Simulation studies of satellite spectral radiances for 11.1, 8.3, 7.3, 6.7 μm IR bands

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Abstract. Starting with an outline of spectral radiance calculation for infrared bands, a simulation study has been presented for 11.1, 8.3, 7.3 and 6.7 μm . Detailed response of the spectral radiance to the meteorological parameters has been studied by generating simulated atmospheric temperature and moisture profiles for Indian atmospheres. It has been found that the responses of the 8.3, 7.3 and 6.7 μm bands are maximum for lower troposphere, middle troposphere and upper troposphere respectively, and these spectral radiances are linearly related to the water vapour content of the relevant layers. The thermal IR window at 11.1 μm is found to be sensitive to not only surface temperature and total water vapour content but also to the boundary layer values of mean temperature and water vapour content.

Keywords. Satellite infrared radiances ; radiative transfer ; atmospheric transmittances ; atmospheric simulations ; infrared absorption ; water vapour content.

1. Introduction

Multichannel infrared radiometers on board meteorological satellites provide information on the distribution of temperature and minor constituents in different layers of the atmosphere. For actual extraction of parameters, however, the relationship of the spectral radiance measured at the top of the atmosphere and the associated parameters is not a simple one. (Smith and Woolf 1976). In general, the total vertical temperature profile and the vertical distribution of all the constituents determine the radiance emerging from the top of the atmosphere and reaching the satellite sensor. The situation, however, is greatly simplified if observations are performed in a few select bands *e.g.* 6.7 μm water vapour band, where only one or two minor constituents play dominant roles. In such cases, the atmospheric transmittances can be modelled in a much more simpler form than other frequencies. (for example, Prabhakara *et al* 1974). Numerous simulation studies have already been conducted (Anding *et al* 1971; Sidran 1980; Weinreb and Grossby 1972) to assess the response of the spectral radiances to the meteorological parameters and several models have been proposed for parameter extraction (Smith and Woolf 1976). The common feature of all these studies is known from

transmission characteristics of the atmosphere by assuming climatic profiles for parameters. The moisture distribution is known to vary considerably specially in Arabian sea (Ramage 1971) and therefore, a mean climatic profile for water vapour absorption calculations will lead to an erroneous conclusions. In this paper, therefore, we have attempted to see the effect of perturbations over water vapour profile within the first few kilometers of the troposphere. We consider the water vapour sensitive channels at $11.14 \mu\text{m}$, $8.3 \mu\text{m}$, $7.3 \mu\text{m}$ and $6.7 \mu\text{m}$.

2. Procedure for radiance calculations

2.1 Radiative transfer

The radiative transfer equations are a set of integro-differential equations (Chandrasekhar 1960) whose solutions can be written formally in terms of deriving source function (absorption/scattering) in atmosphere. Skipping the details of these analytic expressions, we can directly construct the solution for upwelling radiance for a plane parallel atmosphere which is divided into n -homogeneous layers. We construct the radiance function by summing contributions from all individual layers which are assumed to contain only molecular absorbers and a haze medium which causes absorption as well as scattering. We also assume it is *cloud-free*.

The spectral radiance emerging at top of the atmosphere at any frequency, ν , and direction, θ , will have contributions from (a) the surface emissions, (b) upward emissions from each layer of the atmosphere and (c) the downward atmospheric emissions reflected back by the surface.

Let T_s be the surface temperature and $E_s(\nu)$ its emissivity at frequency ν , then the spectral radiance from surface emission reaching the satellite height is,

$$E_s(\nu) B(\nu, T_s) \tau_{0,n}(\nu, \theta) \quad (1)$$

where $\tau_{0,n}$ is the total transmittance of the atmosphere and $B(\nu, T)$ is the Planck's function,

$$B(\nu, T) = \frac{2h\nu^3}{C^3(e^{h\nu/kT} - 1)} \quad (2)$$

The upward radiance from the atmosphere will be given by,

$$B(\nu, T_i) \epsilon_i(\nu, \theta) \tau_{i+1,n}(\nu, \theta) \quad (3)$$

for i th layer with mean temperature T_i and emissivity $\epsilon_i(\nu, \theta)$ ($\tau_{i+1,n}(\nu, \theta)$ is the transmittance of the rest of the atmosphere. The total atmospheric emission in upward direction will be,

$$\sum_{i=1}^n B(\nu, T_i) \epsilon_i(\nu, \theta) \tau_{i+1,n}(\nu, \theta) \quad (4)$$

The downward atmospheric emission will be

$$\sum_{i=1}^n B(\nu, T_i) \epsilon_i(\nu, -\theta) \tau_{0,i-1}(\nu, -\theta)$$

Where $\tau_{0, i-1}$ is atmospheric transmittance between surface and $(i-1)$ th layer. This radiance will be reflected back by the surface with an effective reflectance $(1-E_s)$ and transmitted through the atmosphere to the satellite sensor. Therefore, the total upwelling radiance emerging at the top of the atmosphere will be

$$\begin{aligned}
 I(\nu, \theta) = & B(\nu, T_s) E_s(\nu) \tau_{0, n}(\nu, \theta) \\
 & + \sum_{i=1}^n B(\nu, T_i) \epsilon_i(\nu, \theta) \tau_{i+1, n}(\nu, \theta) \\
 & + (1 - E_s) \tau_{0, n} \sum_{i=1}^n B(\nu, T_i) \epsilon_i(\nu, -\theta) \tau_{0, i-1}.
 \end{aligned} \quad (5)$$

For normal sea conditions, the sea surface emissivity is close to unity for thermal infrared band, therefore, the reflectance term in (5) can be dropped.

Assuming that the scattering at any layer does not contribute significantly to radiance in direction other than the direction of propagation, we can approximate (using Kirchoff's law).

$$\epsilon_i(\nu, \theta) = 1 - t_i^{\text{abs}}(\nu, \theta), \quad (6)$$

Where t_i^{abs} is the transmittance due to absorption alone of the i th atmospheric layer.

2.2 Atmospheric transmittance

The atmospheric transmittance depends entirely upon the distribution of its constituents. In general, the infrared active molecules like H_2O , CO_2 , CO , N_2O , CH_4 , O_3 , etc and the atmospheric haze will contribute to the transmittances in infrared spectral bands. The molecular absorption by atmospheric gases has been studied in detail and vast amount of data are available for the IR region (Prabhakara *et al* 1974; McClatchey *et al* 1972). The molecular scattering is given by λ^{-4} dependence (Rayleigh) which becomes important for $\lambda \leq 4\mu\text{m}$. The haze effects are discussed separately.

Apart from molecular line absorption, the distant line wings and molecular collision which introduce line broadening effects contribute a slow varying continuum absorption. The continuum contribution is still not very well understood and the absorption coefficients are known only approximately (Roberts *et al* 1976; Selby *et al* 1976). Since the radiometric observations provide radiance measurements over certain spectral width, the transmittance function could be averaged over the spectral interval of interest. These degraded (in resolution) transmittances can be calculated by employing one of the many band models available (e.g. Goody 1964; Elsasser 1938).

For calculations presented in this paper, we have used transmittance (Selby *et al* 1978) based on a generalised band model of the type

$$\tau = f(c(\nu) \triangle LP^n), \quad (7)$$

where $c(\nu)$ are spectral coefficients for 10 cm^{-1} resolution, $\triangle L$ is the equivalent

path length of the absorber and P is the effective pressure. The spectral coefficients are obtained by a statistical fit to actual laboratory measurements and the molecular absorption spectra calculations. (McClatchy *et al* 1972).

2.3 Haze-effects

The haze medium is generally considered to be spherical water droplets. For oceanic regions, salt spray is expected to perturb the size-distribution while preserving the shape. For land surfaces, dust and soot particles also contribute. The haze contribution is given by Mie theory where the shape of the scattering particles is assumed to be spherical. The magnitude of the extinction as well scattering at any layer of atmosphere is expressed in terms of particle number density for the layer and the spectral coefficients (β_α and β_s). These spectral coefficients are known in terms of Mie scattering functions and particle size distribution $n(a)$,

$$\beta_{\alpha,s}(\nu) = \int_0^\infty Q_{\alpha,s}(\nu, a) \pi a^2 n(a). \quad (8)$$

The Mie scattering functions for spherical particles are explicitly known in terms of Ricatti-Bessel functions and complex index of refraction for the scattering medium (Born and Woolf 1964). For our calculations we have used Elterman (1968) aerosol profile for particle number density distribution for a clear sky. The spectral functions are used from calculations of Selby *et al* (1976) which incorporate optical properties of liquid water (Hale 1973) and a drop size distribution of standard Maritime haze.

3. Climatic model for Indian region

Climatic mean relative humidity profile has been studied by Nieman (1977) for tropics using radiosonde data. It has been found that in all cases where there is no inversion layer existing, the height variation of relative humidity is more or less linear up to 200 mb (figure 1). Though this data is heavily weighted by land observations, the data from ships and island over north Atlantic and North Pacific support this structure. We assume that for northern Indian Ocean also this structure will be valid except for the periods during summer monsoon where the western Arabian Sea develops a strong inversion layer below 800 mb (Ramage 1971).

Further, to study the effect of the tropospheric boundary layer perturbations, we introduce a relative humidity drop at a height, h_i , (< 3 km). This relative humidity drop, is generally observed when there is a temperature inversion present (Ramage 1971). This phenomena of inversion usually occurs over Arabian Sea during southwest monsoon period and is accompanied by a temperature warming within the boundary layer. It is also known that the higher the temperature warming, the larger will be the relative humidity drop across the inversion layer.

For the calculations presented in this paper we assume a moisture variation for normal profiles governed by,

$$RH(h) = RH_s - \mu h, \quad (9)$$

Figure 1. Relative humidity and temperature profile for tropics (non-inversion case), Nieman 1977.

where RHS is surface relative humidity and μ is lapse rate of relative humidity. It is believed that most of the moisture is distributed in the first few kilometers of the troposphere, therefore, (9) will adequately describe the moisture distribution up to 10 kms and then the values from U. S. Standard Tropical Atmosphere are used. The temperature profile is considered to be independently varying (except that its lapse rate is adjusted according to moisture in boundary layers) with a variation of type,

$$T(h) = T_s - \lambda h, \quad (10)$$

where T_s is surface temperature.

To simulate inversion phenomena, we introduce a temperature warming, ΔT (2° to $4^\circ K$) at a height 1.25 km and a relative humidity drop, ΔRH (40% to 60%) across the inversion layer.

Thermodynamically, the temperature lapse rate is dependent upon the moisture held by the air. For our purpose instead of implementing a detailed adiabatic calculations, we obtained temperature and moisture profiles by small perturbations over the climatic averages for Indian region (Sasi and Sengupta 1979). We assumed a lapse rate (temperature)

$$\begin{aligned} \lambda &= 5^\circ \text{ to } 6.5^\circ \text{ (steps } 0.5^\circ) \quad h < 7 \text{ Km} \\ &= 6^\circ \text{ to } 8^\circ \quad 7 < h < 16 \text{ Km} \end{aligned}$$

and normal above tropopause. For humidity, a lapse rate between 6-8% was considered (in step of 0.5). The surface values T_s was varied between 297° to 302°K and the relative humidity from 60% to 85% (steps of 5%), for oceanic regions. The pressure for each layer (1 km thick up to 25 km of height) was calculated using the hydrostatic equation,

$$dp = -g\rho dZ,$$

and the surface pressure P_s is normally fixed at 1013 mb.

4. Sample calculations for 11.1, 8.3, 7.3 and 6.7 μm

Calculations have been made on the basis of foregoing discussion for 11.1, 8.3, 7.3 and 6.7 μm IR bands for various meteorological conditions with varying water vapour content. The direction of radiance measurement is chosen to be nadir ($\theta=0$) for simplicity. For larger look angles, the limb effects become important. Normally, it is believed that 11.1 μm thermal IR 'window' is free from serious absorptional effects (Streete 1968) and therefore, best suited for sensing surface temperatures through corrections to the order of 8° in surface temperature ($\sim 300^\circ\text{K}$) were found to be necessary because water vapour continuum (Mc Coy 1969) interferes with this observation. The other three bands *i.e.*, at 8.3, 7.3 and 6.7 μm are located in the wing of strong water vapour absorption band at 6.3 μm and therefore, are expected to sense the water vapour content of lower, middle and upper troposphere respectively.

4.1 Vertical structure of radiance contribution

Let us reexamine (5) and (6).

We can write

$$\tau(\nu, \theta) = \tau^{\text{abs}}(\nu, \theta) + \tau^{\text{scat}}(\nu, \theta),$$

where 'abs' stands for total absorption and 'scat' for scattering contributions. Equation (6) is then rewritten as

$$I(\nu, \theta) = B(\nu, T_s) E_s(\nu) \tau_{0,n}(\nu, \theta) + \sum_{i=1}^n W_i(\nu, \theta, T_i) \tau_{i+1,n}^{\text{scat}}(\nu, \theta) \Delta Z_i, \quad (11)$$

$$\frac{\Delta Z_i}{W_i(\nu, \theta, T_i)} = \frac{B(\nu, T_i)}{B(\nu, T_s)} \frac{\tau_{i+1,n}^{\text{abs}} - \tau_{i,n}^{\text{abs}}}{\tau_{i+1,n}^{\text{abs}}} \Delta Z_i \quad (12)$$

The function, W_i (12) denotes emission of individual layers of the atmosphere. Due to pressure dependence for the total atmospheric attenuation, W_i is expected to show preference to only a few particular layers of the troposphere. A typical variation for undisturbed profiles of W_i with height is shown in figure 2 for IR bands under consideration. For inversion case, variation of W_i is compared against normal case (figure 1) for 11.1 μm spectral channel in figure 3.

Figure 2. Atmospheric emission, W_i , for different layers of troposphere for undisturbed profile figure 1.

Figure 3. Comparison of atmospheric emissions for normal and with-inversion for 11.14 μm channel.

4.2 Response of radiance to temperature and water vapour content

The total radiance can be written as

$$R = R_a + R_e$$

where R_s is contribution from surface alone while R_e is emission from troposphere. As mentioned above R_e will get contribution from a limited portion of troposphere. Therefore, if the mean temperature of the region be $\bar{T}$ and water vapour content (of the region) $\bar{W}$ then a simple relationship :

$$R_s = b(\bar{T}) + a(T)W \quad (13)$$

might be adequate to exhibit response of the spectral radiance in a given range of variations for T . The surface term, R_s on the other hand will be dependent upon the total water vapour content (W_{tot}) because it involves whole of the atmosphere. Therefore, we could possibly have

$$R_s = d(T_s) + C(T_s)W_{tot}. \quad (14)$$

On the basis of atmospheric simulations as described in §§ 2 and 3, it has been found that for 7.3 and 6.7 μm channels, the surface term accounts for less than 1% of the total radiance. On the other hand, these contributions are very significant (above 30%) for 11.1 and 8.3 μm . A linear regression analysis for surface term leads to an extremely high ($\geq 95\%$) correlation. The results are summarised in table 2.

4.2a Emission term The emission presents a very different picture. The 6.7 μm radiance terms exhibit 83% correlation with the water vapour of 4-11 km irrespective of boundary layer distributions of water vapour. The 7.3 μm

Table 1. Surface term for 11.14 and 8.3 μm ($R_s = a + bW$)

Wave length	Correlation	a	b
11.14 μm	-99.2%	118.13	-1.413
8.3 μm	-99.2%	37.33	-3.9696

Table 2. $R = R_0 + bW + c\bar{W}$

Wave length μm	Layers kms	R_0	b	c	correlation coefficients		
					r_w	$r_{\bar{w}}$	X^2
11.14	0-2	120.23	-0.403	0.1810	-98%	-90%	0.01
8.3	0-3	53.43	-0.214	-0.01	-98%	-95%	0.01
7.3	1-6	18.50	-0.003	-0.068	-92%	-94%	0.01
6.7	4-11	7.44	-0.0048	-0.02	-83%	-83%	0.01

Table 3. Brightness temperatures for normal and disturbed profiles for 11.14 μm .

Surface temp. ($^{\circ}\text{K}$)	Surface relative humidity	Normal	Brightness Temp. ($^{\circ}\text{K}$) inversion
296	65	292.25	293.63
	70	291.99	293.41
	75	291.74	293.17
	80	291.48	292.93
	85	291.21	292.67
300	65	295.38	297.28
	70	295.05	296.98
	75	294.70	296.67
	80	294.34	296.33
	85	293.98	295.98

spectral radiances show a cut back in the lower level terms in presence of boundary layer inversion enhancing the observed correlation (92%) for the undisturbed (figure 1) profile.

The 8.3 μm and 11.14 μm spectral radiances vary very differently because of their sensitivities for lowest levels of the troposphere. There is an enhanced emission from layers up to inversion height so much so that the aggregate reaching the top of atmosphere (including surface term) increases quite appreciably (table 3). Since 11.14 μm is normally used to estimate surface temperature by incorporating water vapour corrections based on climatological mean-profile-behavior of transmittances, the boundary layer perturbation will lead to an effectively low value of water vapour correction, thus the boundary layer response of 11.14 μm channel should be taken into account for any proper sea surface temperature estimation.

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