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A microwave radiative transfer model

L. Phalippou

Research Department

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European Centre for Medium-Range Weather Forecasts
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Abstract

A microwave (1-300 GHz) radiative transfer model for a non-scattering plane parallel atmosphere is presented. It allows the user to compute the monochromatic brightness temperature at the top of the atmosphere for a given atmospheric profile including non-precipitating clouds and various surface conditions. After an introduction and a brief background on radiative transfer, the main features of the model are described.

1. INTRODUCTION

Microwave radiometry has become a key element for sensing the ocean/atmosphere system from space. One of the main advantages of microwave over infra-red measurements is the capability to sense the atmosphere and/or the surface through moderate clouds. Among the most widely used microwave radiometers currently operating are MSU (Microwave Sounding Unit) aboard National Oceanic and Atmospheric Administration (NOAA) operational meteorological satellites, SSM/T and SSM/I (Special Sensor Microwave/Imager) aboard the DMSP (Defence Meteorological Satellite Programme) satellites. At present, the 22 GHz water vapour absorption line and the 50-60 GHz oxygen absorption band are used to measure the precipitable water content and the temperature of the atmosphere. Window frequencies below 85 GHz are used for sensing cloud, rain and surface parameters. The "new generation" of microwave radiometers (SSM/T2, AMSU.A/B) includes additional channels near the 183 GHz water vapour absorption line which will allow the measurement of the vertical distribution of humidity, even in the presence of moderate clouds.

Previous experiences with Tiros Operational Vertical Sounder (TOVS) have shown that the interpretation and the exploitation of meteorological satellite data for numerical weather prediction (NWP) requires a radiative transfer (RT) model which computes the radiance at the top of the atmosphere (ToA) for a given atmospheric profile and surface conditions. Numerous RT models have been developed in the past. For instance, *Eyre* (1991) describes a fast RT model for TOVS variational assimilation in NWP.

When research on SSM/I data started at ECMWF in 1990 a RT transfer code was needed to help in the interpretation of the data. One of the first needs was to evaluate the sensitivities of the radiances to geophysical parameters (*Phalippou*, 1992) and at this time Eyre's RT model was not set up for SSM/I. For example it did not include cloud and surface contributions were not modelled.

The model presented here is a part of the software developed at Alcatel Espace to simulate the measurements for a proposed microwave radiometer in geostationary orbit. This study was performed for the European Space Agency (ESA) in the framework of the Meteosat Second Generation project (*Phalippou et al.*, 1989). The complete software including the radiative transfer in a scattering atmosphere and the simulation of the instrument is described in *Phalippou* (1991).

The non-scattering part of the radiative transfer code has been implemented on the Cray and on the CDC Cyber at ECMWF . Provisions have been made to allow a future modelling of the scattering by the hydrometeors contained in clouds or precipitation. The cloud absorption along the profile and the scattering effect of rough surfaces are already implemented.

The main features of the model are as follows:

- use of Stokes parameters to describe the radiation
- line by line computation of the gaseous absorption coefficients
- modelling of the absorption along the cloud profile
- Lambertian reflection over land
- specular reflection over calm sea
- model of surface scattering for a rough sea (wind effect).

This RT model is adequate for computing the brightness temperature for the 1-300 GHz frequency range in clear sky and in most of cloudy conditions as long as the scattering effect is negligible.

It is intended to be a research tool with the software being written in a flexible manner.

2. BACKGROUND

The main relationships used to compute the brightness temperature at the ToA are introduced in this section. For a comprehensive introduction to microwave remote sensing the reader is referred to *Ulaby et al.* (1981, 1986).

2.1 Brightness temperature at the top of the atmosphere

In the following we will assume that the Rayleigh-Jeans approximation of Planck's law can be used, i.e. the brightness of a blackbody is given by $2k\lambda^{-2}T$ where k is the Boltzman constant, λ the wavelength and T the temperature (K). The Rayleigh-Jeans's fractional deviation from Planck's law is less than 1% if $\lambda T > 0.77$ mK. For a plane parallel atmosphere, the upwelling monochromatic brightness temperature T_b at the *ToA* (Fig. 1) is given by

$$Tb_{ToA}(\nu,\theta) = Tb_{up}(\nu,\theta) + tr_{\nu,\theta}(o,z_t)\left[Tb_s(\nu,\theta)+Tb_{sca}(\nu,\theta)\right] \tag{1}$$

where ν is the frequency, θ is the incidence angle and z_t the vertical coordinate at the *ToA*.

Tb_{up} represents the upward emission by the atmosphere, and can be written (using Rayleigh-Jeans approximation)

$$Tb_{up}(\nu,\theta) = \sec\theta \int_o^{z_t} K_\nu(z)\ T(z)\ \exp[-\tau_\nu(z,\infty)\sec\theta]\ dz \tag{2}$$

$K_\nu(z)$ is the absorption coefficient (Np/m) of the atmosphere. It depends on the frequency, the concentration of the constituents of the atmosphere and on the temperature. $T(z)$ is the physical temperature (Kelvin).

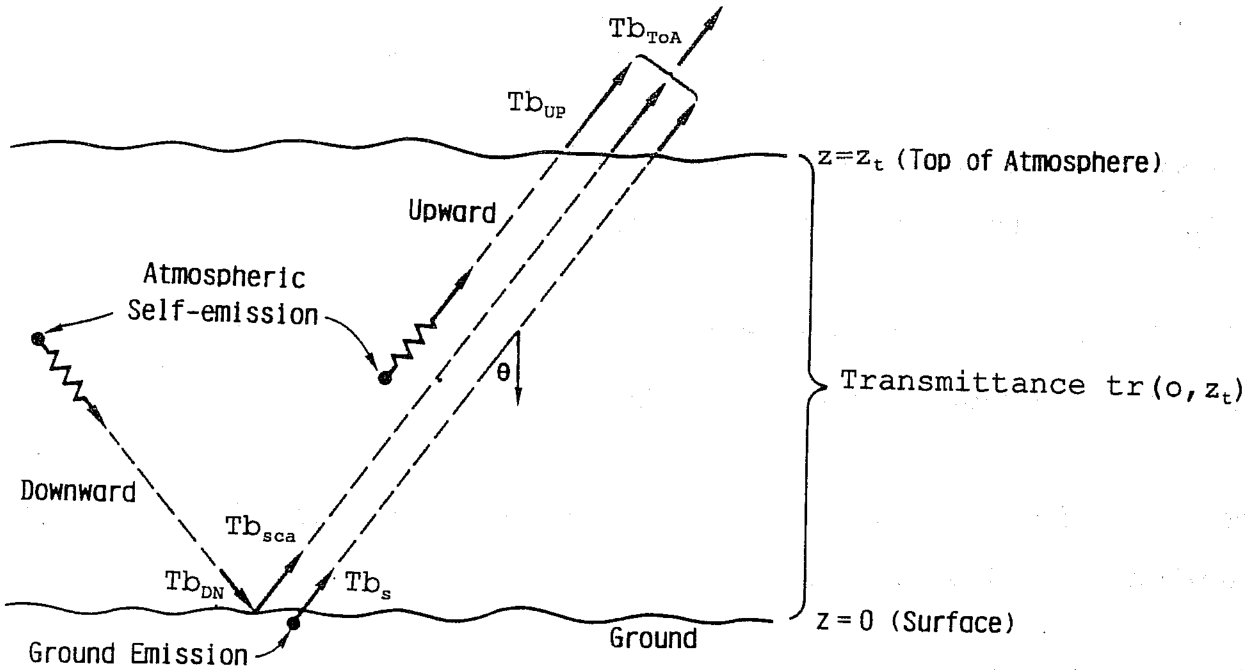

Fig. 1 Brightness temperature at the ***ToA*** (adapted from *Ulaby et al.*, 1986)

The optical thickness τ_ν of a slab (z_1,z_2) is defined as

$$\tau_\nu(z_1,z_2) = \int_{z_1}^{z_2} K_\nu(z)dz \quad \text{with} \quad z_2 > z_1 \tag{3}$$

The transmittance (tr) of the atmosphere for a path defined by (z_1, z_2) and the angle θ is given by

$$tr_{\nu,\theta}(z_1,z_2) = \exp[-\tau_\nu(z_1,z_2)\sec\theta] \tag{4}$$

The second term of (2.1) represents the surface contribution at the ***ToA***. The definitions of $Tb_s(\nu,\theta)$ and $Tb_{sca}(\nu,\theta)$ are given in the following section.

2.2 Brightness temperature of the surface

In the microwave region the emission of the surface is generally written as the product of the surface emissivity e_s and the surface temperature T_s,

$$Tb_s(\nu,\theta) = e_s(\nu,\theta)\ T_s \tag{5}$$

This expression assumes that the emitted radiation comes from a semi-infinite terrain with homogeneous dielectric properties and temperature. In practice (5) is valid for an isothermal medium with finite thickness H, provided its optical thickness is $\tau(0,H) >> 1$. For instance, for sea water and frequencies above 10 GHz, the penetration depth of the radiation (defined as the inverse of the absorption coefficient K) is very small and T_s can be taken equal to the skin temperature. For other surfaces such as snow, this approximation does not hold and a RT equation needs to be solved in the terrain itself.

The emissivity of the surface is usually polarisation-dependent, so for a non-scattering atmosphere the surface is the only source of polarisation of the radiation at the *ToA*.

$Tb_{sca}(\nu,\theta)$ is the brightness temperature scattered by the surface and is related to the surface roughness and to the downwelling atmospheric radiation Tb_{DN} (Fig. 1).

For a rough surface the general mathematical expressions of e_s and T_{sca} are complex, but may be simplified considerably for the following cases:

2.2.1 *Specular surface*

If the surface is smooth (height variations are smaller than the wavelength) the theory of reflection of electromagnetic waves at a plane interface introduces the well-known Fresnel reflection coefficients. In this case, one can show that the emissivity is a function of polarisation (the usual vertical V and horizontal H convention is assumed) and is given by

$$e_V(\theta,\nu) = 1 - \Gamma_V(\theta,\nu)$$
$$e_H(\theta,\nu) = 1 - \Gamma_H(\theta,\nu) \tag{6}$$

where $\Gamma_{V,H}$ are the reflectivities in V, H polarisations. $\Gamma_{V,H}$ are related to the Fresnel reflection coefficients $R_{V,H}$ by the expressions

$$\Gamma_V(\theta,\nu) = |R_V(\theta,\nu)|^2 = \left|\frac{\varepsilon(\nu)\cos\theta-(\varepsilon(\nu)-\sin^2\theta)^{½}}{\varepsilon(\nu)\cos\theta+(\varepsilon(\nu)-\sin^2\theta)^{½}}\right|^2$$

$$\Gamma_H(\theta,\nu) = |R_H(\theta,\nu)|^2 = \left|\frac{\varepsilon(\nu)-(\varepsilon(\nu)-\sin^2\theta)^{½}}{\varepsilon(\nu)+(\varepsilon(\nu)-\sin^2\theta)^{½}}\right|^2 \tag{7}$$

where $\varepsilon(\nu)$ is the complex dielectric constant of the surface.

For a specular surface one can show (*Ulaby et al.*, 1981) that T_{sca} is given by

$$T_{sca}(\nu,\theta) = [1-e_s(\theta,\nu)]\ T_{DN}(\theta,\nu) \tag{8}$$

where $e_s = e_V$ or e_H given by (6). $T_{DN}(\theta,\nu)$ is the downwelling brightness temperature in direction θ, due to atmospheric emission. The expression of $T_{DN}(\theta,\nu)$ is of course very similar to (2), i.e.

$$T_{DN}(\nu,\theta) = \sec\theta \int_{o}^{z_t} K_\nu(z)\ T(z)\ \exp[-\tau_\nu(o,z)\sec\theta]\ dz + Tb/cosmic\ \exp[-\tau_\nu(0,z_t)\ \sec\theta] \qquad (9)$$

where *Tb/cosmic* is the downwelling brightness temperature at the *ToA* due to extraterrestrial origin (set equal to 2.7 K for the microwave spectrum).

The common form of the radiative transfer equation in the atmosphere with specular reflection at the surface is then obtained by inserting (5) and (8) in (1)

$$Tb_{ToA} = Tb_{up} + tr(o,z_t)\ [e_s T_s + (1-e_s) T_{DN}] \qquad (10)$$

where the (ν,θ) dependence has been omitted for clarity.

2.2.2 *Lambertian surface*

In this case, according to Lambert's law, the emitted scattered radiation is isotropic in the upper hemisphere and polarisation independent. If the atmosphere is symmetric around the z axis, Tb_{sca} is given by

$$Tb_{sca} = 2a \int_{o}^{\pi/2} T_{DN}(\theta)\ \sin\theta\ \cos\theta\ d\theta \qquad (11)$$

where a is the albedo of the surface.

The emission of the surface is then given by

$$Tb_s = (1-a)\ T_s \qquad (12)$$

3. RADIATIVE TRANSFER SCHEME

The scheme used to compute the atmospheric contributions (Tb_{up}, Tb_{DN}) and the interface atmosphere/surface are described in this section. The RT code is written in FORTRAN. The user must provide RADABS with an atmospheric profile, surface conditions (roughness, surface temperature), frequency and the viewing angle. RADABS returns the so-called I and Q Stokes parameters at the *ToA*, defined by

$$I = \frac{Tb_V + Tb_H}{2} \qquad Q = \frac{Tb_V - Tb_H}{2} \qquad (13)$$

where $Tb_{V,H}$ are the brightness temperature in the V and H polarisations. If the atmosphere is cloudy, RADABS compute (I,Q) at the *ToA* with and without the cloud contribution. The reason for using I, Q comes from the previous version of the code developed for a scattering atmosphere (see *Phalippou*, 1991). This should make a future implementation of the scattering part easier.

3.1 Atmospheric contribution

The atmosphere is described by the pressure, temperature, specific humidity, cloud liquid and ice density on N levels between the ToA and the surface (ToA = level 1). For computing the atmospheric upwelling and downwelling radiation (2), (9), it is assumed that the layer $i-1$ defined by the levels $(i-1,i)$ is homogeneous, i.e. that the temperature and the absorption coefficient are constant. The program does not check that this approximation is valid. However, one can do so by looking at the convergence of (I,Q) returned by RADABS in function of the vertical discretization of the atmosphere. For instance, for MSU channels (50-60 GHz), 40 levels are generally used for representing the atmosphere between the surface and the ToA taken to be equal to 0.1 mbar (*Eyre*, 1991).

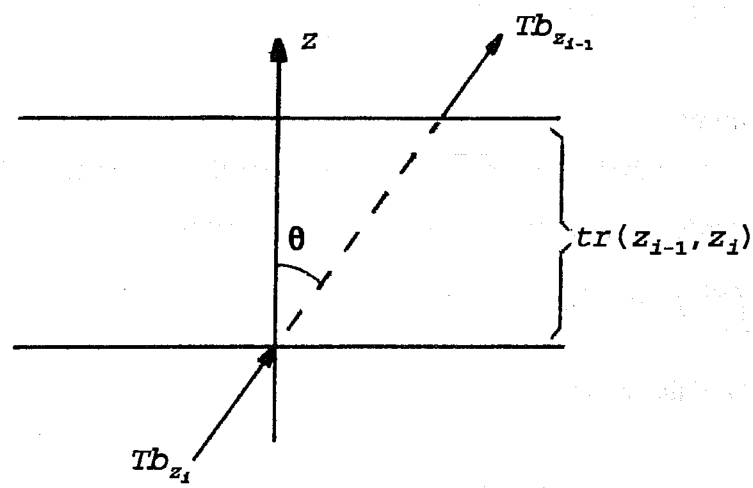

Fig. 2 Radiative transfer in an homogeneous atmospheric layer

For an homogeneous layer and the geometry of Fig. 2, it is easy to show that the upwelling radiation at level $i-1$ is given by

$$Tb_{z_{i-1}} = Tb_{z_i}\, tr(z_{i-1},z_i) + [1-tr(z_{i-1},z_i)]\; T_{i-1,i} \tag{14}$$

where $T_{i-1,i}$ is the temperature of the layer $i-1$, assumed to be constant between z_i and z_{i-1}.

Tb_{up} and Tb_{DN} (see (2), (9)) are computed using (14) for each layer. In the code the temperature of the layer $T_{i-1,i}$ is taken as

$$T_{i-1,i} = \frac{T_{i-1} + T_i}{2} \tag{15}$$

and the transmittance is computed using

$$tr(z_{i-1},z_i) = \exp[-\sec\theta(K_{cloud}(i-1,i) + K_{gas}(i-1,i))(z_{i-1}-z_i)] \tag{16}$$

where the absorption coefficients are given by

$$K_{gas}(i-1,i) = \frac{K_{gas}(i-1) + K_{gas}(i)}{2} \tag{17}$$

and if $K_{cloud}(i-1)$ and $K_{cloud}(i)$ non null then (18)

$$K_{cloud}(i-1,i) = \frac{K_{cloud}(i-1) + K_{cloud}(i)}{2} \tag{19}$$

otherwise

$$K_{cloud}(i-1,i) = 0 \tag{20}$$

(19) shows that the model allows multi-layer clouds. It must be noted that if a level i contains liquid or ice phase, and if the levels $i-1$, $i+1$ are liquid or ice free, then the cloud absorption at level i will be discarded due to condition (18).

The thickness of the layer $(z_{i-1}-z_i)$ is computed using the hydrostatic equation.

3.2 Interface with the surface

For a specular surface it has been shown in section 2.2.1 that the knowledge of the downwelling radiation T_{DN} is only needed along the specular direction θ, see (10).

For the more general case of a rough surface, it is necessary to know T_{DN} along any θ direction of the upper hemisphere (see (11) for instance). Therefore, before computing the surface contribution the program checks if the surface is rough or specular. If the surface is rough then T_{DN} is computed for 10 values of θ between [0, 90°] with a step of 10°. A linear interpolation in θ is then used to compute the downwelling radiation along any direction θ if needed (see section 5).

4. MODEL OF ABSORPTION BY THE ATMOSPHERE

4.1 Gaseous absorption

The oxygen and the water vapour are the two main absorbing gases of the atmosphere in the microwave spectrum. *Liebe* (1981, 1989), *Liebe et al.* (1991) propose regularly up-dated versions of a line by line model using all the O_2 and H_2O absorption line lying in the band 1-1000 GHz. This model has been tested in the frame of ITRA (Intercomparison of Transmittance and Radiance Algorithms) by comparing radiances measured by an upward-looking radiometer at 20.6, 31.6 and 90 GHz with radiances computed using collocated radiosoundings. Table 1 gives a summary of the statistics of the differences between the observed and computed brightness temperatures for various versions of Liebe's model. They give very similar results for the frequencies reported in Table 1. The 1989 and 1991 versions are available in the code.

Table 1: (*Tb* observed - *Tb* computed) statistics for 110 ITRA profiles, and 3 versions of Liebe's model.

F	20.6 GHz			31.6 GHz			90 GHz		
Model	B	SD	RMS	B	SD	RMS	B	SD	RMS
Liebe 81	1.33	1.95	2.4	-.34	0.77	0.8	-.38	2.02	2.1
Liebe 89	2.38	1.48	2.8	0.98	0.75	1.2	0.91	1.88	2.1
Liebe 91	2.51	1.47	2.9	0.82	0.75	1.1	1.00	1.86	2.1

B: Bias (Kelvin)
SD: Standard Deviation (Kelvin)
RMS: Root Mean Square (Kelvin)

4.2 Cloud absorption

Generally cloud droplets (liquid and ice) produce absorption and scattering of the radiation (*Liou*, 1980). We have assumed in this RT model that the Rayleigh approximation is valid (see *Ulaby et al.*, 1981 for details). Moreover, as for water and ice sphere, the refractive index is such that the absorption coefficient is much larger than the scattering coefficient in the Rayleigh region, we only need the cloud absorption coefficient K_{cloud} given by:

$$K_{cloud} = 6\pi \ 10^{-14} \frac{\varepsilon''}{(\varepsilon'+2)^2 + (\varepsilon'')^2} \nu \, d \qquad (Np/m) \qquad (21)$$

with ν: frequency (Hz)

$\varepsilon', \varepsilon''$: real and imaginary parts of the water (ice) dielectric constant

d: water (ice) content of cloud in g/m^3

For water phase, ($\varepsilon', \varepsilon''$) are computed using the model reported in *Ulaby et al.* (Annex E.2.1 (1986)), where the dielectric constant depends on frequency and temperature of water. In the RT model the cloud droplets are assumed to be at the air temperature. For ice phase, very few data are available over the 1-300 GHz range. As reported in *Ulaby et al.* (1986), Annex E.3, we use $\varepsilon'_{ice} = 3.15$ and we take a "mean value" of $\varepsilon''_{ice} = 5 \ 10^{-3}$ independently of the temperature and the frequency.

5. MODEL OF SURFACE BRIGHTNESS TEMPERATURE

For window frequencies, i.e. when the transmittance is close to 1, one can show that the surface contribution at the ***ToA*** is usually larger than the upwelling brightness temperature due to the atmosphere alone (see equation 1) and therefore the surface contribution deserves careful modelling.

Two models for computing the sea surface brightness temperature are briefly described in sections 5.1 and 5.2. A very basic model for land brightness temperature is given in section 5.3.

5.1 Specular sea

We assume that the sea surface roughness responds instantaneously to the surface wind vector. This hypothesis is probably true for capilarity waves but is obviously false for larger wavelength. In the RT model the sea surface is considered a specular surface when the wind is calm. The contribution of the surface is then given by (5), (6), (7) and (8). The saline water dielectric constant ε, is computed using *Klein and Swift* model (1977). The salinity of the sea is taken equal to 36 parts per thousand. Above 10 GHz the effect of the variation of sea salinity on ε is negligible.

5.2 Rough sea

The rough sea model is largely inspired by *Stogryn* (1967), *Wilheit* (1979), *Prigent-Benoît* (1988), *Abba* (1990) among others. The basic idea is to represent the sea surface as a collection of flat facets, each one acting as a specular surface element. Each facet is described by the components of its slope along and across the wind direction. The *Cox and Munk* (1954) bidimensional slope distribution is used. The shape of this distribution is nearly gaussian with mean-square slope components (cross wind and up/down wind) linearly related to the wind speed.

The contribution of one facet to the upwelling radiation at sea level can then be written as

$$P(\vec{F},d)\ [1-\Gamma(\vec{K}_o,\vec{F},p))\ T_s + \Gamma(\vec{K}_o,\vec{F},p)\ T_{DN}(\vec{K}_o,\vec{F})] \qquad (22)$$

with the geometry given in Fig. 3, where

$P(\ )$:	weighting function of the facet
$\vec{K}_o$:	observation direction
$\vec{F}$:	facet normal
p:	polarisation (***H*** or ***V***), the incident plane being defined as containing the normal to an imaginary flat sea surface and $\vec{K}_o$
d:	Cox and Munk slope distribution (windspeed dependent)

$\Gamma(\,)$: facet reflectivity in the (H, V) reference axes

$T_{DN}(\,)$: downwelling brightness temperature reflected along $\vec{K}_o$.

T_s: sea surface temperature

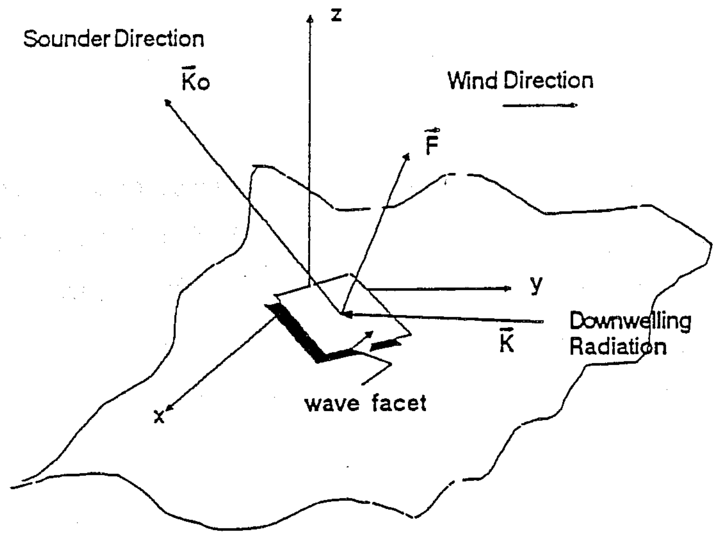

Fig. 3: Sea surface geometry

The brightness temperature of the sea surface at sea level is then obtained by integrating equation (22) over the upper hemisphere (UH) i.e. omitting variable dependences

$$\int_{UH} P(1-\Gamma)\, T_s\, d\Omega + \int_{UH} P\, \Gamma\, T_{DN}\, d\Omega \tag{23}$$

where $d\Omega$ is the solid angle element.

The weighting function P must obviously satisfy the normalisation relation

$$\int_{UH} P\, d\Omega = 1 \tag{24}$$

Inserting (24) in (23) leads to

$$(1-\int_{UH} P\, \Gamma\, d\Omega)\, T_s + \int_{UH} P\, \Gamma\, T_{DN}\, d\Omega \tag{25}$$

Comparing (25) with (1) and (5) the rough sea emissivity is given by

$$e_s = 1 - \int_{UH} P\, \Gamma\, d\Omega \tag{26}$$

and the scattered contribution becomes

$$T_{sca} = \int_{UH} P\, \Gamma\, T_{DN}\, d\Omega \tag{27}$$

Equation (27) shows that the contribution of the surface cannot be simply written as the product of the surface reflectivity by the downwelling radiation along only one direction, as it was done for a specular surface.

The integrals of (25) are numerically computed using discrete summations using 51 x 51 points to sample the 2-D slope distribution within $\pm$ 2.5 times the standard deviations. The facets which are not viewable from the satellite (i.e. $\cos(\vec{K}_o,\vec{F}) < 0$) are not taken into account in the summation (shadowing effect). If the downwelling radiation Tb_{DN} comes from below the horizon, Tb_{DN} is set equal to the sea surface temperature. For all others $\vec{K}$ direction, Tb_{DN} is computed using a linear interpolation of pre-computed Tb_{DN}'s against the incidence angle, as explained in (14).

Although sea foam may have a large effect on sea brightness temperature for wind speed greater than 10 m/s, its effect has not been taken into account because sea foam emissivity and coverage dependency are difficult to model. The reader will find some useful references on sea foam brightness temperature in *Smith* (1988).

5.3 Land surfaces

The modelling of land microwave brightness temperatures is a very difficult task. The 3-D inhomogeneity of the dielectric constant, the variability of surface types, the roughness are some examples of the problems one has to deal with for land surface Tb's modelling. Once again, a very good introduction will be found in *Ulaby et al.* (1981, 1986). We have assumed that the land surfaces are Lambertian (see 2.2.2). This assumption is known to be valid for microwave frequencies when the surface is very rough as for rain forests. The user of the code has to specify the albedo of land surface.

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