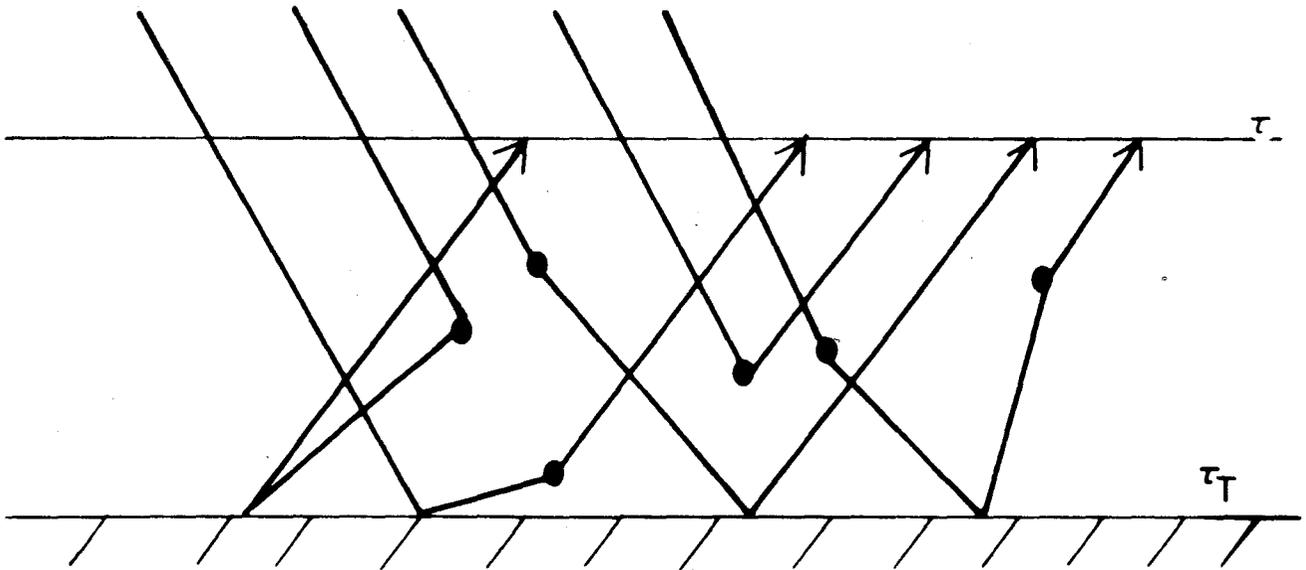


Figure 8. Pictorial representation of various single scattering processes in the atmosphere.

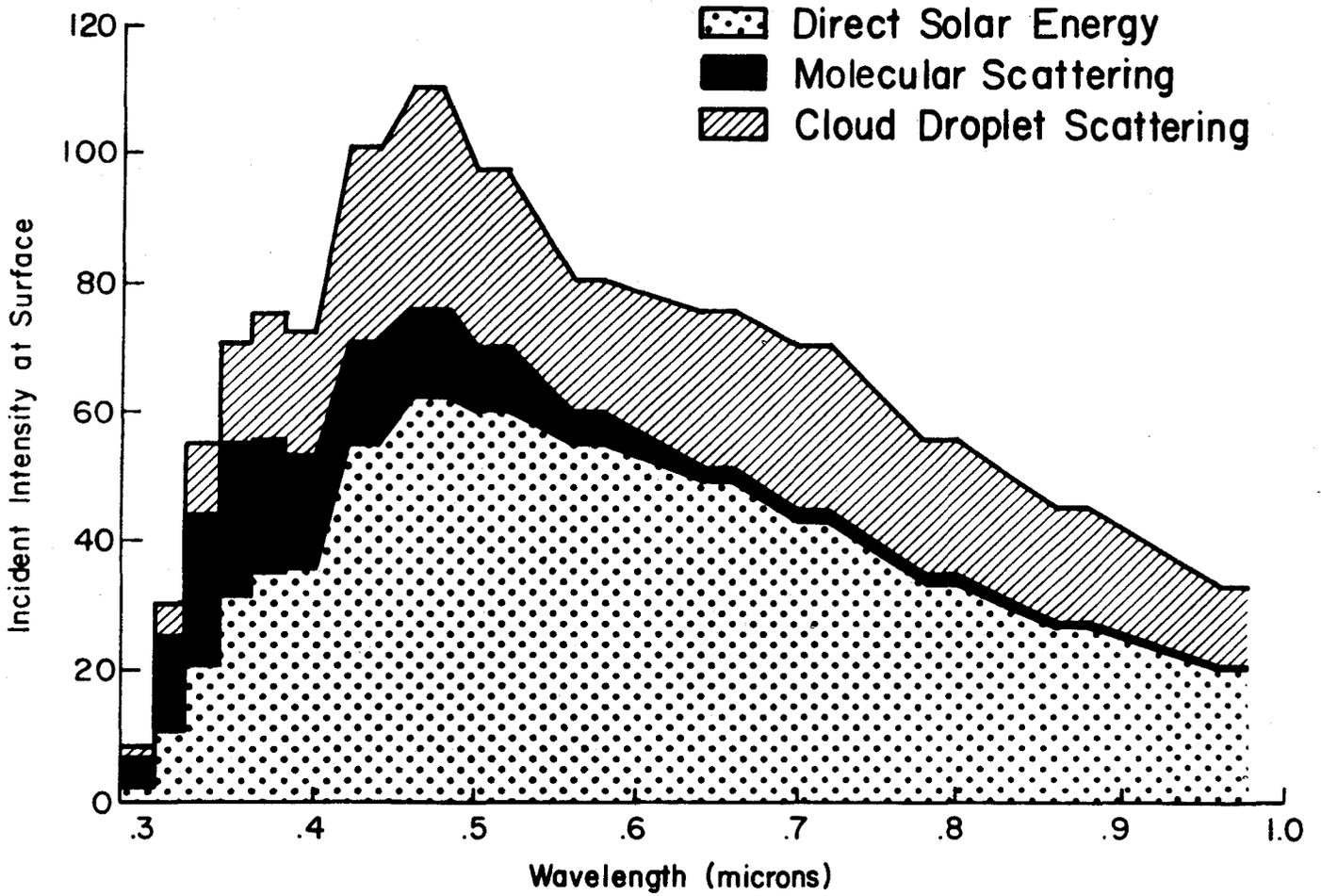


Figure 9. Contributions to the radiation (measured in relative units) incident on the earth's surface made by direct solar radiation, scattering by molecules, and scattering by a typical cloud.

$$I(\tau_T, \mu, \phi) = \underbrace{F(0, -\mu_0, \phi_0) e^{-\tau_T / \mu_0} r(\mu, \phi, -\mu_0, \phi_0)}_{\text{Direct component}}$$

$$+ \underbrace{\int_{\text{upper hemisphere}} I(\tau_T, \mu', \phi') r(\mu, \phi, -\mu', \phi') d\omega'}_{\text{Diffuse component}}$$

$$I(\tau, \mu, \phi) = \underbrace{I(\tau_T, \mu, \phi) e^{-(\tau_T - \tau) / \mu}}_{\text{Attenuated radiation from source}}$$

$$+ \underbrace{\int_{\tau'=\tau}^{\tau_T} e^{-(\tau' - \tau) / \mu} J(\tau', \mu, \phi) \frac{d\tau'}{\mu}}_{\text{Contribution from within atmosphere}}$$

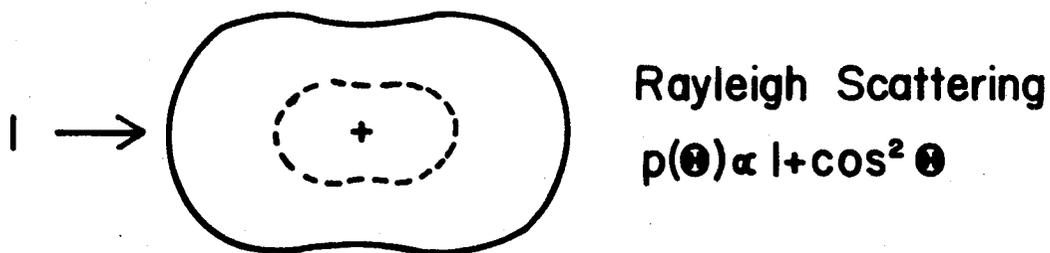
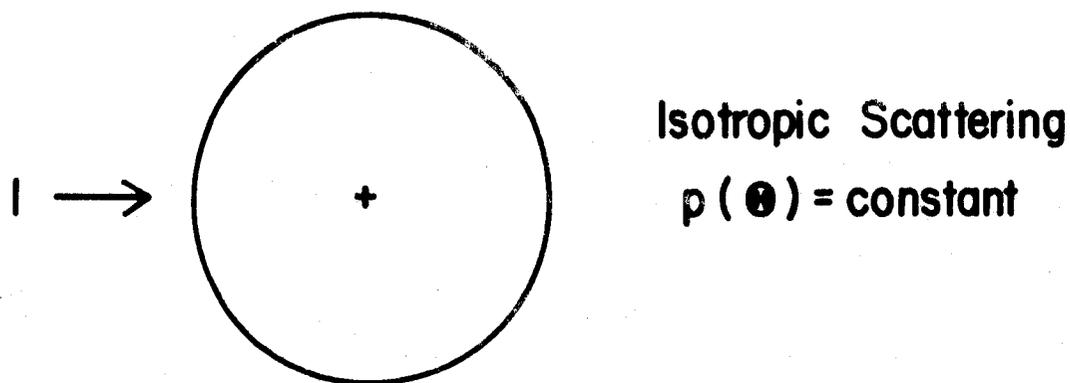
$$J(\tau', \mu, \phi) = \rho(\mu, \phi, -\mu_0, \phi_0) F(\tau', \mu_0, \phi_0) + \int_{\text{all } \omega} \rho(\mu, \phi, \mu', \phi') I(\tau', \mu', \phi') d\omega'$$

Figure 10. Expressions for the radiative intensity leaving the earth's surface and the upward-directed intensity within the atmosphere including a source function for atmospheric scattering.

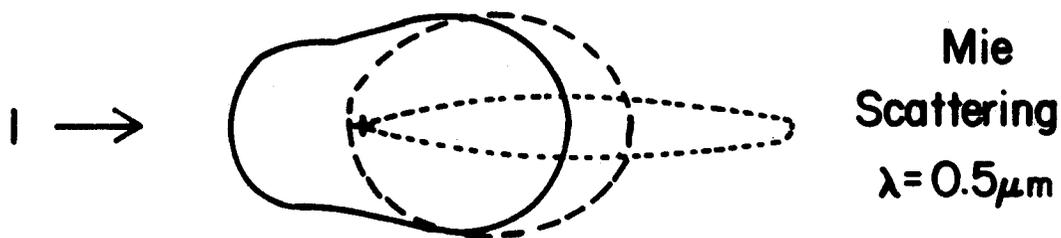
scattering phase function are shown in Figure 11 (Robinson, 1966). In each case the incident intensity arrives from the left, scattering occurs at the point marked +, and the curve is a locus of the relative magnitudes of scattered intensities for all θ . The simplest type of scatterer to envision is the isotropic scatterer for which $p(\theta)$ is constant. The surface of the earth is often modelled as an isotropic scatterer. The angular dependence of molecular, or Rayleigh, scattering is shown next, where $p(\theta) = 1 + \cos^2 \theta$. Two curves are given to point out the relatively greater scattering of short wavelengths. The green light is scattered more than the red, but in each case the intensities scattered in the forward and backward directions equal twice those scattered at right angles to the incident beam. As particle size increases greater and greater asymmetry appears, with the strongest scattering taking place in the forward direction. The scattering phase function no longer possesses a uniform simple form as seen above, but varies with particle size and composition. As a result, the signal measured at an aircraft, for example, will depend upon the angle of observation of the surface. The determination of surface reflective properties is clearly complicated by the presence of the atmosphere. It is the goal of our efforts to develop operational models which can account for these atmospheric effects.

4. NON-SCATTERING CASES

For wavelengths in the infrared spectrum the magnitude of scattered radiation can be considered negligible with respect to emissions by the surface and atmosphere. Thus, the source function represents emission alone, and can be shown to be equal to the black-body, or Planck, function $I_{\lambda b}$. See Figure 12. The solution of the transfer equation for $I(\tau, \mu, \phi)$ is also given in Figure 12, where it is seen that the emitted surface intensity is given by the product of the surface emissivity ϵ and the Planck function $I_{\lambda b}(T(\tau_T))$ for the surface temperature. At these wavelengths the critical surface parameter is the emissivity. This quantity can be extracted from a measurement given knowledge of the vertical temperature distribution and gaseous absorption and emission in the atmosphere, which



— $\lambda = 0.5 \mu\text{m}$ (green)
- - - $\lambda = 0.7 \mu\text{m}$ (red)



— $r = 0.05 \mu\text{m}$
- - - $r = 0.10 \mu\text{m}$
- · - · - $r = 0.50 \mu\text{m}$

Figure 11. Scattering phase functions for an isotropic scatterer, a very small particle obeying the Rayleigh scattering law, and larger particles obeying the more general Mie scattering theory.

Non-scattering Cases

For wavelengths in the infrared region scattering can be considered negligible compared to the emissive properties of the surface and atmosphere. Thus, the source function now represents emission alone. Through application of basic principles such as Kirchhoff's law, one can show that as long as thermodynamic equilibrium exists.

$$J_{\lambda}(\tau', \mu, \phi) = \frac{E_{\lambda}(\tau', \mu, \phi)}{k_{\lambda t}} = I_{\lambda b}(\tau')$$

The transfer equation now becomes

$$I(\tau, \mu, \phi) = I(\tau_T, \mu, \phi) e^{-\frac{(\tau_T - \tau)}{\mu}} + \int_{\tau' = \tau}^{\tau_T} e^{-\frac{(\tau' - \tau)}{\mu}} I_{\lambda b}(\tau') \frac{d\tau'}{\mu}$$

(0 < μ ≤ 1)

where $I(\tau_T, \mu, \phi) = \epsilon I_{\lambda b}(\tau_T)$
 and ϵ = emissivity of the target

Figure 12. The source function and solution to the radiative transfer equation for a non-scattering atmosphere.

allows evaluation of the source term. The most important gases in this wavelength region are water vapor, carbon dioxide, and ozone. Absorption spectra of these and several other gases are presented in Figure 13 (Fleagle and Businger, 1963), along with a composite atmospheric absorption spectrum. Beyond the 6.3 μm bands of H_2O there is an almost transparent region, called the atmospheric window, which lies between 8 and 12 μm . In the region between 13 and 100 μm lie the nearly completely absorbing rotational bands of CO_2 and H_2O . The window region is commonly used for surface observations because of its near-complete transparency.

5. SUMMARY

We have briefly discussed two essentially distinct regions of the electromagnetic spectrum: 1) the scattering region in which the radiation energy is provided by the incident solar flux, and 2) the infrared region in which emission by the earth's surface and atmospheric gases supply radiative energy.

However, in each of these spectral regions the atmosphere performs its dual function with respect to a remote sensing measurement of surface properties. The atmosphere acts both as a filter and as a noise generator removing and obscuring sought after information. Nevertheless, with proper application of concepts such as we have just considered, it will be possible to remove these unwanted atmospheric effects and to improve identification techniques being developed.

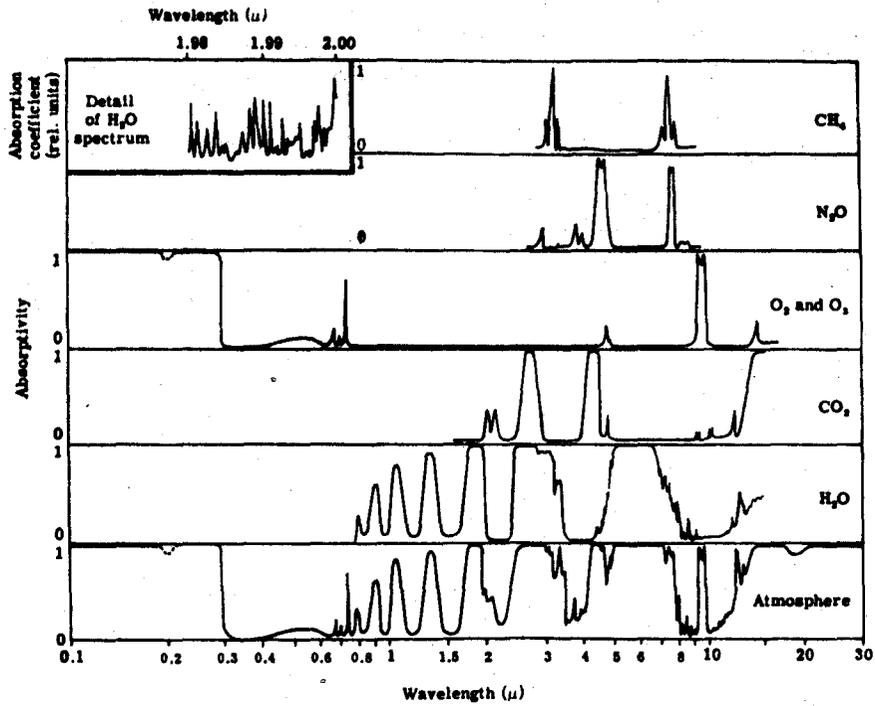


Figure 13. Absorption spectra for CH₄, N₂O, O₂, O₃, CO₂, H₂O and the absorption spectrum for the atmosphere.

REFERENCES

- Elterman, L., 1970: Vertical-attenuation model with eight surface meteorological ranges 2 to 13 kilometers. Air Force Cambridge Research Laboratories, Environmental Research Papers No. 318, Bedford, Mass., 56 pp.
- Fleagle, R. G., and J. A. Businger, 1963: An Introduction to Atmospheric Physics. Academic Press, New York, 346 pp.
- Goody, R. M., 1964: Atmospheric Radiation, Clarendon Press, Oxford, 436 pp.
- Robinson, N., 1966: Solar Radiation. Elsevier Publishing Co., Amsterdam, 347pp.