Formation of the Dry Surface Layer and Its Effect on Bare Soil Evaporation : Field Observation and Numerical Experiments

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Abstract

Field measurements of soil water content, temperature, and relative humidity were made at several depths within a lysimeter where the dry surface layer (DSL) formed. The observed water content was almost constant in the DSL, while it increased toward deeper depth below the DSL. In addition, the peak of vapor density was observed at the bottom boundary of the DSL. Similar features to those observed was shown by numerical computations using a one-dimensional model of water and heat with high vertical-resolution. Furthermore, the result of numerical computations also demonstrated that the evaporating zone was located at the bottom boundary of the DSL, and that the thickness of the evaporating zone was extremely small, except for highly retentive soils. These results of both the field observation and numerical experiments indicate that information of the thickness of the DSL is required rather than that of water content of surface soil, to evaluate evaporation from bare soil surfaces.

1. Introduction

Soil moisture conditions of land area can have major effect on global climate through evapotranspiration. Many of the previous studies have empirically parameterized the availability of surface moisture for evaporation from bare soil, by using water content of surface soil layer. On the other hand, it is well known that so-called 'dry surface layer' (DSL) forms during soil drying.

According to several previous works, evaporation of soil water occurs mainly at the bottom boundary of the DSL instead of at the soil surface. In such case, the direct determining factor of the surface-moisture availability is considered to be the thickness of the DSL rather than the surface water content. In order to clarify the formation mechanism of the DSL and the effect of the DSL on evaporation from bare soil surfaces, field observation and numerical experiments were carried out.

2. Field observation

A field observation was conducted by using a lysimeter in which uniform fine sand was packed and groundwater level was maintained at a depth of 0.8 m, under the framework of the TSUKUBA 92 field campaign (Sugita *et al.*, 1993) held in Tsukuba, Japan, in the summer of 1992. Soil water content, soil temperature, and relative humidity in soil pores were measured at several depths below the soil surface by means of heat probe type soil moisture sensors, platinum resistance thermometers, and capacitance humidity sensors, respectively. The soil moisture conditions were fairly dry due to low precipitation during the field campaign, and the DSL which had a few to 10 cm thickness was observed. The observed soil water content showed clear difference between in and below the DSL (Fig. 1), that is, it was almost constant in the DSL, while it increased toward deeper depth below the DSL. The relative humidity of pore air was almost unity below the DSL. On the other hand, the relative humidity in the DSL was less than unity, and shows the tendency to decrease toward the soil surface. As a result of this humidity gradient and opposite temperature gradient, the peak of the vapor density was found out around the bottom boundary of the DSL. These observed results indicate that the divergence zone of vapor flux, namely 'evaporating zone', was located at the bottom boundary of the DSL.

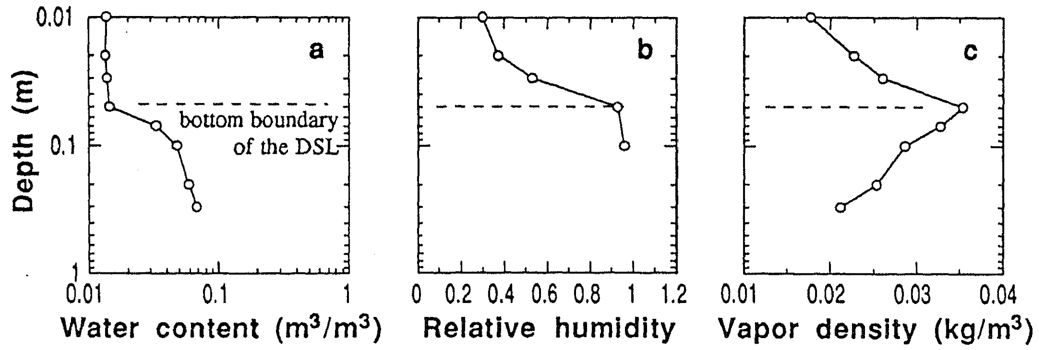

Fig.1 Vertical profiles of (a) water content, (b) relative humidity and (c) vapor density observed at 15:00 (JST) on Aug.10, 1992. The thickness of the DSL was determined by the difference in color of soil profile.

3. Numerical experiments

In order to clarify the microscopic structure of evaporating zone, numerical experiments was conducted using a one-dimensional model of water and heat flow with high vertical-resolution. The model used was fundamentally based on the modified Philip and de Vries theory (Philip and de Vries, 1957; Milly, 1982). In the model, the water flux consisted of liquid and vapor water flux driven by matric-head gradient and temperature gradient. The liquid water flux driven by the temperature gradient, however, was not considered in the present study, since it is less important in dry soils. The heat flux consisted of conduction and latent heat transfer due to vapor movement.

At the upper boundary, evaporation into the atmosphere (E_s) was given as the sink of water, and R_n-H-lE_s was given as the source of heat, where R_n is the net radiation, H the sensible heat flux into the atmosphere, and l the latent heat for vaporization. The E_s can be expressed as

$$E_s = \rho_a C_E u (h_s q_{sat}(T_s) - q_a) \qquad (1)$$

where ρ_a is the density of air, C_E the bulk transfer coefficient, u the wind speed, $q_{sat}(T_s)$ the saturated specific humidity at the soil surface, h_s the relative humidity at the soil surface, and q_a the specific humidity of atmosphere. The hs can be given by following equation, as

$$h_s = exp(\psi_s g / R T_s) \qquad (2)$$

where ψ_s and T_s are the matric head and temperature at the soil surface, g is the gravitational acceleration, and R the gas constant of water vapor. Diurnal variation of atmospheric conditions was not considered, and constant values of atmospheric conditions were given in the present study. The lower boundary conditions were constant temperature and no water flux at a depth of 0.6 m. This model consisted of 144 layers, and had very fine resolution (min., 0.5 mm) near the soil surface.

Table 1 Soil hydraulic parameters for the three example soils

Soil name	θ_{sat} (m^3/m^3)	θ_r (m^3/m^3)	α (m^{-1})	n (-)	K_{sat} (m/s)
Toyoura Sand	0.412	0.0239	3.14	7.0	1.25×10^{-5}
Silt Loam G.E.3 †	0.396	0.131	0.423	2.06	5.73×10^{-7}
Beit Netofa Clay†	0.446	0.0	0.152	1.17	9.49×10^{-9}

† data from van Genuchten (1980)

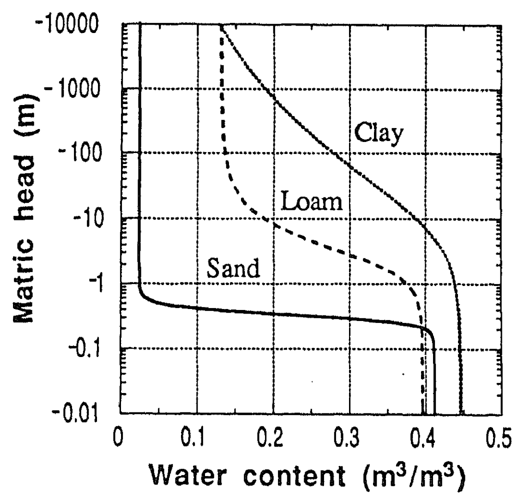

Fig.2 Soil moisture characteristic curves for the three soils used in the numerical experiments.

Hydraulic properties of soils were given by van Genuchten (1980)'s model. The parameters for three example soils used in the present study are listed in Table 1, and soil moisture characteristic curves for those soils are shown in Figure 2.

As an example, computational results for sand are shown in Figure 3. Very good similarities were recognized between observed and computed profiles of water content, relative humidity, and vapor density, respectively. These similarities demonstrate the validity of the model used. From comparison between the observed and computed results, the position of a inflection point in water content profile was considered to be corresponded to that of the bottom boundary of the DSL. It is remarkable that the evaporating zone was located at the inflection point in water content profile and that the evaporating zone was extremely thin.

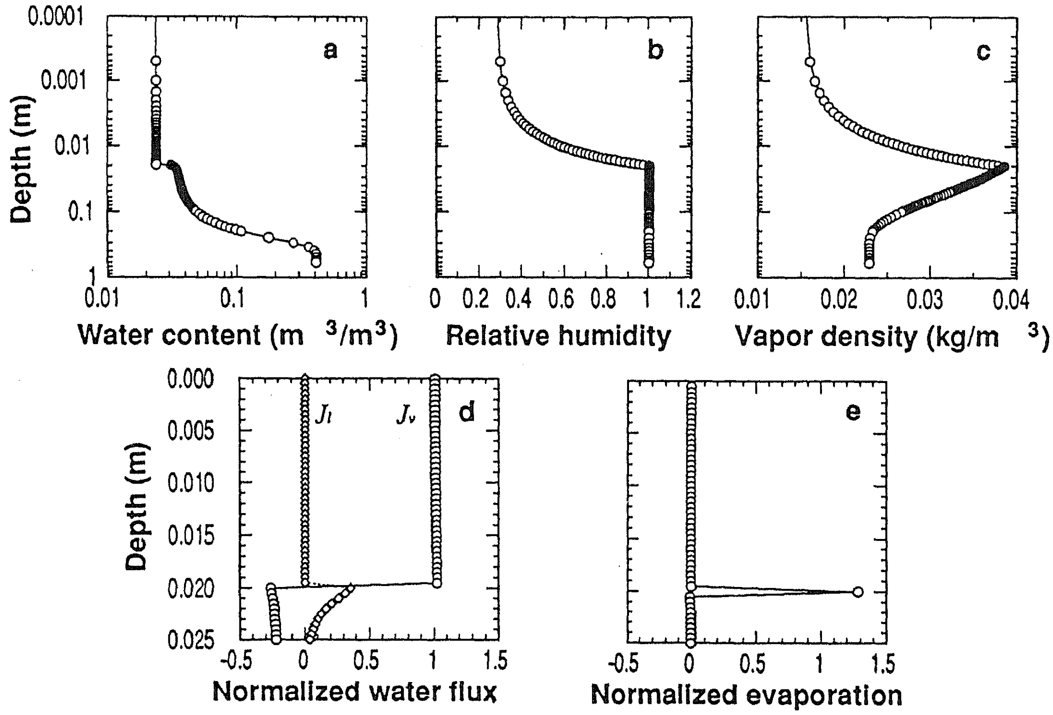

Fig.3 Computed vertical profiles for sand of (a) water content, (b) relative humidity, (c) vapor density, (d) liquid (J_l) and vapor (J_v) water fluxes, and (e) evaporation from each calculation node. The water fluxes and evaporation are normalized by evaporation into the atmosphere E.

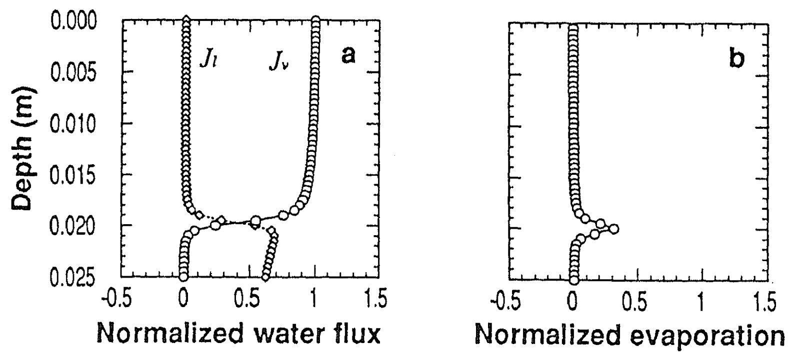

Fig.4 Computed vertical profiles for clay of (a) liquid (J_l) and vapor (J_v) water fluxes, and (b) evaporation from each calculation node. The water fluxes and evaporation are normalized by E.

The features like the above were also found out in the computed result for loam. On the other hand, for clay, two major points of difference can be found out (see Fig. 4). The first is that the decrease of liquid water flux toward soil surface around the bottom boundary of the DSL was relatively gentle, and thus, the thickness of the evaporating zone was relatively large. The second is that the vapor flux slightly increased toward soil surface at the upper region of the DSL, where no liquid water transport from deeper soil layer existed. The first point of difference is thought to be due to the difference of the properties of the unsaturated hydraulic conductivity and vapor conductivity between clay and other types of soil (see Fig. 5). For sand and loam, the unsaturated hydraulic conductivity rapidly decreased against the decrease of matric head, while it did not for clay. Therefore, the gentle shifting of water flux components for clay is thought to be owing to the gentle decrease of its hydraulic conductivity against matric head.

The second point of difference is related to the retention properties of soils. From the soil moisture characteristic curves (Fig. 2), the water contents of sand and loam demonstrate to have almost constant values at low matric-head level, where water transport in liquid phase is difficult to be made. On the other hand, it is shown that clay can retain abundant water at same level of matric head. Thus, it can be concluded that the soil water can evaporate from dry layer, where no liquid water supply from underlying soil, for highly retentive soils.

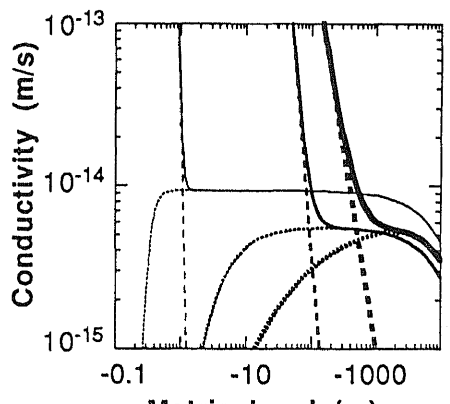

Fig.5 Relationships between conductivity and matric head for sand (thin line), loam (medium line) and clay (thick line). Broken and dotted line represent unsaturated hydraulic conductivity and vapor conductivity, respectively, and solid line represents the total of those.

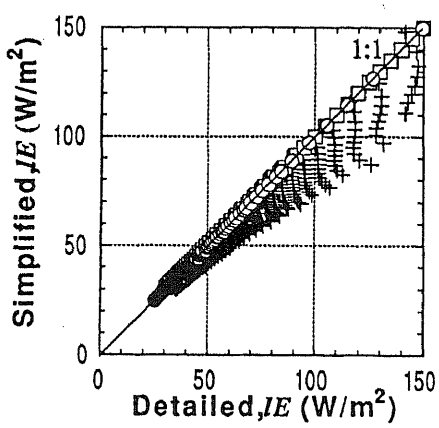

Fig.6 Comparison between evaporation estimated by the simplified method with eq. (3) and that obtained by the detailed model for sand (circle), loam (square) and clay (plus). Evaporation rate is presented as latent heat flux, where l is the latent heat for vaporization.

Assuming that the evaporating zone can be regarded as a horizontal plane, and that relative humidity at the evaporating zone is equal to unity, the evaporation rate can be represented by following equation, as

$$E_s' = \frac{\rho_a (q_{sat}(T_e) - q_a)}{1/C_E\, u + z_e/D_{ve}} \qquad (3)$$

where T_e is the temperature at the evaporating surface, z_e the depth of evaporating surface, which is corresponded to the thickness of the DSL, and D_{ve} the effective diffusivity of water vapor in the DSL. Figure 6 shows comparison of the evaporation estimated by the simplified method with equation (3) and that obtained by the detailed model mentioned the above. The values estimated by simplified method agree very well those obtained by the detailed model for sand an loam, although the simplified method tends to be lower than the detailed model due to the contribution of the evaporation occurred within the DSL. This indicates that equation (3) is effective to estimate the evaporation into the atmosphere, except for highly retentive soils.

Although it is not easy to directly measure the depth and temperature of the evaporating zone, the method to estimate the above two parameters from the conventional micro-meteorological measurement has been proposed by Yamanaka (1995).

4. Concluding remarks

It was clarified that the evaporating zone was located at the bottom boundary of the DSL, and was extremely thin for sand and loam. It was also shown that the parameterization, which regards the evaporating zone as a plane, is effective for those soils. These results indicate that information of the thickness of the DSL is required rather than that of water content of surface soil, to evaluate evaporation from bare soil surfaces. However, It should be noted that the above conclusions are obtained by assuming somewhat ideal conditions. The effects of the spatial variability in soil properties, the variation of atmospheric conditions, and the other factors, on the formation of the DSL and the structure of evaporating zone need to be investigated further.

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