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**Research Article**

# Remote Sensing and Atmosphere

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**Abstract:** The intensities observed along nadiar at the top of atmosphere as a function of solar zenith angle for $\lambda = 0.55$ micron, haze o refractive index m = 1.50 – 0.031 and aerosols distributed over 0.03 to 10 micron range. As the solar zenith angle increases, the increases in effective atmosphericpath leads to decrease in intensity – approaching to zero a solar zenith of 90. The rate of decrease of intensity with solar zenith angle is more for higher values of reflectivity. The variation with the solar zenith angle at the top of the atmosphere of upward – travelling radiance for each of the lands at bands as seen at an altitude of 45.538 for a surface reflectivity of 0.2. this uses an atmospheric model based on the vertical distribution and content of ozone, aerosol and water vapour for an average mid-latitude summer. Atmosphere.since the solar flux is highest in the spectral interval 0.5-0.6 micron, the upward radiance received by that hand is higher than any other band. Also as the solar zenith angle increases, the upward radiance diminishes as expected because of the added path length through which the solar flux must pass.

**Keywords:** Remote Sensing, Atmosphere Absorption Band, Atmospheric Windows Aerosol

## 1. Introduction

Remote sensing of earths surface from any space based plateform involves the effect of earths atmosphere as the reflected / emitted spectral energy from a ground pixel has to pass through the atmospheric gases, and suspended aerosols environment, the major constituents of atmosphere (n2 o2 etc …) are more or less transparent to visible, near infra-red and thermal infra – red spectral regions (o2 has obsorption in microwave region of the spectrum).

Water vapour, ozone, oxides of nitrogen, co2 are the major gases. absorbers in the thermal IR and near IR contribute little to absorption invisible region.

Aerosols mainly effect the remote sensing in the visible region.

Fig. 1 pictorially expresses the atmospheric modes of modifying. the signatures received from the target at the surface of the earth. the radiant flux from the sun is partially absorbed and scattered as it passes down through the atmosphere to the surface of the earth.

So the surface b on the diagram is irradiated by the direct radiant flux from the sun as well as by the scattered flux from the surrounding hemisphere of sky (an example is the flux from g).

The ground scene reflects part of the flux incident on it in the direction of orbital sensor.

This reflected flux passing through the atmosphere is again absorbed and scattered as shown at f, but to it is added scattered light from the atmosphere, shown at d, that has not been reflected from the ground scene. Also added to the flux reflected from b is the flux reflected from a that is subsequently scattered at into the field of view of the remote sensing system.

These upward – scattering effects are the most insidious effects of the atmosphere, as the flux appears to remote sensor as if it came from the ground scene of interest.

In fact, it contains no spatial or spectral modulation from the scene.

In fact it radueas the contrast of the scene and thus make – fine detail invisible or at least harder, further more, in adding a uniform flux level to that from the ground scane, it confuses the interpretation of the spectral signature of the scene.

## 2. Atmospheric Windows and Absorption Bands

Spectral distridution of radiation $_{emited}$ from a perfect black body is given by,

$$m\lambda = \frac{2\pi hc2}{\lambda[\exp(chl\ \ kt)-1]}$$

### 2.1. Where

Mλ= Spectral radiant exitance in watt $m^{-2}um^{-1}$
H = planks constant ($6.6256\times10^{-34}ws^{2}$)
C = velocity of light ($2.997925\times10^{8}ms^{-1}$)
K = Boltzmann's constant ($1.38054\times10^{-23}wsk^{-1}$)
T = Absolute temperature in degree (k)
λ= wavelength in meters.
With constant values incorporated, mλ is given by

$$M \quad = \frac{3.74151\ \times 108}{5[\{\exp(1.34879\ \times\ 104)t\} - 1]}$$

Here λ is in micrometers.

### 2.2. Here the Quantities $3.74151*10^{8}wm^{-2}\mu m^{4}$ and $1.43879*10^{4}$

µm k are referred to as first and second radiation constants and are usually assigned the symbol c1and c2. the estimated standard deviation errors in c1 and c2 are 0.0027 and 0.0042 respectively.

Radiometric computation become simple if we approximate the solar irradiance distribution outside the earth's atmosphere as that due to a disk at a certain black body temperature in place of the sun.

Table – 1 is showing the wave 1ength interval and the corresponding temperature of the equivalent black body source to replace the photosphere.

Peek in the energy distribution curve of received in sun radiation at the top of the atmosphere corresponds to a temperature of 5950 k.

***Table 1.*** *Effective blackbody temperature of the sun as a function of wavelength.*

| Wavelength or wavelength interval in micron | Temperature of effective black body source to replace sun photosphere in |
|---|---|
| 0.2-0.25 | 5000 |
| 0.30 | 5500 |
| 0.35-0.40 | 5700 |
| 0.45 | 5900 |
| 0.48 | 5950 |
| 0.50 | 5900 |
| 0.55-0.60 | 5800 |
| 0.65-0.90 | 5700 |
| 0.95-1.20 | 5800 |
| 1.30-4.50 | 6000 |
| 5.00 | 5500 |
| 6.00-20.00 | 5000 |

The spectral region below 2.5 micron is termed as solar reflected flux region while the spectral region above 6 micron is termed as self – emitted thermal radiantflux

The special region between 2,5 to 6,0 micron contains a mixture of both radiation processes.

Atmosphere is a strong scatterer in the 300 – 400 micron region, generally reducing image contrast to an unacceptable level, and for this reason remote sensing below 400 nm is seldomat tempted.

Atmosphere transmission visible region, the losses there are due primarily to molecular aerosol solthering hardly molecular absorption occurs in a naturally clear, unpolluted atmosphere.

Absorption is a thermo – dynamically irreversible transformation of radcuent energy into heat.

Absorption is a thermo - dynamically scattering precess depends on the size distribution of scattering elements, their composition and concentration, and the wavelength or wavelength distribution of the radiant flux included on them Table 2 Gives the principal atmospheric windows on the electromagnetic spectrum.

***Table 2.*** *Prnicipalatmospheric windows in electromagnetic spectrum.*

| Window type | Wavelength (micron) | Frecuancy (gh2) | Absor Lower boundary | Bing gas Upper boundary |
|---|---|---|---|---|
| 1 | 2 | 3 | 4 | 5 |
| Visibal | 0.3-0.9 | - | O3 | H2o |
| Near ir | 1.5-1.6 | - | H2o | H2o |
| Near ir | 2.0-2.3 | - | H2o | H2o |
| Intermediateir | 3.5-4.5 | - | H2o | H2o |
| Far ir | 8.0-9.0 | - | H2o | O3 |
| Farir | 10.0-12.5 | - | O3 | Co2 |
| Microwave | $0.03\text{-}0.3^{6}$ | 100-80 | H2o | O2 |
| Micro wave | 0.7 – 1.0 | 45.30 | O2 | H2o |
| Micromove | 2.0- 30 | 15 to <1 | H2o- | - |

Absorption by atmosphere in the absortion bands will take place throught the atmosphere and for computation sake it could be divided into varios layer assuming meen values for absorption parameters in the layer.

The absorption will depend on the temperature, gas concentration or density and the spectral wavelength.

If the gas participating in absorption process is thoroughly mixed and is having uniform distribution (for example $co^{2}$, $o^{2}$ etc…..) then the measurement on a given wavelength band can give information on the temperature at various levels of the atmosphere.

If the participating gas (in absorption process is not uniformly distributed then with the knowledge of temperature profile the spectral information in an absorption band could be related to the concentration distribution of the gas in the atmosphere (example are the estimation of water vapour distribution in troposphere, estimation of stratosphere pollution concentration which are also known as the minor constituents of the atmosphere, totalozone measurement, ozone profile measurement using multispectral information etc….). Let us now look at conceptual aspect of absorption bands utilization which is being exploited in the atmosphere sciences studies using satellites.

The energy, reaching the sensor on board the satellite i (r, β), could be approximated by.

$$I\ (v.e) = b\ [v,t^{(no)}]\ t\ (v,no,\beta) - \frac{B[v,t(n)]dt(v,n,e)}{dn}$$

Where isa single valued function of pressure p and refers to surface pressure, B[v, t (N)] refers to plank black body function.

$$\frac{dt(v.n.\beta)}{dn}$$

Refers to the variance of transmittance function (averaged over the slit/filter spectral interval of the detector) with pressure.

Refer to spectral frequency, a single valued function of pressure and viewing geometry angle respectively.

The first turn of the r.h. s. of above equation refers to lower boundary (surface) contribution while the second termrefers to contribution of the atmosphere.

During the process of emission and absorption taking place at all leves, the emission spectral lines at the lower level (in troposphere) will be broader as compared to the absorption in the center of the line.

Thus the temperature information of the lower levels will be contained in the wings of the spectral lines. due to increasing temperature and greater absorption towards line center in stratospheric levels the radiance in the central part of spectral lines will be carrying information about these levels. as the radiance emitted from layers will be much attenuated while the emission.

From upper levels will be little due to low deasity of the gas most of the rediance received at satellite level will correspond to intermediate levels and under such varying amount of energy the information is to be retrieved.

That weighting function for a single collisional broadened line has lese half width as compared to that for a spectral Interval consisting of several lines leading to a lower vertical resolution in case of a spectral interval.

As the pressure will be different foe different spectral intervals in a band due to rapid variation of absorption coefficient with frequency the plotting of weighting function against pressure will have peaks at different altitudes, resulting in the overlapping of weighting functions and thus the amount of radiance received in one spectral interval will not only correspond to radiation emitted from the level where its weighting function is peaking but will also be consisting of radiation emitted from other levels making the problem of retrieval of information more difficult.

# 3. Contrast Reduction

Let uslook at the concept of contrast.

Contrast ratio (cr) could be defined as follows:

$$Cr = bmax.(maximumbrightnessofthescene) \\ /b\ \mathrm{min}.(minimumbrightnessofthescene$$

Other modes of defining contrast are:

Differential contrast, =

$$cd = \frac{emax. - emin.}{emin}.$$

Logaithmic contrast.

$$Cl = log10\frac{emax}{emin}$$

Modulation in contrast,

$$m = \frac{emax - emin}{emax + emin}$$

These contrast have been defined in terms of brightness which is defined as the magnitude of the response produced in the eye by light that can only be determined approximately.

Brightness variation may be calibrated with a gray scale.

LUMINANCE is a quantitative measure of intensity oflight from a source, and is measured with a device called a photometer light meter.

The term tone is used for each distinguishable shade from black and white in practice, most interpreters do not use an actual gray scale the way one would use a centimeter or intermediate scale and characterize areas on an image as light, moderate or intermediate, or dark in tone.

Thus the variance contrast parameters defined for brightness could also be defined in terms of object or image radiances or luminances.

Atmospheric physicists frequently use a measure of contrast.

Called the contrast transmittance or contrast Trasmittance coefficient, y. if the target, background, and path radiances are lt, lb and lb respectively and the atmospheric transmittance ist, then the differential contrast at the observer co is defined as:

$$Cod = \left(\frac{(ltt + lp) - (lbt + lp)}{LBT + LP}\right)$$

$$\frac{(\mathrm{lt} - \mathrm{lr})\mathrm{t}}{\mathrm{lbt} + \mathrm{lp}} - 9$$

And the differential contrast at the ground is given by

$$Cgd = \frac{lt - lb}{lb}$$

The contrast transmittance y is defined as

$$\mathrm{Y} = \frac{\mathrm{cod}}{\mathrm{cgd}}$$

$$\left\{1 + \frac{lb}{lbt}\right\} - 1$$

It is unfortunate the y not only depends on the path radiance but also on radiance or the target surround.

Contrast reduction in the atmosphere is primary due to scattering which is a selective (1.0. wavelength dependent) as well as non-selective process. rayleigh and mie scattering are selective processes. rayleigh scattering is due to gas molecules while mie scattering is caused due to particles of smoke, fumes and haze (sizes are comparable to incident wavelength). scattered light luminates shadows that are never completely dark, but are blueish in colour.

Scattered illumination is also referred as skylight to distinguished it from direct sunlight.

Nonselective scattering is caused by dust, fog and clouds with particle sizes more than 10 times the wavelength of light.

These particles scatter all wavelengths equally.

There fore clouds and fog appear white although their water particles are colourless.

# 4. Analytic Treatment of Atmospheric Effects

The interaction of incident solar energy which gets scattered by the atmosphere and reflected by the remotely sensed object and rinally reaches to a detector placed in space.

Here the component refers o the contribution of atmosphere alone.

A. Component refers to the contribution of photons directly transmitted to the ground and reach after reflection

B. Component refers to the photons diffusely transmitted to the ground and directly transmitted after reflection

C. And € refer to photons directly or diffusely transmitted to the ground and diffusely transmitted after reflection (no direct looking by the detector)

Refers to multiple interactions between ground and atmosphere (here only double interaction is shown).

The appearent albedo measured in a direction above atmosphere of optical thickness t illuminated by a solar flux f from the direction.

The terms (a) through (f) on the r. h. s. of this equation refer to te radiance components discussed prior to the writing of this equation.

e\are the scattering and transmission functions of the atmosphere along and are given by and in general.

Here p is the uniform lambertion ground albedo and the equation for p.

Is for uniform lambertion ground and prim to coordinates refer to appearent aspect.

Is relates to the direction of observation specified by the nadir angle a n = co2.

The angle is the azimuth angle referred to a vertical plane passing through the sun and the satellite.

Similarly s * and t *

Could be defined for the atmosphere illuminated at the bottom/(t) m and saredefinel.

Similary the case of a nonuniform, nonlabertian ground could be analytically.

Table 3 gives the astimTES OF tmospheic contributions as a function of Bserved wavelength and e.

***Table 3.*** *Atmospheric Contributton for Vertical Observation.*

| In mm | In degrees | Ratlight contribution | Turbid atmosphere for a visibility of 23 km. | Turbid atmosphere for a visibility of 5 km |
|---|---|---|---|---|
| 450 | 15 | 0.0838 | 0.1050 | 0.1603 |
| 550 | 15 | 0.0376 | 0.0567 | 0.1071 |
| 650 | 15 | 0.0184 | 0.0366 | 0.0815 |
| 850 | 15 | 0.0061 | 0.0208 | 0.0559 |
| 450 | 60 | 0.0988 | 0.1281 | 0.2027 |
| 550 | 60 | 0.0448 | 0.0708 | 0.1420 |
| 650 | 60 | 0.0228 | 0.0454 | 0.1096 |
| 850 | 60 | 0.0077 | 0.0250 | 0.0747 |

One could draw comparative conclusion for the effects due to variation or o and TMOSPHERE VISIBILITY FOR A GIVEN WAVIENGHT, AND ALSO COMPAISON OF THEES AMONG DIFFERENT WAVEIENGTHS.

Except the second term of equation (1) all other terms on the right hand side are the unwanted terms in remote sensing.

Let us now have a view of variability of optical thiciness (table 4).

And function s * (table 5) with reference to the following three atmosphere models:

A. Pure molecular atmosphere.

B. Molecular atmosphere with serosols corresponding to the model defindedbymeclatchey et al, for ground visibility of 5 km.

C. Same as (b) but with ground visibility of 23 km.

***Table 4.*** *Optical thickness for molecules (t) and aerosols (t) for two visibilities, V 23 km. and v 5 km.*

| | | | |
|---|---|---|---|
| Micron | 0.4500 | 0.5500 | 0.8500 |
| Ttr | 0.2157 | 0.0948 | 0.0163 |
| Tr | 0.2801 | 0.2348 | 0.1550 |
| Tp(v 5 km) | 0.9305 | 0.7801 | 0.5151 |

***Table 5.*** *Function s * for four wavelengths and there atmosphere models.*

| Wavelengths in mm | Molecular atmosphere | Turbid atmosphere for v 23 km | Turbid atmosphere for v 5 km |
|---|---|---|---|
| 450 | 0.1605 | 0.2128 | 0.3080 |
| 550 | 0.0807 | 0.1403 | 0.2432 |
| 650 | 0.0438 | 0.1038 | 0.2056 |
| 850 | 0.0157 | 0.0698 | 0.1606 |

Value of s* increases with turbidity and toward shortwave length due to Rayleigh scattering. as s* is multiplied by p in multiple interaction TERMS, it is concluded thet over ocean or low reflecting grounds (say p 0.05 )

The corresponding contribution is of the order of one percent in the worst cases and it can be neglected. over high reflecting grounds ( say p 0.50), the contribution reaches is percent. but in remote sensing of terrestrial sites.

Such high reflectances are generally found for vegetation only in the neer infrared: therefore, the interactions term is only about 7 percent for the worst visibility cass.

# 5. Discussion on Aerosols

Embedded in the gaseors atmosphere is a semi-permanent suspension of liquid and solid particles called serosols.

theparticies arise form a variety of natural and anthropogenic sources such as volcaness, forest fires,dust storms, sea spray.

Industrial smokestekes, automobile exhaust etc.

From such varied the particalecoalexe, and condence to produce a distribution of shpes, sizes, and composition. the shape of liquid particaly that of a sphere whereas solid particles may have shape whatever, however, for a collection of particales in random orintations we can probably assume that the scattering effect is nearly the same as that for a collection of spheres. the sizes of particale range from $10^{-7}$ cm to $10^{-4}$ cm with an approximate Gaussian type distribution.

The composition of aerosols can very from pure water to highly absorbing soot like particles.

The complex index of refraction (m) for aerosols could be given by:M ( ) n ( ) -I k ( ).

Where n given the magnitude of scattered energy K relative amount of absorption.

Both n and k vary with the type of aerosol – a collection of relatively large particles of differing sizes. water and dust haze will have significantly different value of n and k in visible and neerir regions. if the imaginary part k ( ) Is zero, the absorption can be quit important.

The aerosole size distribution function, in (r), under various condition of the atmosphere can contain rather more or lese particles of quite small or quite large sizes.

The resultant scattering of radiation is sensitive to the relative and absolue abundance of each.

The other aspect of aerosols is their height distribution function, n (z). meteorological condition of winds and temperature determine whether particulate matter is confined near the surface or is distributed quite uniformaly with height, and the measured radiations at any given height in the atmosphere will be irfluenced by these differing conditions.

# References


[1] Chahine, m. t: an analytical transformation for remote sensing of caear-colomn atmosphere temperature profiles j atmos. sel, 1995.

[2] Estes, J, E., Manual of remte Sensing, 2ed Edition –volume II., American Socity of photogammetry – the Sheridan press, USA, 1983.

[3] Giboson, P. J, Introductory Remote Sensing Principles and Concept. St Edmunds Bury Press, Great Britanin, 2000.

[4] Fleming h. e: retrieval of atmospheric temperature profiles from satellite measurements for dynamical forecasting 1992.

[5] Frits. s. wark, d, q: temperature sounding from satellits, noaatr ness 59 national oseanic atmospheric Washington d.c. 1972. p. 49.

[6] Kunzi.k.f remote sensing of atmospheric temperature profiles 2009.

[7] Mather, P, M. and Brand, T. Classification Methods For remotery Sensed date, (2nded), Taylor and Francis Group, LLC, 2009.

[8] Miler, g. f: fredholm equations of the first kind, in numerical solution of integral equatios, edited by l.m. delvas and j waisvhclarendon press, oxford, 1994.

[9] Polishchuk, a. l estimations of the information content of meteorological observing system, Leningrad gl, geofis. observ 2002.

[10] Smith w. l inversion techniques for remote sensing of atmosphere temperature drofiles 2005.

[11] Thomas, E, G. and Paul, J. C., Atmosphere Climate and Changes, Scientific American Library, New Yurok, 1999.

[12] Weley, K.K. Modern Physical, 1996.

[13] Westwater, ed. r inversion techniques in remote sensing of the troposphere, edited by y. e derr, chapter 16, national oceanic and atmospheric administration, Washington, d.c 1972.