A Parameterization of Evaporation from Bare Soil Surfaces

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(Manuscript received 14 July 1989, in final form 21 October 1989)

ABSTRACT

A simple model of evaporation from a bare soil surface is developed. This model combines two processes of water vapor transport: one is the vapor transport in air expressed by the bulk formula, and the other is molecular diffusion of vapor in the surface soil pore with the vapor being carried from the interior of the soil pore to the land surface. The resistance to the vapor diffusion in the soil pore is expressed using a new parameter, estimated by experimental means.

General formulation of the so-called "surface moisture availability" is expressed with this model. The formulation shows that the "surface moisture availability" depends not only on volumetric soil moisture, but also on wind velocity, and on the ratio of the specific humidity of the air to that of the saturation value at the soil surface temperature. This dependence agrees with experiments performed with loam and sand under various conditions.

In the evaporation parameterization used in current numerical simulations, the humidity of the air adjacent to the water in the soil pore, which is determined thermodynamically, is often substituted for the land surface humidity. The present study suggests, however, that such a parameterization is invalid except for saturated conditions.

1. Introduction

Surface evaporation is one of the main processes in the air–land energy exchange, but much about this process remains to be solved. The evaporation rate is controlled by atmospheric conditions, surface soil wetness, and moisture transport in the soil layer affected by the soil moisture. For examining the air–land energy exchange, the heat and moisture transport should be estimated simultaneously in the soil and in the atmosphere.

Several numerical simulations, taking the heat and moisture budgets into account, have been made (e.g., Sasamori 1970; McCumber and Pielke 1981; Camillo et al. 1983). In these models, the evaporation process is usually parameterized by the so-called "surface moisture availability." Although there are several ways to express surface moisture availability, it has not been determined which is the most suitable to give a realistic estimate of evaporation.

The two main forms defining surface moisture availability are written as the α-method:

$$E = \rho C_E u(\alpha q^*(T_S) - q), \tag{1}$$

and the β-method:

$$E = \rho C_E u \beta(q^*(T_S) - q), \tag{2}$$

where E is the evaporation rate, ρ the density of air, u the wind speed, C_E the bulk coefficient of evaporation, q the specific humidity of the air, $q^*(T_S)$ the saturation value at the soil surface temperature T_S, while α and β are the coefficients representing the surface moisture availability.

Apparently, α expresses the land surface moisture defined as the air relative humidity at the humidity roughness height Z_q. But the relative humidity of the air adjacent to the water in the soil pore, h, is often used instead of α in Eq. (1). This humidity h is determined through the thermodynamical relationship between the liquid and vapor phases. According to Philip (1957), h can be expressed as

$$h = \exp(\psi g / R T_S), \tag{3}$$

where ψ is the soil water potential at the surface, g the acceleration of gravity, and R the gas constant for water vapor. Several investigators have used this parameterization (Sasamori 1970; McCumber and Pielke 1981; Camillo et al. 1983; Chung and Horton 1987), but h is not always the same as the humidity α at the land surface, which is discussed in section 4.

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Barton (1979) obtained α from field experiment data as a function of volumetric soil moisture content θ. Yasuda and Toya (1981) suggested a similar relationship for $\alpha(\theta)$, and Toya and Yasuda (1988) confirmed the relationship by a lysimeter experiment. It should be noted that their θ is the mean value in the soil between depths of 2.5–5 cm. On the other hand, an evaporation formula using the β-method has been proposed by Kondo (1971) in a boundary layer prediction model, and Nappo (1975) examined the difference between α-method and β-method with calculating diurnal surface latent heat flux in boundary layer models. It is not clear whether surface moisture availability depends on other conditions such as wind speed and the humidity and temperature of the air.

In this study, a simple model of evaporation from a bare soil surface is presented. This model includes a new parameter related to the soil moisture θ of the first 2 cm layer, and allows for a general expression of surface moisture availability. This thickness of 2 cm was chosen to consider the diurnal change of surface moisture. Also, this soil thickness is suitable for direct soil-gathering and practical remote sensing observations. Note that the "bucket model" (e.g., Manabe 1969; Hansen et al. 1983) cannot predict the diurnal cycle of evaporation since it relates the moisture content of a rather thick layer to evaporation (Dickinson and Henderson-Sellers 1988).

2. Model

In an unsaturated soil, water is retained in the form of "free water" and "bound water." The former is retained in the soil matrix by capillary force. The latter is bound firmly to soil particles by intermolecular force or electrical force. Bound water surrounds a soil particle several times as thick as a water molecule. Figure 1 shows a schematic illustration of these conditions.

During the drying process of the soil, the free water retained in large pores vaporizes earlier than that in small pores due to the weaker capillary force on the water in larger pores. As time passes, soil water is left only in the interior of very small pores. The bound water is finally released, with oven drying necessary to remove all the bound water in the soil

The evaporation from soil consists of two processes. In the first process, water vapor is transported by molecular diffusion from the water surface in the soil pore to the "land surface." In the second process, water vapor is carried from the land surface to the atmosphere by laminar or turbulent airflow. Except for a case that water vapor is condensed at the land surface, these two fluxes of each process are equal to the ground surface, and expressed as

$$E = \rho D_{\mathrm{atm}} \frac{q^*(T_S) - q_s}{F(\theta)} \tag{4}$$

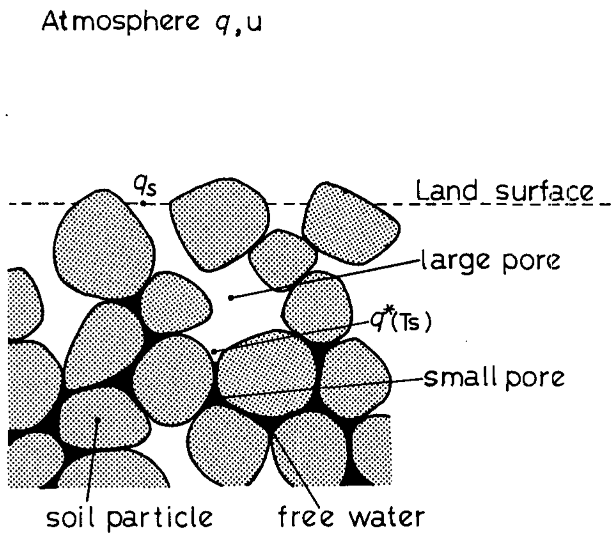

FIG. 1. Schematic illustration of a surface soil layer. Water vapor is transported by molecular diffusion from the soil water surface localized in the pores to the land surface.

for the molecular diffusion, and as

$$E = \rho C_E u(q_s - q) \tag{5}$$

for atmospheric transport. Here, D_{atm} is the molecular diffusivity of water vapor, $q^*(T_S)$ the saturated specific humidity at the surface temperature T_S, q_s the humidity of the air at the land surface (see Fig. 1) and $F(\theta)$ is newly introduced and is a function of the surface volumetric soil moisture θ. The D_{atm} has been formulated by Camillo et al. (1983) as

$$D_{\mathrm{atm}} = D_0(T_S/273.16)^{1.75}, \tag{6}$$

where $D_0 = 0.229 \times 10^{-4}\ \mathrm{m^2\ s^{-1}}$.

In Eq. (4), $q^*(T_S)$ is the specific humidity of the air adjacent to the soil water in the pore, and $F(\theta)$ expresses the resistance to water vapor diffusion from the interior of the soil pore to the land surface. The functional form of $F(\theta)$ should be known in advance for each soil type.

The air adjacent to the soil water actually equilibrates with the water potential of the soil water and its humidity is less than 100%; however, such air is assumed to be saturated in the present model. This is because the soil water is highly localized in the unsaturated soil layer, leading to the complexity in determining the water potential of such soil water from averaged soil moisture θ. The function $F(\theta)$ will be determined by experimental means in section 3.

By eliminating q_s from Eqs. (4) and (5), the evaporation rate is expressed as

$$E = \rho C_E u \frac{q^*(T_S) - q}{1 + C_E u F(\theta)/D_{\mathrm{atm}}}, \tag{7}$$

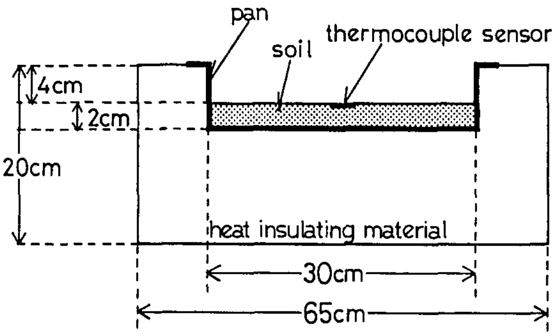

FIG. 2. Illustration of the evaporation pan. The top level of the soil is 4 cm lower than the edge of the pan to prevent soil erosion by the wind.

while α and β are obtained from Eqs. (1), (2) and (7) as follows:

$$\alpha = \frac{q}{q^*(T_S)} + \beta\left(1 - \frac{q}{q^*(T_S)}\right), \quad (8)$$

and

$$\beta = \frac{1}{1 + C_E u F(\theta)/D_{\mathrm{atm}}}. \quad (9)$$

These formulas show that β depends on θ and u, while α depends on θ, u and $q/q^*(T_S)$. Thus the parameter β is more useful in a practical sense than the parameter α.

3. Experiments and results

The functional form of $F(\theta)$ was obtained from experiments performed with a layer of loam or sand packed in an evaporation pan, with care taken to make sure of the formulation of α and β [Eqs. (8) and (9)]. The value of $F(\theta)$ was determined from Eq. (7), using the measured values of E, C_E, u, T_S and q, and then values of α and β were determined from Eqs. (8) and (9).

Figure 2 displays a schematic illustration of the evaporation pan. The soil is 2 cm thick in the pan which has a diameter of 30 cm. Values of E and θ are obtained by weighing the pan. Here, θ is defined as the averaged soil moisture content of the 2 cm thick layer. The surface temperature T_S is measured with a thermocouple, and the air temperature T, wind speed u and the specific humidity of the air q are measured at a height of 50 cm above the soil surface. The bulk coefficient C_E is determined from the simultaneous measurement of evaporation from a fully wet soil, where q_s in Eq. (5) is equal to $q^*(T_S)$.

The experiments were performed during the period from midsummer (3 August) to the beginning of winter (23 December) of 1988. The ranges of experimental values and instruments are shown in Table 1. The averaging time of each value is 1 hour. The daily average temperature difference between the surface and the bottom of the soil layer was 1°–2°C. It should be noted that C_E-value depends on the Reynolds number based on the pan size, while C_E of the natural open ground surface depends only on the soil type and atmospheric stability.

Figures 3a and 3b show the experimental relationship between $F(\theta)$ and θ for loam and sand, respectively. It can be seen that $F(\theta)$ is well defined as a function of θ, supporting the present model. The solid line in each figure is expressed as

$$F(\theta) = F_1(\theta_{\mathrm{sat}} - \theta)^{F_2}, \quad (10)$$

for loam:

$$F_1 = 2.16 \times 10^2\,(\mathrm{m}), \quad F_2 = 10.0, \quad \theta_{\mathrm{sat}} = 0.490,$$

and for sand:

$$F_1 = 8.32 \times 10^5\,(\mathrm{m}), \quad F_2 = 16.6, \quad \theta_{\mathrm{sat}} = 0.392,$$

where θ_{sat} is the saturation soil moisture content for loam and sand, which was determined experimentally.

Defined by Eq. (4), $F(\theta)$ is related to the resistance to the vapor diffusion in the soil pore. When θ is large, water exists in large pores near the land surface and water vapor is easily transported to the land surface, resulting in quite small $F(\theta)$. For small θ, the distance for the water vapor to reach the land surface from the pore is relatively large and water vapor transport is prevented since water exists only in the interior of small complicated pores. Thus $F(\theta)$ increases with a decrease in θ.

For large θ where $F(\theta) \approx 0$, q_s is regarded as equal to $q^*(T_S)$. Here, the evaporation rate expressed by Eq. (5) corresponds to "potential evaporation."

TABLE 1. The ranges of experimental values during the experiments and instruments.

Symbol	Unit	Minimum	Maximum	Instrument
u	$\mathrm{m\ s^{-1}}$	0.3	2.7	3-cup anemometer (MAKINO, Type KC101)
T	°C	4.3	30.9	Ventilated thermistor thermometer
q		1.8×10^{-3}	2.5×10^{-2}	Ventilated thermistor psychrometer
T_s	°C	8.8	49.1	Copper-constantan thermocouple
C_E		8.5×10^{-3}	2.5×10^{-2}	(Estimated)
E	$\mathrm{mm\ day^{-1}}$	1.1	21.9	Pan and weighing machine
Solar radiation	$\mathrm{W\ m^{-2}}$	59	1029	NEO pyranometer (EKO, Type MS-43F)

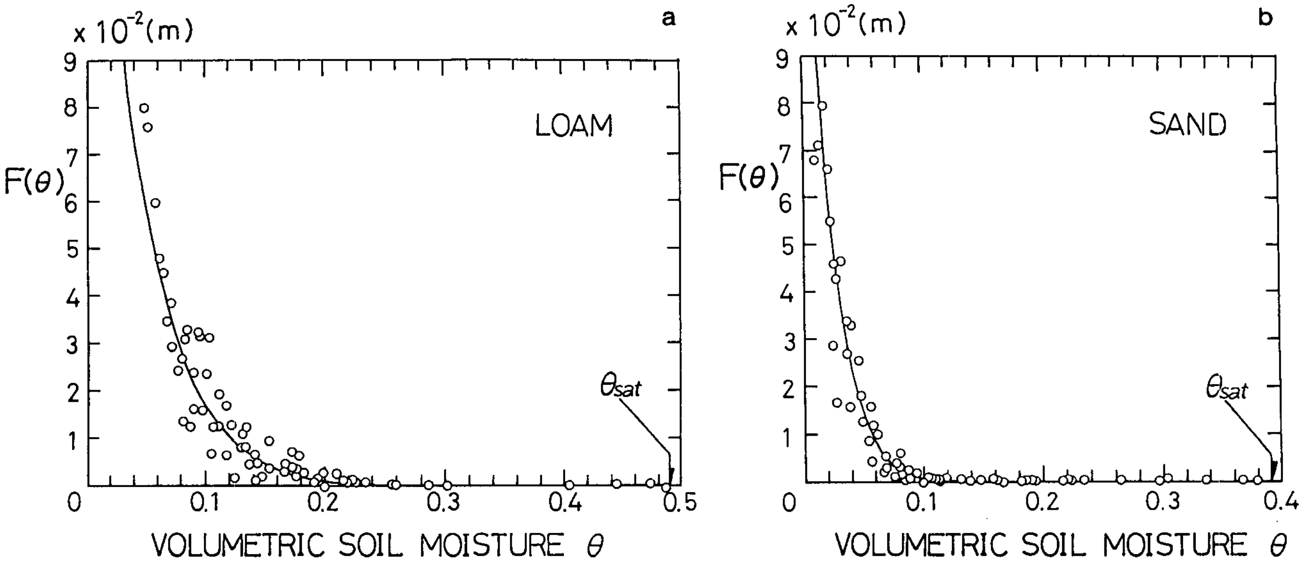

FIG. 3. Experimental relationship between $F(\theta)$ and the volumetric soil moisture content θ for (a) loam and (b) sand. Open circles are experimental values. The curve is the empirical relation expressed by Eq. (10).

One of the causes of the scatter in the observed values of $F(\theta)$ is expected to be the existence of θ-profile in the 2 cm soil layer (θ almost always increases with depth during a drying period). If θ near the surface is very small then the water vapor transport to the land surface is small, so $F(\theta)$ becomes larger than that in the case of even profile of θ. Very small θ near the surface is often led by such as strong solar radiation and very dry air.

The value of θ is the average soil moisture content within the 2 cm soil layer in this study. For application, if the soil layer is thicker than 2 cm then $F(\theta)$ is larger than the value calculated by Eq. (10), and if it is thinner than 2 cm, $F(\theta)$ is smaller, because the averaged θ increases with depth.

In the real soil, water vapor and liquid water are carried from the lower part of the soil to the surface 2 cm layer. A soil model considering the water transport from the lower part is now under development, and $F(\theta)$ obtained in the present study has been confirmed to be effective to use at the first 2 cm layer.

Figures 4a and 4b are plots of the values of α and β for loam as a function of θ. Each symbol denotes an experimental value, and the curved lines express Eq. (8) to Eq. (10). As expected from Eq. (8), α depends to a great extent on $q/q^*(T_S)$: for small values of θ, α does not become zero but instead approaches $q/q^*(T_S)$ with decreasing θ. This feature differs from previous works on α (for example, see Barton 1979). Figure 4b shows that β depends slightly on u. The experimental data have shown that the present model can satisfactorily describe the dependence of α and β on external conditions.

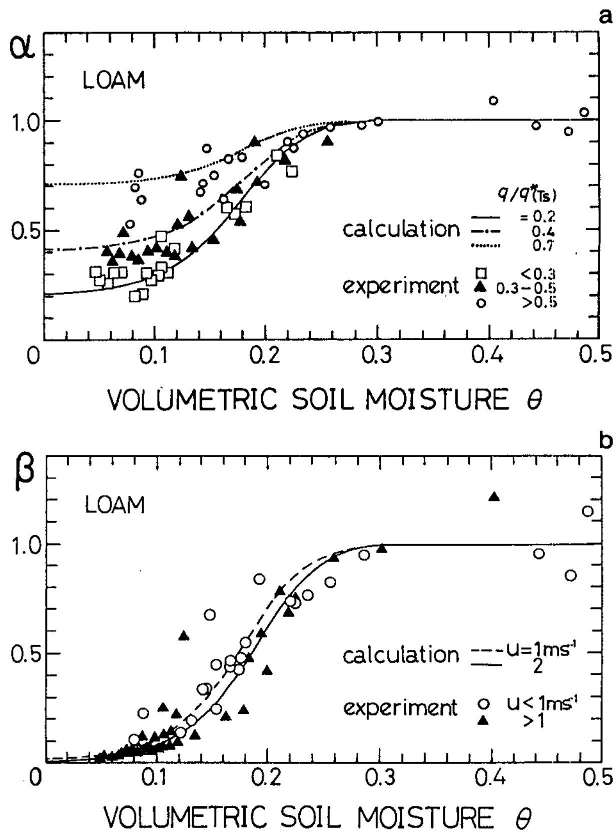

FIG. 4. (a) Surface moisture availability α for loam. Each symbol represents an experimental value. The curved line is calculated from Eqs. (8) and (10) with $u = 1\ \mathrm{m\ s^{-1}}$. (b) Surface moisture availability β for loam. The curved line is calculated according to Eqs. (9) and (10).

In the case of sand, the experimental results of α and β display a similar dependence on external conditions (figures are not shown).

4. Discussion

Surface relative humidity h defined by Eq. (3) has often been used instead of α in current numerical simulation models. To be exact, h is the humidity of the air that is in equilibrium with the soil water. For example, Fig. 5 shows h versus θ for loam. In the calculation, ψ is given as

$$\psi(\theta) = \psi_{sat}(\theta/\theta_{sat})^{-b}, \tag{11}$$

which is the usual parameterization of matric potential (Clapp and Hornberger 1978). Here ψ_{sat} is the matric potential at saturation, and b depends on the soil type (for loam $\psi_{sat} = -0.478$m and $b = 5.39$).

When $\theta = 0.2$, the water in the soil pore being localized as in Fig. 1, the air above the land surface is unsaturated ($\alpha < 1$; Fig. 4a), but the air adjacent to the water in the soil pore is saturated ($h = 1$; Fig. 5). The difference between α and h can be summarized as follows: α decreases slowly with a decrease in θ, with great dependence on $q/q^*(T_S)$ for small θ; h changes sharply from 1 to 0 around $\theta = 0.1$ (for loam), and is independent of q and T_S. These differences result from the fact that α is the humidity of air at the land surface while h is that of air adjacent to the water in the soil pore.

On the basis of similar reasons, Wetzel and Chang (1987) pointed out the inaccuracy of substituting h for the relative humidity at the plant canopy air layer or that at the humidity roughness height Z_q.

5. Concluding remarks

A simple model for the parameterization of evaporation from bare soil surfaces is developed by introducing a new parameter, $F(\theta)$, which represents the resistance to the diffusion of vapor in the soil pores. With this model an expression for the surface moisture availability is derived. The surface moisture availability depends not only on θ but also on T_S and other conditions of the air. The features compare favorably with the experimental results.

The difference between the humidity of the local air adjacent to the water in the soil pore and the macroscopic land surface humidity has been discussed. The former is often substituted for the land surface humidity in current numerical simulation models. However, these simulation models cannot represent the real evaporation under unsaturated soil conditions due to the essential difference in humidity between these two locations.

Future studies should concentrate on determining $F(\theta)$ for various soil textures in order to calculate the evaporation from actual soil surfaces more realistically.

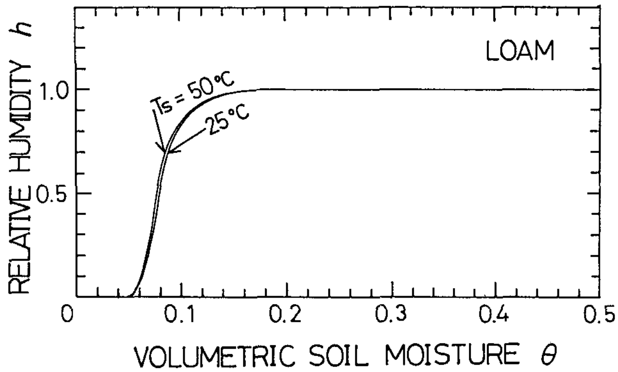

FIG. 5. Relative humidity defined by Eq. (3) as a function of θ.

Also, it is necessary to consider the transport of water from deep soil layers to the surface soil layer when predicting the time variation of surface soil moisture. These studies are needed to establish a concise and exact method of estimating the diurnal and seasonal variation of the evaporation rate from various bare soil surfaces.

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