IMPROVED MODELING OF THE MICROWAVE RADIOMETRIC RESPONSE OF BARE GROUND

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ABSTRACT

The heat radiation of the ground as detected by a microwave radiometer depends both upon the emissivity and the temperature of the ground. Earlier modeling has mainly been focused on the emissivity dependence and its relation to the ground characteristics. In this paper, an extended model is presented, which also takes into account the temperature and soil-moisture profiles in the top-layer and how they depend on soil and weather parameters. Besides predictions of the microwave emission of the ground, the model also gives an improved understanding of the influence of wavelength, weather effects, time of acquisition, and the electromagnetic, thermal and hydraulic characteristics of the ground material.

1. INTRODUCTION

The thermal radiation of the ground as detected by a microwave radiometer is mainly controlled by the emissivity and the temperature of the ground surface. At high frequencies, in particular, the atmospheric transmission and reradiation have also significant effects. Most work on microwave radiometric emission of ground has primarily been focused on the emissivity and atmospheric influence. Usually, a constant surface temperature with no depth-variation is assumed at the interpretation of the data. This simplified approach may be satisfactory over areas with high emissivity fluctuations due to strong variations in soil-moisture. If weaker effects are to be studied, however, emissivity variations of the order of one or two per cent should be detectable, which corresponds to radiometric temperature contrasts of only a few degrees. Obiously, an improved model is then required in order to separate emissivity and temperature effects. Measurements and analysis by Njoku [1] and others have shown the need for improved modeling of the temperature influence upon the microwave radiation of the ground.

In this paper, an extended model is described, which besides emissivity also predicts the temperature of the upper soil-layer. The temperature profile is computed from a finite-difference model of the heat-diffusion equation with the heat-exchange at the ground-surface as a boundary-value. Inputs to the model are the electromagnetic and thermal characteristics of the ground material, and the weather conditions during the time-period before the acquisition of data. For moisty soils, the model also describes the moisture profile and the heat and water transfer at the ground-air interface. Specific submodels are also discussed, which translate porosity, soil-moisture to thermal and electromagnetic characteristics.

2. MICROWAVE EMISSION

2.1 Introduction

Usually, the black-body temperature of the outgoing microwave radiation is modeled by

$$T_B = \varepsilon_s \, T_s + (1-\varepsilon_s) \, T_a \tag{2.1}$$

where ε_s is the surface emissivity, T_s is the ground temperature, and T_a is the radiation temperature of the incident sky radiation.

A more accurate model, which also takes into account the semi-transparent characteristics of the top-layer and its temperature variation with depth, $T(z)$, is given by

$$T_s = \int_0^\infty \alpha(z) \, T(z) \, \exp\left(-\int_0^z \alpha(x) \, dx\right) dz \tag{2.2}$$

where $\alpha(z)$ represents the variation with depth of the microwave absorption factor at the actual frequency.

2.2 Emissivity

The emissivity of bare soils of varying water content is typically in the range 0.6-0.95, which means temperature variations in the range 170-300 K when the atmosphere is high-transparent. At high frequencies, or when air-born measurements are carried out below a thick cloud-cover, temperature contrasts on the order of 5-20 K are more typical. For a smooth soil-surface the emissivity of a smooth homogeneous soil-layer can be described as

$$\varepsilon_s(\Theta) = 1 - R_s(\Theta) \tag{2.3}$$

where $R_s(\Theta)$ is equal to the power Fresnel reflection coefficient at incidence Θ and defined by the complex dielectric constant of the soil-layer.

2.3 Surface roughness effects

For rough surfaces, an approximate model has to be used [2], [3]

$$\varepsilon_s = 1 - R_c - R_c \tag{2.4}$$

where R_c is the reflection due to coherent scattering in the specular direction [3]

$$R_c = R_s(\Theta) \exp\left(-h_a \cos^2\Theta\right) \tag{2.5}$$

with $h_a = (4\pi\sigma/\lambda)^2$ and σ^2 is the variance of surface roughnesses. The incoherent component R_i is also related to the correlation length of the roughness (L) and has to be computed numerically e.g. [3]. A one-parameter approximation of the total reflectivity of the rough surface is [4]

$$R = 1 - \varepsilon_s = R_s(0) \exp(-h) \tag{2.6}$$

where h is an effective roughness parameter ($h < h_a$).

2.4 Absorption factor

The absorption factor of soil depends upon soil-type, water content and porosity. For homogeneous materials the absorption factor is defined by the imaginary part of the complex dielectric constant of the material

$$\varepsilon = \varepsilon' - j\,\varepsilon'' \qquad (2.7)$$

as follows

$$\alpha = \frac{2\pi}{\lambda_o} \operatorname{Im}\{\sqrt{\varepsilon}\} \qquad (2.8)$$

For mixtures, no general theory exists. For a mixture of water inclusions in a host material, with the water modeled as ellipsoidal discs that are randomly distributed, both spatially and with respect to the angular orientation, Loor [5] derived the following formula of the complex dielectric constant of the mixture

$$\varepsilon_m = \varepsilon_h + \frac{V_i}{3}(\varepsilon_i - \varepsilon_h)(2 + \varepsilon_m/\varepsilon_i) \qquad (2.9)$$

where ε_h, ε_i are complex dielectric constants of the host material and the inclusions respectively, and V_i is the volume fraction of the inclusions.

Mostly, the imaginary part of the host material can be taken close to zero, since the water content of semi-dry soils gives a dominating contribution to the loss-factor. The complex dielectric constant of water can be estimated from the Debeye expression

$$\varepsilon = \varepsilon_\infty + \frac{\varepsilon_s - \varepsilon_\infty}{1 + j2\pi f} - j\,\frac{\sigma}{2\pi f \varepsilon_o} \qquad (2.10)$$

where f is the frequency (Hz), ε_∞ is the dielectric constant at infinite frequency, ε_s is the static dielectric constant, τ is the relaxation time(s), σ is the ionic conductivity (mho/m) and $\varepsilon_o = 8.854 \cdot 10^{-12}$ (F/m) is the permittivity of free space. The parameters ε_∞^o, ε_s and σ depend on temperature and salinity. Models showing these relations are described in [6].

Examples of other types of mixture formulas can be found in [7]. The following multi-component relationship was also suggested

$$\sqrt{\varepsilon_m} = \sum_{i=1}^{n} W_i \sqrt{\varepsilon_i} \qquad (2.11)$$

where W_i, ε_i are the volumetric portion and dielectric constant, respectively, of the i:th component of the mixture. An empirical model of the complex dielectric constant of soils as a function of soil-moisture content was developed by Wang and Schmugge [8]. According to this model, the real part of the relative dielectric constant of four different soils at 5 GHz starts from about $\varepsilon_r \approx 3$ for dry soil and increases near-linearly to 11-15 for a volumetric moisture content of 30 per cent. The imaginary part starts at zero (dry soil) and is close to 2 at 30 per cent soil-moisture.

3. THERMAL MODELS

Introduction

The temperature profile of the top-layer of the soil is controlled by the heat transfer at the surface boundary. If a one-dimensional model is assumed, the temperature U(x,t) and the heat flow Q(x,t) are determined by

$$Q = -\lambda\frac{\partial U}{\partial x} \tag{3.1}$$

$$-\frac{\partial Q}{\partial x} = \frac{d}{dt}(CU) \tag{3.2}$$

where t is time, x is depth, λ is the thermal conductivity, and C is the heat capacity per unit volume.

The boundary condition at the surface is determined by the heat balance equation

$$Q_o = -\lambda\frac{\partial T}{\partial x}\bigg|_{x=0} = I(1-A)-\varepsilon\sigma(T^4-T_{sky}{}^4)-H-q_e-q_s \tag{3.3}$$

where T is the surface temperature, I is the short-wave irradiance, A is the albedo of the ground surface, ε is the long-wave emissivity, σ is the Stefan-Bolzmann constant, T_{sky} is the long-wave sky-temperature, H is the sensible heat flux, q_e is the latent heat flux, and q_s is the geothermal heat flux.

The solar irradiance I(t) in (3.3) is for horizontal surfaces

$$I(t) = \tau_s I_s(t) + I_d(t) \tag{3.4}$$

in which $I_s(t)$ denotes the irradiance at the top of the atmosphere, τ_s is the transmittance, and $I_d(t)$ is the diffuse sky component. The irradiance $I_s(t)$ is well-defined when latitude, solar declination, local time, and surface elevation are known. Due to variations of τ_s and I_d, however, I(t) has to be estimated from ground-based measurements.

Sensible heat flux

The sensible heat flux of (3.3) can be described by

$$H = k_o(T-T_a) \tag{3.5}$$

where T_a is the air-temperature. The factor k_o is related to the aerodynamic diffusion resistance r_a of the rough surface boundary

$$r_a = \rho c_p/k_o \tag{3.6}$$

in which ρ is the air density, and c_p is the specific heat of dry air at constant pressure. The factors k_o and r_a are highly influenced by the local wind-speed (w), but also the surface roughness and the atmospheric stability have significant effects. An approximation, which is often used under conditions of neutral stability, is

$$r_a = \frac{\ln^2((z-d)/z_o)}{k^2 w} \tag{3.7}$$

In (3.7), k is the von Karman constant (0.35), d is the zero displacement, and z is the height level, where wind velocity and air temperature are measured. The parameter z_0, which represents the effective roughness parameter for water vapour and heat exchange, is usually one order lower than the real surface roughness, e.g. 0.0003 m for sand and 0.001 m for smooth fields.

More general relationships, which also include stable and unstable conditions can be found in e.g. [9]-[10].

Latent heat flux

For moisty evaporating surfaces, the latent heat flux q_e has a very significant effect upon the surface temperature. Like the sensible heat transfer, q_e is difficult to model with high accuracy. A commonly used expression is

$$q_e = Mk_0(L/c_p)(X_s(T)-h_a X_s(T_a)) \qquad (3.8)$$

in which M represents the ratio of the actual evaporation to the potential one, L is the latent heat of vaporization, X_s (T) is the saturation mixing ratio, and h_a is the relative humidity of air.

The factor M, which controls the evaporation rate can be expressed as a ratio

$$M = r_a/(r_a+r_g) \qquad (3.9)$$

where r_a is the aerodynamic diffusion resistance, and r_g describes the resistance of the soil layer to water transport.

For a wet surface saturated by water, r_g is low, giving $M \approx 1$. When the surface dries up, M is reduced as a result of the increased r_g-value.

Finite difference model

The surface temperature can be computed from Eqs. (3.1)-(3.3) under very general conditions using the method of finite differences [11]. The upper ground surface layer is then subdivided into thin horizontal layers (Δx) to a depth of D. Before the computer processing starts, the initial temperatures of the layers and all parameters involved have to be defined. Temperatures and heat flows at time $t = (m+1)\Delta t$ are then updated from Q and U at time $t = m\Delta t$, using (3.1)-3.2) expressed in terms of differences and (3.3) as boundary condition.

A great advantage of the finite difference model is the fact that it can be used under very general conditions, for instance, in-homogeneous soil layers and time-varying weather parameters. The ultimate accuracy of the algorithms primarily depends on how well start conditions, weather data and thermal characteristics of the ground can be defined [12].

Main factors, which influence the temperature of a bare soil surface, are the actual to potential evaporation ratio (M) and the thermal inertia ($P = \sqrt{\lambda C}$) of the ground. Also the albedo (A), surface roughness (z_0), surface slope, air-temperature, dew-point and wind-speed have significant effects, however. The thermal inertia is influenced by the mineral content of the soil, water-content and porosity. Sub-models which describe the thermal characteristics of mixtures have been formulated by de Vries [13].

The heat capacity per unit volume (C) can be expressed as the average of the heat capacity of the various components. If we only distinguish between air, water, minerals and organic matter:

$$C = 4.19 \cdot 10^6 (0.46\ X_m + 0.60\ X_o + X_w)\ [J/m^3 K] \tag{3.10}$$

where X_w, X_m and X_o represent the volume fractions of water, minerals and organic matter, respectively.

The model of the thermal conductivity of mixtures is less straight-forward, since the conductivity is also influenced by the shape and the orientation of the granules. A model suggested by de Vries [13] expresses the conductivity as a weighted average:

$$\lambda = \sum_{i=1}^{N} K_i X_i \lambda_i / \sum_{i=1}^{N} K_i X_i \tag{3.11}$$

where N is the number of the soil components; each with different conductivity (λ_i) and volume fractions (X_i). The weighting factor K_i is for spheroidal particles determined by:

$$K_i = \frac{2}{3[1+(\lambda_i/\lambda_o - 1)g_i]} + \frac{1}{3[1+(\lambda_i/\lambda_o - 1)(1-2g_i)]} \tag{3.12}$$

where λ_o is the conductivity of the medium.

The shape factor g_i has to be calibrated by experiments for different soil types. For sand de Vries found g_i=0.141. In [12] this value was applied for the organic, mineral and water components. The shape-factor of the air-inclusions was modeled by

$$g_a = 0.013 + 0.3(X_w/(X_a + X_w))^2 \tag{3.13}$$

4. WATER PROFILE

The soil-moisture profile of the ground can be predicted in similar way as the heat flow and temperature profile, if precipitation history, the atmospheric water exchange and the hydraulic soil-characteristics are known. For vertical flows and unsaturated soil-water conditions

$$\frac{\partial}{\partial z}\left(K \frac{\partial \psi}{\partial z}\right) - \frac{\partial k}{\partial z} = \frac{\partial \Theta}{\partial t} \tag{4.1}$$

where Θ is the water content (by volume fraction), K is the hydraulic conductivity, and ψ is matrix potential [14]. The process is controlled by the water exchange at the surface boundary due to precipitation and evaporation.

By chain rule of differentiation (4.1) can be written as

$$\frac{\partial}{\partial z}\left(D \frac{\partial \Theta}{\partial z}\right) - \frac{\partial K}{\partial z} = \frac{\partial \Theta}{\partial t} \tag{4.2}$$

where D is the soil water diffusivity defined by the ratio of the hydraulic conductivity to the water capacity of the soil

$$D = K(\Theta)/\frac{\partial \Theta}{\partial \psi}. \tag{4.3}$$

A number of models have been developed for defining relationships between D, K and the structure of a soil-layer e.g. [14], [15].

As shown in [16], Eq. (4.1)-(4.2) can readily be extended to also include the vapour flow and the influence of the temperature profile on the diffusion process.

5. CONCLUDING REMARKS

The main-structures of a simulation model have been described for the prediction of microwave radiation of ground and its dependence on weather and soil-characteristics.

The primary soil-factor in the model is the water content but also mineral content, porosity, surface roughness and the temperature profile have significant effects, however. The soil-moisture controls both the emissivity and the thickness of the radiating top-layer. Even the temperature profile is influenced, since both evaporation and thermal inertia increase at high soil-moisture contents.

The results of computer simulations show that the surface temperature of ground is mainly affected by the thermal inertia and the evaporation which both are correlated to the soil-moisture. The thermal inertia controls the amplitude of the diurnal variations of the surface temperature, while the evaporation reduces its mean-value. Also weather parameters (incident radiation, wind-speed, air-temperature) and albedo, surface elevation and roughnesses have a significant influence [11], [12].

Simulation results also show that for attenuation factors of the microwaves (α) on the order of 2-3 m^{-1} (i.e. penetration depths of 0.3-0.5 m), the outgoing radiation of the ground is mainly affected by the mean surface temperature of the day. If the attenuation is significantly increased (due to higher frequency or soil-water content), the diurnal variations of the surface temperature become much important.

At the interpretation of microwave radiometric data, an improved accuracy is possible, if the weather influence is suppressed or compensated for. At soil-moisture estimations, for instance, preliminary approximations of emissivity and water content are obtained from the measured brightness temperature and an initial guess of the ground temperature. From the thermal model, using the actual weather parameters and the preliminary soil-moisture content as inputs and the temperature profile as output, the emissivity and water content estimates can be further improved. The estimation errors are reduced, if also albedo, soil-type, and surface roughness information is available at the data analysis.

A parallel use of multi-frequency radiometers, including thermal-IR and millimeter waves, should increase the possibilities to resolve the moisture profile and surface effects. The diurnal changes of the microwave radiation of ground and its potential use at the data analysis should also be further investigated.

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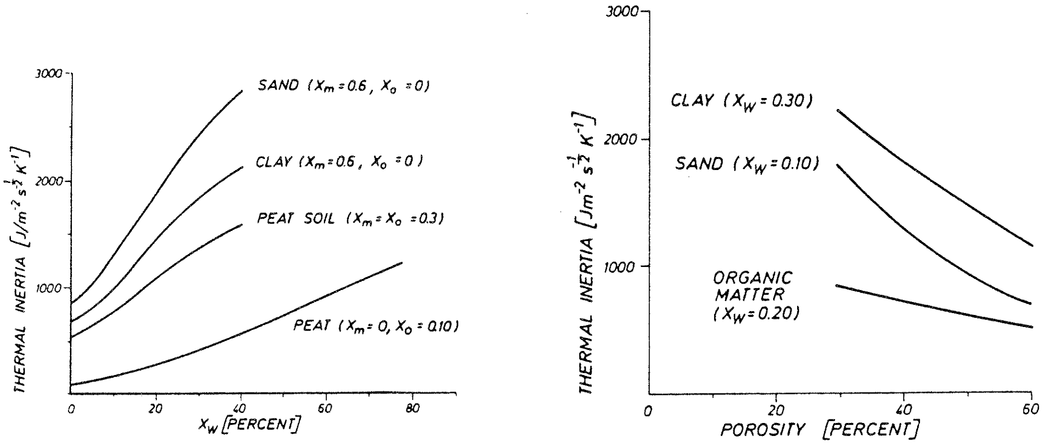

FIGURE 1. Predicted thermal inertia dependence on soil-type, porosity and water content.

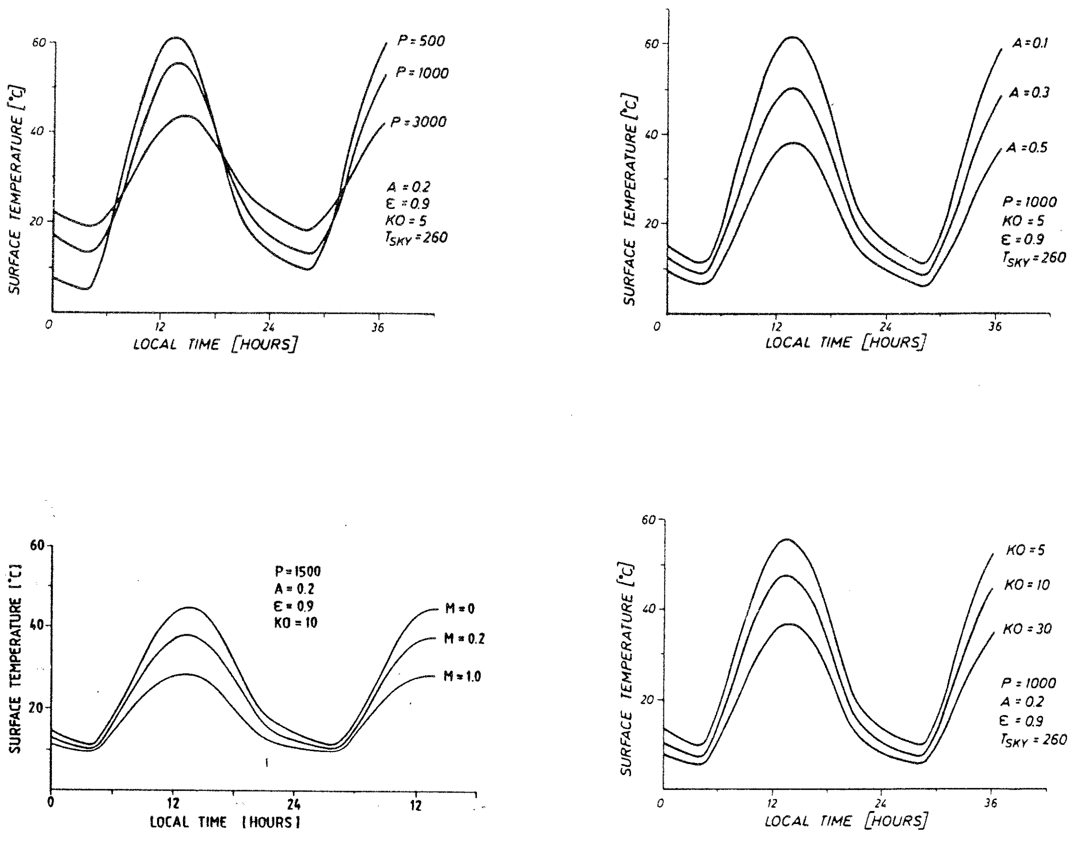

FIGURE 2. Diurnal ground surface temperature predictions for clear sky. Latitude 58°, solar declination 20° and diurnal air-temperature 6 - 27 °C.

TABLE I. Relationships between penetration depth ($d = \alpha^{-1}$), time of acquisition and the efficient surface temperatur [^{o}C] as defined by Eq. (2.2)

$$T_s = \int_0^\infty \alpha\, T(z) \exp(-\alpha z)\, dz$$

The temperature profiles of dry and wet clay soils were predicted using the thermal model for clear sky conditions and low wind-speeds. Solar declination: 23^{o}, latitude: 58^{o}, and air-temperature 11-26^{o} C.

OBJECTS AND PENETRATION DEPTHS	LOCAL TIME			
	06.00	12.00	18.00	24.00
1. Dry clay (P=1000, M=0)				
- d = 0.33 m	24.6	29.5	28.4	24.6
- d = 0.11 m	23.4	33.2	30.0	22.4
- d = 0.04 m	23.1	39.0	30.7	18.6
- d = 0.00 m	24.7	43.2	29.3	15.3
2. Wet clay (P=1800, M=0.5)				
- d = 0.33 m	23.5	26.0	25.4	23.5
- d = 0.11 m	21.6	26.6	24.8	21.0
- d = 0.04 m	20.5	28.6	24.2	17.9
- d = 0.00 m	21.4	30.4	23.2	15.7

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