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Modeling the radiometric signatures of the earth from space: a tool to study the performance of new radiometers.

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ABSTRACT

In the last years two new kinds of microwave radiometers are being studied for Earth observation: aperture synthesis interferometric radiometers and polarimetric radiometers. The first ones are formed by an array of small antennas whose outputs are cross-correlated and then, properly processed to obtain a map of the apparent brightness temperature of the whole scene being imaged. One- and two- dimensional systems have been studied by some space agencies, e.g. ESTAR by NASA, and MIRAS by ESA, as a solution that avoids the implementation of large steerable antennas at low frequencies (L-band), while reaching a relatively high spatial resolution: about 20-30 Km. More recently preliminary studies of mm-wave systems have also been studied to improve the spatial resolution achieved by today's radiometers. On the other hand, polarimetric radiometers are formed by a dual-polarization antenna. The real and the imaginary parts of the complex cross-correlation computed from the H/V outputs leads to the third and fourth Stokes parameters of the incoming thermal radiation, which are basically related to roughness state of the surface being imaged. At present, a number of studies are being conducted to establish the relationship with the wind direction over the sea surface. The performance analysis of those systems requires the modeling of the apparent brightness temperature map of the Earth and/or sea surface that would be imaged at the microwave and the mm-wave frequencies, which is the object of this paper.

Keywords: radiometry, polarimetry, brightness temperature, imaging, simulator.

1. INTRODUCTION

The modeling of the vertical and horizontal brightness temperatures of the Earth surface and their relationship with geophysical parameters is today an open issue, specially for vegetation and snow covered surfaces at high frequencies. On the other hand, the behavior of the sea surface roughened by the wind, as well as that of the atmosphere are much better understood, and global wind maps over the ocean, rain intensity maps, water vapor maps ... are daily generated from SSM/I measurements. Recently, numerical techniques have also been developed to analyze the polarimetric emission behavior of the sea surface.

In addition, a detailed modeling of the radiometric signatures at a global scale would require a complete data base of a number of geophysical parameters which are not available at present, and the available ones have a spatial resolution worse than required. In the apparent brightness temperature generator that is presented in this paper, vertical and horizontal brightness temperatures are computed from the following physical parameters: soil and snow albedos, snow depth, soil roughness, vegetation albedos, soil moisture, soil surface temperature, ocean salinity, zonal and meridional winds over the oceans, vegetation height, ocean surface temperature and ocean ice cover for the earth and sea surfaces, while the atmosphere upwelling, downwelling and attenuation are computed for a standard atmosphere in which the effect of rain intensity and water vapor concentration maps have been included. These parameters correspond to December 1998, and most of them have been extracted from the CD set: "Global Data Sets for Land-Atmosphere Models ISLSCP Initiative 1: 1987-1988 Volumes 1-5" of the NASA Goddard Space Flight Center^{1,2}. The data contained in this CD set have been acquired

from a variety of sources, including model outputs, satellite and ground measurements and are mapped in a common spatial resolution: a $1^\circ \times 1^\circ$ grid, with a monthly temporal resolution. However, some parameters are given with a 6 hour temporal resolution, for which the December 18 data have been selected. The $1^\circ \times 1^\circ$ grid corresponds to a pixel's size of 110 Km x 110 Km over the equator, which is larger than the spot over the Earth's surface of any of the above mentioned sensors. To overcome this problem the resulting brightness temperature pixels have been bilinearly interpolated to a thinner $1/12^\circ \times 1/12^\circ = 9.26 \text{ Km} \times 9.26 \text{ Km}$ grid given by the NOAA ETOPO5 5 minutes resolution global digital elevation model, which preserves the high frequency content given by the coast line (figure 1).

Consequently, the main limitations of this software are: 1) the limited validity of some models for certain frequency ranges and/or due to the lack of some geophysical parameters, and 2) the limited spatial resolution of the data available at a global scale.

Figure 1. Effect of interpolation to a $1^\circ \times 1^\circ$ mask from a $5^\circ \times 5^\circ$ mask.

Figure 2. Geometry for the incidence angle computation.

The apparent brightness temperature T_{AP} at p polarization is computed from the following general equations (ch 4³):

$$T_{AP}(\theta, \phi) = T_{UP, atm, p}(\theta, \phi) + \frac{1}{L_{atm, p}} \left(T_B(\theta, \phi) + T_{SC, p}(\theta, \phi) \right) \quad (1)$$

$$T_B(\theta, \phi) = e_p(\theta, \phi) T_{ph}(\theta, \phi); \quad e_p(\theta, \phi) = 1 - \frac{1}{4\pi \cos \theta} \iint_{2\pi} [\sigma_{pp}^0(\theta, \phi; \theta_p, \phi_p) + \sigma_{pq}^0(\theta, \phi; \theta_p, \phi_p)] d\Omega_i \quad (2)$$

$$T_{SC, p}(\theta, \phi) = \frac{1}{4\pi \cos \theta} \iint_{2\pi} [\sigma_{pp}^0(\theta, \phi; \theta_p, \phi_p) + \sigma_{pq}^0(\theta, \phi; \theta_p, \phi_p)] T_{DN, p}(\theta_p, \phi_p) d\Omega_i \quad (3)$$

$$T_{DN, p}(\theta, \phi) = \frac{T_{sky}}{L_{atm, p}} + T_{DN, atm, p}(\theta, \phi) \quad (4)$$

where $T_{UP, atm, p}$ and $T_{DN, atm, p}$ are the atmospheric upwelling and downwelling brightness temperatures at p -polarization (to account for the different rain attenuation), $L_{atm, p}$ is the atmospheric attenuation, T_{sky} is the sky radiation (galactic and cosmic noise), T_B is the brightness temperature of the pixel towards the pencil beam antenna is pointing, which is related to the surface emissivity $e_p(\theta, \phi)$ and the physical temperature of the pixel, and $T_{SC, p}$ is the total downwelling temperature reflected towards the antenna. Both the emissivity and the scattered temperature $T_{SC, p}$ is related to the bistatic scattering coefficients σ_{pp}^0 and σ_{pq}^0 .

Note that in the above equations, for each pixel, all the parameters have to be computed for each polarization at an incident angle that depends on the altitude of the platform where the radiometer is, on its orientation: $\theta_{inc} = \arccos(\text{PS.EP}/\text{IPSL.EP})$ (figure 2), as well as the orientation between boresight and the wind direction for the computation of the third Stokes parameter (section 4). In the next sections the different terms appearing in equations (1) to (4) are computed and examples of simulated apparent brightness temperatures from the space, at different frequencies are presented.

2. MODELING THE UP-WELLING AND DOWN-WELLING ATMOSPHERIC RADIATION

It is known that the amount of noise introduced by an attenuator at each port is given by:

$$N_{att} = k_B \left(1 - \frac{1}{L_{att}} \right) T_{ph} B \quad (5)$$

where L_{att} are the attenuator losses, T_{ph} is its physical temperature, and k_B and B are the Boltzmann constant and the system's noise bandwidth. If the atmosphere were an attenuator with losses L_{atm} at a uniform temperature T_{ph} , the upwelling and downwelling temperatures would be given by: $T_{UP,atm} = T_{DN,atm} = (1 - 1/L_{atm}) T_{ph}$. However, each atmospheric layer acts as an attenuator at a different temperature, whose attenuation depends on the gas composition, mainly water-vapor concentration, and the presence or not of hydrometeors, specially rain. The atmospheric upwelling and downwelling emitted radiation (eqns. 1 and 4) at a height H can then be computed as:

$$T_{UP,atm}(\theta, H) = \sec\theta \int_0^H k_a(z') T(z') e^{-\tau(z', H) \sec\theta} dz' \quad (6)$$

$$T_{DN,atm}(\theta) = \sec\theta \int_0^\infty k_a(z') T(z') e^{-\tau(0, z') \sec\theta} dz' \quad (7)$$

$$\tau(z_1, z_2) = \int_{z_1}^{z_2} k_a(z) dz \quad (8)$$

where $k_a(z)$ is the absorption coefficient at the height z , and can be decomposed in three terms: the contribution of the different gas constituents, k_{gas} , and eventually the extinction coefficients of clouds and rain precipitation, $k_{ext,clouds}$ and $k_{ext,rain}$.

$$k_a = k_{gas} + k_{ext,clouds} + k_{ext,rain} \quad (9)$$

Except for water-vapor variations and rain, the relative composition of the atmosphere is rather constant up to 90 Km above the sea level. For mid-latitudes, the 1962 US Standard Atmosphere gives a generalized model of the vertical structure of the earth's atmosphere: temperature, density, pressure and water-vapor density profiles (ch. 5³). Below 100 GHz the absorption due to gas constituents is mainly contributed by water-vapor and oxygen, around 22 and 60 GHz respectively.

Water clouds extend roughly up to 10 Km, and the attenuation coefficient can be computed from Benoit's empirical formula⁴:

$$k_{1,ext,clouds} = f^{1.95} e^{-6.866 (1 + 0.00457)} \quad (10)$$

where $k_{1,ext,clouds}$ is given in (dB/Km) / (g/m³) and f in GHz. The height distribution of the total water-vapor of the clouds⁵ is shown in figure 3a. Due to the lack of global data on ice clouds, they have not been included in our simulator. The vertical distribution of rain in a cell⁵ roughly extends up to 4 Km (figure 3b) and a logarithmic model (ch. 5³) is used the extinction coefficient for V/H polarizations vs. height at each frequency.

Figure 3a. Cloud water-vapor distribution (%) vs. height.

Figure 3b. Rain distribution (%) vs. height.

Figure 4. $T_{em up}$ and $T_{em dn}$ for $\theta = 0^\circ$ and rain rates 0 (-) and 20 mm/h (- -).

Figure 4 shows the upwelling and downwelling atmospheric temperatures vs. frequency for the standard atmosphere, a zenith angle of 0° , and two rain rates: 0 mm/h (continuous line) and 20 mm/h (dashed line). Note the peak near 22 GHz due to the water vapor absorption, and the oscillations in the upwelling brightness temperature near 60 GHz, due to the oxygen resonance lines.

3. MODELING THE EMISSION BEHAVIOR OF THE LAND

As commented in the introduction three different cases have been considered: bare soil, soil covered by vegetation, and soil covered by snow. Since most of current models for vegetation and snow covered soils have a limited frequency range of validity, and require very detailed information, which is usually not available, simplified models have been adopted and some parameters, e.g. the albedos, have been adjusted to fit modeled brightness temperatures with reported experimental data.

3.1. BARE SOIL

Bare soil emission $T_{B,p}^{bare}(\theta) = e^{\epsilon_r \sigma} T_{ph}$ is modeled⁶ with the following emissivities $e_p(\theta)$ (eqn. 2):

$$\begin{aligned} e_p^{bare}(\theta) &= 1 - \Gamma_p(\theta) \\ \Gamma_p(\theta) &= \left[(1-Q) \Gamma_p^p(\theta) + Q \Gamma_p^q(\theta) \right] e^{-4k^2 \sigma_{\text{rms}}^2 \cos^2 \theta} \\ Q &= 0.35 \left(1 - e^{-0.6 \sigma_{\text{rms}}^2 f_{\text{cm}}} \right) \end{aligned} \quad (11)$$

where $\Gamma_p(\theta)$ is the effective reflection coefficient, $\Gamma_p^p(\theta)$ is the Fresnel reflection coefficient at p -polarization, Q is a polarization-mixing parameter, and σ is the rms height of the surface.

The emissivity dependence on the soil moisture (SM) is introduced through the dielectric permittivity of the soil used in the Fresnel reflection coefficient. In addition to the soil moisture data, mixing models⁷ require data about the sand and clay fractions, the soil bulk and solid soil material densities and the soil dielectric constant. In our simulator the empirical MAC-HYDRO '90 model has been used⁸ which is reported to agree within a 10% with the Dobson mixing model⁷ for the soils samples of that campaign.

$$\epsilon_r = 2.583 e^{0.0612M} \quad (12)$$

3.2. VEGETATION COVERED SOIL

Vegetation covered soil is modeled by means of the three layer approach described by Ulaby et al. (ch. 4³), which simplifies to the usual models, whenever the air-vegetation reflection coefficient is negligible⁹ ($\Gamma_{\text{air-veg}} \approx 0$):

$$T_B^{canopy}(\theta) = \frac{1 - \Gamma_{\text{air-veg}}}{1 - \frac{\Gamma_{\text{air-veg}} \Gamma_{\text{veg-soil}}}{L_{\text{veg}}^2}} \left[\left(1 + \frac{\Gamma_{\text{veg-soil}}}{L_{\text{veg}}} \right) \left(1 - \frac{1}{L_{\text{veg}}} \right) (1-a) T_{\text{veg}} + \frac{1 - \Gamma_{\text{veg-soil}}}{L_{\text{veg}}} T_{\text{soil}} \right] \quad (13)$$

The computation of reflection coefficients $\Gamma_{\text{air-veg}}$ and $\Gamma_{\text{veg-soil}}$ uses a quadratic model for the effective dielectric permittivity of the vegetation¹⁰, with Mätzler's model for the dielectric constant of the leaves¹¹:

$$\epsilon_{\text{veg}} = \left(1 + vol \cdot (\sqrt{\epsilon_{\text{leaves}}} - 1) \right)^2 \quad (14)$$

$$\epsilon_{\text{leaves}} = 0.522 (1 - 1.32 m_d) \epsilon_{sw} + 0.51 + 3.84 m_d \quad (15)$$

where vol is the volumetric fraction of vegetation, m_d is the dry mass fraction (≈ 0.3), and ϵ_{sw} is the dielectric constant of saline water, approximately 1%. The vegetation attenuation is computed by means of eqns. 14 and 15, taking into account the transmission angle and the rms height of the vegetation and vegetation type data available at a global scale.

Since the data corresponding to the fractional area of each pixel covered by vegetation (C) is not available the following mixing formula³ can not be applied:

$$T_B(\theta) = [1 - C(\theta)] T_B^{bare} + C(\theta) T_B^{canopy} + [1 - C(\theta)] T_B^{interact \text{ veg-bare}}(\theta) \quad (16)$$

3.3. SNOW AND ICE COVERED SOIL

Snow covered soil is modeled in a similar way as the vegetation covered soil (eqn. 13), by interchanging the subscripts veg by $snow$, with the following considerations: i) since the liquid-water content data is not available, snow is assumed to be dry, ii) the dielectric constant of the dry snow is computed for $f < 13$ GHz with the formulas provided by A. Fung¹² (eqn. 17), iii) soil is assumed¹² to be frozen at a temperature of -1°C , iv) its dielectric constant is approximately 3, v) the variations of the physical temperature and the snow density inside the snow layer have an exponential variation¹³.

$$\epsilon'_{dry\ snow} = (1 + 0.51 \rho_s)^3 \quad (17a)$$

$$\epsilon''_{dry\ snow} = 3 v_{ice} \epsilon''_{H_2O} \frac{\epsilon'^2_{dry\ snow} (2\epsilon'_{dry\ snow} + 1)}{(\epsilon'_{H_2O} + 2\epsilon'_{dry\ snow}) (\epsilon'_{H_2O} + 2\epsilon'^2_{dry\ snow})} \quad (17b)$$

where $\epsilon'_{dry\ snow}$, $\epsilon''_{dry\ snow}$, ϵ'_{H_2O} and ϵ''_{H_2O} are the real and imaginary parts of the dielectric constant of dry snow and pure water, ρ_s is the snow density and v_{ice} is the ice volume fraction.

Ice covered soil is computed in a similar way, but the dielectric constant of the ice is used instead¹²

$$\epsilon'_{saline\ ice} = 3.05 + 0.0072 v_b (S_p T_i) \quad (18a)$$

$$\epsilon''_{saline\ ice} = 0.024 + 0.0033 v_b (S_p T_i) \quad (18b)$$

where $\epsilon'_{saline\ ice}$ and $\epsilon''_{saline\ ice}$ are the real and imaginary parts of the dielectric constant of saline ice, and v_b is the brine volume fraction that depends on the sea ice salinity S_i and the temperature T_i .

4. MODELING THE POLARIMETRIC EMISSION BEHAVIOR OF THE SEA

The emissivity of an irregular surface over the XY plane, at H/V polarization observed from the direction (θ, ϕ) is related to its scattering properties¹⁴.

$$\epsilon_i(\theta, \phi) = 1 - \frac{1}{4\pi \cos \theta} \int_{2\pi} \cos \theta_i \left[\gamma_{mm}(\theta, \phi, \theta_p, \phi_p) + \gamma_{vv}(\theta, \phi, \theta_p, \phi_p) \right] d\Omega_i \quad (18)$$

where $\cos \theta_i = \gamma_{pqpq} \Delta \sigma_{pq}^0$ and $p, q = H/V$ (compare eqns. 2 and 18). The extension to the polarimetric case (third and fourth Stokes parameters) was performed by Yueh et al.¹⁵.

$$\epsilon_v(\theta, \phi) = - \frac{1}{4\pi \cos \theta} \int_{2\pi} \cos \theta_i 2 \frac{R}{S} \left[\gamma_{vv}(\theta, \phi, \theta_p, \phi_p) + \gamma_{vv}(\theta, \phi, \theta_p, \phi_p) \right] d\Omega_i \quad (19)$$

where the polarimetric bistatic scattering coefficients $\gamma_{mpq}(\theta, \phi, \theta_p, \phi_p)$ are related to the polarimetric scattering matrix elements $f_{pq}(\theta, \phi, \theta_p, \phi_p)$ -incident field from direction (θ_p, ϕ_p) at q -polarization, reflected field to direction (θ, ϕ) and p -polarization- and the illuminated area.

The computation of $\gamma_{mpq}(\theta, \phi, \theta_p, \phi_p)$ was performed in 1967 Stogryn¹⁶ by means of the *Kirchhoff Method under the Stationary Phase Approximation* for the anisotropic geometry of the sea surface roughened by the wind, allowing different rms slopes along the upwind (Y-axis, $\phi=90^\circ$) and crosswind directions, leading to the following expression for the scattering coefficients:

$$\gamma_{pqpq}(\theta, \phi, \theta_p, \phi_p) = \frac{\sigma_{pq}^0(\theta, \phi, \theta_p, \phi_p)}{\cos \theta_i} = \frac{1}{\cos \theta_i} \frac{k^2 q^2 |U_{pq}|^2}{2 q_x^4 g_x g_y} \exp \left[- \frac{1}{2 q_x^2} \left(\frac{q_x^2}{g_x^2} + \frac{q_y^2}{g_y^2} \right) \right] \quad (20)$$

where the parameters involved are defined in (ch. 18³). It should be pointed out that the U_{pq} terms depend on the Fresnel reflection coefficients for H/V polarizations, that depend on the incidence angle θ_i and the dielectric permittivity of sea water, which is a function of the salinity (S) and the temperature (T)¹⁷. The g_x^2 and g_y^2 factors in eqn. 20 are the variances of the slopes along the upwind and crosswind directions respectively, and are related to the wind speed at 12.5 m height¹⁸. However, Wilheit¹⁹ suggested that: i) for frequencies below 35 GHz, the values of the slope variances given by Cox and Munk¹⁸ must be multiplied by a factor $0.3 + 0.02 f_{(GHz)}$, and ii) the effect of the sea foam induced by the wind was an isotropic reduction of the reflectivity (integral term in eqn. 18) by a factor $K = 0.006(1 - e^{-f_{(GHz)}^{17.5}})(u_{(m/s)} - 7)$, for wind speeds larger than 7 m/s. This reduction of the reflectivity produces an increase of the H/V emissivities (eqn. 18) and consequently, in the H/V brightness temperatures. However, when the same factor is applied to eqn. 19, it produces a decrease of the magnitude of the third and fourth Stokes parameters.

Following a similar development Camps et al.²⁰ extended eqn. 20 to the study of polarimetric case of the sea surface roughened by the wind, leading to:

$$\gamma_{\text{mpq}}(\theta, \phi, \theta_p, \phi_p) = \frac{1}{\cos \theta_i} \frac{k^2 q^2 U_{\text{m}} U_{\text{p}}}{2 q_x^4 g_x g_y} \exp \left[-\frac{1}{2 q_x^2} \left(\frac{q_x^2}{g_x^2} + \frac{q_y^2}{g_y^2} \right) \right] \quad (21)$$

Numerical results computed with the above model have been shown to be in good agreement with experimental results²¹, and that the third Stokes (T_U) parameter has a sinusoidal variation of the form $T_U(\phi) = U_x(\theta, f, u) \sin(2\phi)$, with zero crossings at 90° , 180° and 270° , which may be used to estimate the wind direction, once the wind speed is determined from T_{HV} .

Figure 5. Amplitude of the $T_U(\phi=45^\circ)$, for $f = (+)$ 19, (x) 37 and (o) 85 GHz, and different incidence angles and wind speeds.

Figure 5 shows the maximum value of T_U , at $\phi=45^\circ$, versus the incidence angle with respect to nadir, for three different frequencies 19 GHz, 37 GHz, and 89 GHz, and for three different wind speeds at 12.5 m. According to these results and experimental evidence reported²¹: i) the polarimetric emission behavior is a large bandwidth phenomenon, with weak dependence on the frequency, and a very strong dependence on the wind speed, and ii) the sign of the second harmonic of T_U , $U_x(\theta)$, changes between 45° and 65° , decreasing with increasing wind speeds. Note the incidence angle trade-off for maximum sensitivity on H/V polarizations ($\theta_i \approx 40^\circ-60^\circ$) and T_U ($\theta_i \approx 10^\circ-40^\circ$). In the case of sea partially covered by ice the brightness temperature is computed from a mixing formula that takes into account the fraction of the pixel covered by sea and that by iced sea.

5. SIMULATION RESULTS

Figures 6 to 11 show six different geophysical parameters of interest to understand better the appearance of the brightness temperature maps that are presented. They are: the sea surface salinity and temperature, the wind direction map over the oceans, the surface soil moisture, the snow depth and the mean daily rain rate. These, and the rest of the parameters used correspond to the mean values of December 1998, except whenever the daily data was available.

Figure 6. Sea Surface Salinity.

Figure 7. Sea surface temperature.

Figure 8. Wind direction map.

Figure 9. Surface soil moisture.

Figure 10. Snow depth.

Figure 11. Daily rain rate.

Figures 12 to 17 show 5 five different apparent brightness temperatures scenes as they would be imaged by a dual H/V polarization microwave radiometer at different frequencies and at a fixed position at 4.000 Km altitude.

i) Figures 12 a,b show a 1.4 GHz scene of the Amazon river. Note the large contrast between the brightness temperature of the sea and that of the land, and the increase of the brightness temperature of the sea at the Amazonas mouth due to the lower salinity concentration (figure 6) caused by the large amount of fresh water poured into the sea, close to the Earth border. Note also the decrease/increase of the horizontal/vertical brightness temperature with increasing incidence angles. Snow at the Andes (figure 10) appears with lower brightness temperature. Atmospheric effects at this band are almost negligible. The ESTAR and MIRAS interferometric radiometers, under study, are intended to measure sea salinity and surface soil moisture at 1.4 GHz.

Figure 12a. South America: Brightness temperature 1.4 GHz H-Pol.

Figure 12b. South America: Brightness temperature 1.4 GHz V-Pol.

ii) Figures 13 a,b show the 10.7 GHz V/H brightness temperature of the South East Asia and Australia. Atmospheric effects are also almost negligible at this frequency, but now the effect of sea roughness caused by the wind can be more clearly appreciated. Figure 13c shows the third Stokes parameter (T_U) over the sea for the same scene. First of all, it should be noted the small amplitude of this magnitude, about ± 2 to 3 K, even for high wind intensities and proper orientation of the observation angle and the up-wind direction (figures 5 and 8), which poses severe requirements on the radiometric sensitivity and accuracy required to measure it. However, the sign changes and the zero-crossings of T_U may serve in a futur to determine, not only the wind intensity, which is determined from $T_{H/V}$, but its direction, at a global scale.

Figure 13a. Asia and Australia:
Brightness temperature 10.7 GHz
H-Pol.

Figure 13b. Asia and Australia:
Brightness temperature 10.7 GHz
V-Pol.

Figure 13c. Asia and Australia:
Brightness temperature 10.7 GHz
Third Stokes parameter.

iii) Figures 14 to 18 show the horizontal and vertical brightness temperatures at 10.7, 19, 22, 37 and 89 GHz. From this sequence of figures it is clear that at higher frequencies, more important the atmospheric effects are. For example, the rain storms at the western coast of Africa, at the North-West coast of Spain, and Norway (figure 11), have a clear radiometric signature, which is much more apparent at 89 GHz than at 22 or 37 GHz, being almost unnoticeable at 10.7 GHz. Note also that the effect of high surface soil moisture results in a cooler brightness temperature, since the dielectric constant of the soil has increased, and so does the reflection coefficient, e.g. Northern coast of Africa, center of Spain and central Europe. The effect of the snow cover can be clearly appreciated in the Scandinavian peninsula, and in the Alps. At the Eastern part of Russia this effect is not apparent because it is masked by the variation with the ground incidence angle and the higher atmospheric attenuation.

Figure 14a. Europe Brightness
temperature 10.7 GHz H-Pol.

Figure 14b. Europe Brightness
temperature 10.7 GHz V-Pol.

Figure 15a. Europe Brightness
temperature 19 GHz H-Pol.

Figure 15b. Europe Brightness
temperature 19 GHz V-Pol.

Figure 16a. Europe Brightness temperature 22 GHz H-Pol.

Figure 16b. Europe Brightness temperature 22 GHz V-Pol.

Figure 17a. Europe Brightness temperature 37 GHz H-Pol.

Figure 17b. Europe Brightness temperature 37 GHz V-Pol.

Figure 18a. Europe Brightness temperature 89 GHz H-Pol.

Figure 18b. Europe Brightness temperature 89 GHz V-Pol.

6. CONCLUSIONS

This paper has described the implementation of an apparent brightness temperature generator of the Earth, from an arbitrarily located and oriented platform. Examples of the generated images have been presented for several frequencies and regions of the world, showing out the different radiometric signatures of the apparent brightness temperature at different frequencies on several geophysical and atmospheric parameters, that may be susceptible of being measured remotely from space. The main limitations of the implemented brightness temperature generator are: i) the lack of data at a global scale of the different parameters required to model with more detail the brightness temperatures signatures, ii) the reduced spatial resolution of the available data, which has been bypassed by re-gridding the data into a thinner grid given by the NOAA ETOPO5 digital elevation model, and performing a bilinear interpolation of the computed values, and iii) the lack of accurate emissivity models over the whole microwave and millimeter wave frequency range, specially at higher frequencies. Even these limitations may limit the radiometric accuracy of some models at certain frequencies, this brightness temperature generator is a useful tool to study new radiometric systems, such as L-band interferometric radiometers (e.g. MIRAS), new ones under study at mm-wave frequencies, and new polarimetric radiometer systems devoted to the determination of wind speed and direction over the ocean surface.

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