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Study of heat and moisture transfer in soil with a dry surface layer

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Abstract

Mathematical model for describing simultaneous heat and moisture transfer in the porous soil with a dry surface layer was developed by using the volume-averaging method. Numerical simulation was conducted to investigate water evaporation, transient distributions of temperature and moisture in the porous soil at environmental conditions, which might be useful for agricultural application. In order to validate the mathematical model and numerical method, an experiment was conducted under natural environmental conditions. An additional experiment was conducted in a closed-loop wind tunnel to investigate the temperature effect on soil moisture transport. Theoretical and experimental results indicate that the dry surface layer has an important effect on heat and moisture migration in soil and the influence of temperature on moisture transport in unsaturated soil is significant.

1. Introduction

The cultural field remains largely bare from seeding to the early stage of plant growth and is directly exposed to environment. The soil water evaporation may deplete the moisture of soil surface, and the soil might be divided into two regions: the wet layer and the dry-saturated surface layer [1]. As an unsaturated porous medium, the wet soil consists of solid particles, liquid water, gaseous mixture of vapor and air and other chemical and biological substances. Heat and moisture transfers in the wet region are highly coupled. However, there is no free water in a dry layer and soil is saturated with gas.

Since the pioneer work of Darcy, the transport process in humid porous soil has attracted considerable interest over the past decades. A phenomenological model of combined heat and moisture transfer in porous media was developed by Philip and de Vries [2] and de Vries [3,4]. Whitaker [5] introduced a rigorous derivation of the model by volume averaging for pore scale equations. Some relevant studies can also be found in Refs. [6–11]. In recent years, there is increased interest in heat and moisture transfer in soil with a dry surface layer, such as, Ilic [1], Przesmycki and Strumillo [12], and Yamanaka and Yonetani [13]. However, few researches accurately located the transient interface between wet and dry layers. Moreover, to the best of our knowledge, no experimental and numerical studies on heat and moisture migration in soil with a dry-saturated

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Nomenclature

A	area, m^2
c	specific heat, J/(kg K)
D_l	diffusivity of water in porous medium, m^2/s
D_v	molecular diffusivity of vapor in air, m^2/s
D_{Tl}	diffusivity defined in Eq. (4), $m^2/(s\ K)$
D_{Tv}	diffusivity defined in Eq. (13), $m^2/(s\ K)$
D_{lv}	diffusivity defined in Eq. (13), m^2/s
g	acceleration of gravity, m/s^2
h_0	convection heat transfer coefficient, W/(m^2 K)
H_m	convection mass transfer coefficient, m/s
R_s	heat flux from solar radiation, W/m^2
λ_m	apparent thermal conductivity, W/(m K)
$(\rho c)_m$	apparent thermal capacity, J/(m^3 K)
K_l	hydraulic conductivity of liquid, m/s
K_g	infiltrating conductivity of gas-mixture, m/s
$\dot{m}$	mass rate of phase change, kg/(m^3 s)
p_a	atmosphere pressure, Pa
p_v	water vapor pressure on the evaporation front, Pa
τ	time, s
T	temperature, K (°C)
T_{sky}	temperature of the sky, K (°C)
T_I	upper side temperature adjacent to the interface between wet and dry soil layers, K (°C)
T_{II}	lower side temperature adjacent to the interface between wet and dry soil layers, K (°C)
$\vec{V}$	velocity vector, m/s
u_{gI}	upper side velocity adjacent to the interface between wet and dry soil layers, m/s
u_{gII}	lower side velocity adjacent to the interface between wet and dry soil layers, m/s
ω	emissivity of soil surface
ρ_∞	density of water vapor in the atmosphere, kg/m^3
ρ_0	density of water vapor on the evaporation front, kg/m^3
H	height of the analysis domain, m
H_1	height of unsaturated layer, m
RH	relative humidity of the atmosphere, %

Greek symbols

ε	volumetric phase content, %
υ	kinematic viscosity, m^2/s
σ	Stefan–Boltzman constant, W/(m^2 K^4)
γ	latent heat of water, J/kg
ρ	density, kg/m^3
ϕ	porosity, %

Subscripts and superscripts

a	air, ambient
g	gaseous mixture
l	liquid
m	mean
s	solid, solar, sky
v	vapor
∞	atmosphere

layer under natural conditions have been reported in the literature.

Though many hydrological phenomena may be taken to be isothermal [2], analysis of heat and moisture during evaporation from soil necessitates an understanding of the influence of temperature and temperature gradient on moisture transport in soil. Migration of heat and moisture in soil is a coupled energy and mass transport process, which is affected by temperature, phase content, pressure and velocity. Several researchers [14–16] indicated that temperature plays an important role on moisture transfer. However, they did not present a quantitative analysis.

In the present paper, a mathematical model is developed to describe coupled heat and moisture transport process in porous soil consisting of wet-unsaturated layer and dry-saturated layer. Numerical analysis is performed to predict water evaporation rates and transient variations of temperature and moisture in soil under natural conditions. The effects of temperature gradient on water transport and water vapor diffusion were also analyzed.

2. Model development

The following assumptions are made to model the physical mechanisms of mentioned phenomena in a two-layer soil bed.

(i) The medium is homogeneous and isotropic. The distension or contraction is negligible.
(ii) Local thermodynamics equilibrium is satisfied throughout the analytical domain.
(iii) Liquid phase and gas phase exist in funicular (continuous) states, respectively, in wet region.
(iv) Boussinesq's approximation is valid in gaseous natural convection in wet region.
(v) Darcy's law holds both for gas and liquid phases.

(vi) Perfect gas law is valid for gaseous mixture in pore space of porous matrix.

2.1. Thermally induced liquid water transport

Soil moisture transport is affected by the field distributions of temperature, phase content, pressure, velocity and so on. Several investigators demonstrated that moisture transfer under temperature gradients is negligibly small in a very dry or in very wet media, but attains a fairly well-defined maximum at an intermediate moisture content which depends on the soil-water tension and the air-filled pore space. Philip and de Vries [2] developed a coupled model, similar to Luikov's equations, by combining capillary diffusion of water with vapor diffusion under temperature gradient and moisture gradient in porous media. Applying moisture conservation, after considering the combined effect of temperature and moisture content, we can obtain

$$\frac{\partial \varepsilon_l}{\partial \tau} = \nabla \cdot ((D_{Tl} + D_{Tv})\nabla T) + \nabla \cdot ((D_l + D_v)\nabla \varepsilon_l) \tag{1}$$

The vapor diffusion will be discussed in details later, but the thermally induced liquid water transport equation can be written as

$$\frac{\partial \varepsilon_{l,T}}{\partial \tau} = \nabla \cdot (D_{Tl}\nabla T) \tag{2}$$

2.2. Transport of vapor phase

In a porous medium with evaporation of liquid water, diffusive motion of vapor influences the transport process. Several options are available to model the vapor diffusion: (1) introduce the governing equations according to continuum mechanics without involving diffusive mechanism; (2) add the diffusive effects of vapor into the continuity and energy equations without considering it in the momentum equation; (3) use the Stefan diffusion model through a stagnant gas film to calculate the diffusive velocity of vapor, and estimate the bulk velocity of gas phase by assuming that air is inert in the pore space. However, in the present investigation, we combine the diffusive motion of vapor with the motion of gaseous mixture, since any moving mass with an extended velocity can cause momentum change between two neighboring points in the averaging volume, no matter how it is driven by what kind of forces [10].

The vapor diffusion equations, considering the vapor flux resulting from the bulk flow of gaseous mixture (vapor and air), can be written as

$$\vec{q}_v = -D_v \upsilon \alpha \varepsilon_a \nabla \rho_v \tag{3}$$

where $\upsilon = [P/(P - p_v)]$ is the mass flow factor reflecting the bulk flow, α represents the tortuosity factor, ε_a designates the volumetric air content. As an approximation, we let $\varepsilon_a \cong \varepsilon_g = \phi - \varepsilon_l$. Note that ϕ and ε_l stand for the porosity and the liquid content, respectively.

Substituting all these parameters into the Eq. (3), we can obtain

$$\vec{q}_v = -D_v(\phi - \varepsilon_l)\alpha P/(P - p_v)\nabla \rho_v \tag{4}$$

From the Refs. [2,3], we have

$$\rho_v = \rho_{vs} h = \rho_{vs} e^{\psi g/RT} \tag{5}$$

Differentiating Eq. (5) yields

$$\nabla \rho_v = h\nabla \rho_{vs} + \rho_{vs}\nabla h \tag{6}$$

Noting that the relative humidity $h = f(\varepsilon_l, T)$, we can write

$$\nabla h = \frac{\partial h}{\partial \varepsilon_l}\nabla \varepsilon_l + \frac{\partial h}{\partial T}\nabla T \tag{7}$$

Substitution of Eq. (7) into (6) leads to

$$\nabla \rho_{vs} = h\frac{d\rho_{vs}}{dT}\nabla T + \rho_{vs}\left(\frac{\partial h}{\partial \varepsilon_l}\nabla \varepsilon_l + \frac{\partial h}{\partial T}\nabla T\right) \tag{8}$$

Using Eq. (5), one can obtain the partial differentials in the parentheses of right hand side of Eq. (8) as

$$\frac{\partial h}{\partial \varepsilon_l} = \frac{\partial h}{\partial \psi}\frac{\partial \psi}{\partial \varepsilon_l} = \frac{hg}{RT}\frac{\partial \psi}{\partial \varepsilon_l} \tag{9}$$

and

$$\frac{\partial h}{\partial T} = -\frac{hg\psi}{RT^2} \tag{10}$$

Substituting Eqs. (8)–(10) into Eq. (3) and denoting $\vec{V}_v = \vec{q}_v/\rho_v$, we can obtain an equivalent velocity for vapor diffusion motion.

$$\vec{V}_v = -D_{Tv}\nabla T - D_{lv}\nabla \varepsilon_l \tag{11}$$

2.3. Governing equations

2.3.1. Wet region

Previously developed seven-field model [10,11,18] by current authors is used to describe simultaneous heat, moisture and gas migration in unsaturated porous media with phase change in wet region.

1. *Continuity equations*

For liquid water:

$$\frac{\partial(\rho_l \varepsilon_l)}{\partial \tau} + \nabla \cdot (\rho_l \varepsilon_l \vec{V}_l) + \nabla \cdot [\rho_l (D_{Tl} \cdot \nabla T)] = -\dot{m} \tag{12}$$

For water vapor:

$$\frac{\partial(\rho_v \varepsilon_g)}{\partial \tau} + \nabla \cdot [\rho_v \varepsilon_g (\vec{V}_g + \vec{V}_v)] = \dot{m} \tag{13}$$

For gaseous mixture:

$$\frac{\partial(\rho_g \varepsilon_g)}{\partial \tau} + \nabla \cdot (\rho_g \varepsilon_g \vec{V}_g) = \dot{m} \tag{14}$$

2. *Momentum equations*

For liquid phase:

$$\frac{\partial \vec{V}_l}{\partial \tau} + (\vec{V}_l \cdot \nabla)\vec{V}_l - \frac{\dot{m}}{\rho_l \varepsilon_l}\vec{V}_l = -\frac{gD_l}{K_l}\nabla\varepsilon_l - \frac{g\varepsilon_l}{K_l}\vec{V}_l - \frac{g\varepsilon_g}{K_g}(\vec{V}_l - \vec{V}_g) + v_l\nabla^2\vec{V}_l - \vec{g} \tag{15}$$

For gaseous phase:

$$\frac{\partial \vec{V}_g}{\partial \tau} + (\vec{V}_g \cdot \nabla)\vec{V}_g + \frac{\dot{m}}{\rho_g \varepsilon_g}(\vec{V}_g + \vec{V}_v) = -\frac{1}{\rho_g}\nabla P - \frac{g\varepsilon_g}{K_g}(\vec{V}_g - \vec{V}_l) + v_g\nabla^2\vec{V}_g - \vec{g} \tag{16}$$

3. *Energy equation*

$$\frac{\partial}{\partial \tau}[(\rho c)_m T] + c_{p,l}T[\vec{V}_l \cdot \nabla(\rho_l\varepsilon_l)] + c_{p,g}T[\vec{V}_g \cdot \nabla(\rho_g\varepsilon_g)] + \varepsilon_l c_{p,l}[\vec{V}_l \cdot \nabla(\rho_l T)] + \varepsilon_g c_{p,g}[\vec{V}_g \cdot \nabla(\rho_g T)] = \nabla \cdot (\lambda_m \nabla T) - \gamma\dot{m} \tag{17}$$

2.3.2. Dry region

Continuity equation:

$$\frac{\partial(\rho_g\varepsilon_g)}{\partial \tau} + \nabla \cdot (\rho_g\varepsilon_g\vec{V}_g) = 0 \tag{18}$$

Momentum equation:

$$(\vec{V}_g \cdot \nabla)\vec{V}_g = -\frac{1}{\rho_g}\nabla P - \frac{g\varepsilon_g}{K_g}(\vec{V}_g) + v_g\nabla^2\vec{V}_g - \vec{g} \tag{19}$$

Energy equation:

$$\frac{\partial}{\partial \tau}[(\rho c)_m T] + c_{p,g}T[\vec{V}_g \cdot \nabla(\rho_g\varepsilon_g)] + \varepsilon_g c_{p,g}[\vec{V}_g \cdot \nabla(\rho_g T)] = \nabla \cdot (\lambda_m \nabla T) \tag{20}$$

2.4. About governing equations

Since the present mathematical modeling is quite complicated, some discussions, therefore, might be necessary for the above governing equations.

1. The effect of temperature on liquid water transport is considered in the continuity Eq. (12), and the thermally induced diffusivity of liquid D_{Tl} had been defined and measured by Jury and Miller [17].
2. The vapor-diffusion motion in unsaturated soil is affected by temperature gradient and water content gradient, and it plays an important role in transport process. Thus it should be considered in the corresponding governing Eqs. (13) and (16) as follows:

$$\vec{V}_v = -D_{Tv}\nabla T - D_{lv}\nabla\varepsilon_l \tag{21}$$

3. A formula to calculate the gaseous infiltrating conductivity K_g corresponding to the liquid hydraulic conductivity K_l, which can demonstrate the mechanisms of Darcy's drag resistance in terms of gaseous phase and the reaction of liquid to gaseous mixture, had been developed in the previous work [18]. At the same time, many researches [19,20] indicate that hydraulic conductivity K_l is also influenced by temperature. As a result, we defined the K_l as a function of water content ε_l and temperature T in the form of

$$K(\varepsilon_l, T) = K(\varepsilon_l, T_0)\frac{1 + a_1T + a_2T^2}{1 + a_1T_0 + a_2T_0^2} \tag{22}$$

4. Based on the fact $\varepsilon_s + \varepsilon_l + \varepsilon_g = 1$ the mean physical properties of porous media can be written as $(\rho c)_m = \varepsilon_s(\rho c)_s + \varepsilon_l(\rho c)_l + \varepsilon_g(\rho c)_g$. From the thermodynamics principle of constant volume for gaseous components $\varepsilon_g = \varepsilon_a = \varepsilon_v$, we simply set $(\rho c)_g = \rho_a c_a + \rho_v c_v$. The apparent heat conductivity of porous material, according to the mean-weighted method, is simply defined as $k_m = \varepsilon_s k_s + \varepsilon_l k_l + \varepsilon_g k_g$.

2.5. Boundary conditions

The physical model and its thermal and dynamic boundaries under investigation are illustrated in Fig. 1. In the numerical calculations, it is assumed that the lower boundary is adiabatic and impermeable while the top surface is exposed to the environment.

At $x = 0$:

$$\frac{\partial T}{\partial x} = 0, \quad u_l = u_g = 0, \quad \varepsilon_g = 0 \tag{23}$$

At $x = H$:

$$\dot{m}_s = H_m(\alpha_m\rho_{v0} - \rho_{v\infty}) \tag{24}$$

$$-\lambda_m\frac{\partial T}{\partial x} = h_0(T_0 - T) + \omega\sigma(T_{sky}^4 - T^4) + R_s - \gamma\dot{m}_s \tag{25}$$

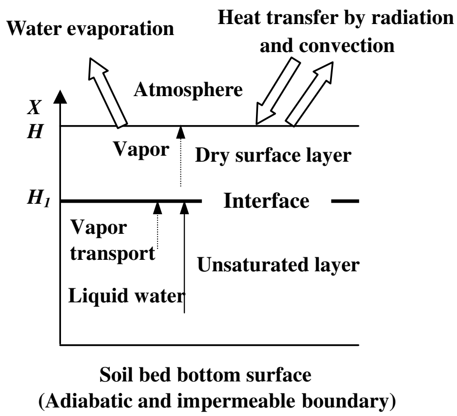

Fig. 1. Soil bed with wet and dry layers.

where R_s is heat flux from solar radiation, $\dot{m}_s$ is water evaporation rate from soil surface, α_m is soil surface specific relativity humidity. In general, the specific humidity of the gaseous mixture contacting with the surface of fairly wet soil can be taken as the saturation. During the soil drying, however, the specific humidity is reduced to below saturation, and then water evaporation is strongly dependent on the soil moisture condition [13]. In addition, $\dot{m}_s = 0$ occurs in the dry surface layer.

At $x = H_1$, the interface between wet and dry regions:

$$-\lambda_I \frac{\partial T}{\partial x}|_I + \lambda_{II} \frac{\partial T}{\partial x}|_{II} = -\gamma \dot{m}_s, \quad T_I = T_{II}, \quad u_{gI} = u_{gII} \tag{26}$$

With the development of dry surface layer (DSL), evaporation front moves from the top surface to the interface between wet and dry layers in the soil. According to the Stefan vapor diffusion law, the evaporation rate on the interface can be written as

$$\dot{m}_s = \frac{D}{R_A T} \frac{p_a}{(H - H_1)} \ln\left(\frac{p_a - p_{v\infty}}{p_a - \alpha_m p_v}\right) \tag{27}$$

where p_a is atmosphere pressure, $p_{v\infty}$ is vapor pressure in atmosphere, p_v is vapor pressure on evaporation front.

3. Numerical simulation and discussions

The Eqs. (12)–(20), together with the boundary condition Eqs. (23)–(27), are solved numerically by finite difference method. Since heat and mass transfer are highly coupled with the pressure, we choose the pressure-based algorithm, incorporated with staggered grid technique in primitive variables and *ADI* technique, to solve the coupled equations. Since there is no pressure-related term in liquid-phase momentum equation, the pressure correction equation is only needed for gas momentum equation, which is developed through gas-mixture continuity Eq. (13). In order to ensure the numerical convergence, the under relaxation and error feedback techniques are applied.

3.1. Numerical uncertainty analysis

The soil thermophysical properties, such as diffusivities, conductivities and apparent thermal conductivities, are required as parameters in simulation. The calculation shows that the soil water diffusivity, the hydraulic conductivity, and the thermal conductivity are the major physical parameters affecting the numerical results. The soil water diffusivity, D_l, and the hydraulic conductivity, K_l, have been discussed in our previous work [11] and can be determined by the following formula for the sandy soil:

$$D(\varepsilon_l) = 5.88 \times 10^{9.13\varepsilon_l - 9.0} \tag{28}$$

$$K_l(\varepsilon_l) = 9.48 \times 10^{12.53\varepsilon_l - 10.0} \tag{29}$$

To investigate the sensitivity of heat and mass transfer to soil property, we have performed the calculation with different D_l, λ_m and K_l values. The results indicate that the soil properties and thermophysical properties will influence the calculation value but will not change the predicting trend of heat and mass transfer in soil. The numerical results, sensitive to the physical properties which are directly taken from the published literature, show the good agreement between calculation and experiment, and thereby testify the reliability of the numerical analysis.

In order to simulate the transient variations of heat and mass transfer in soil under natural condition, the environmental parameters, such as solar radiation, ambient temperature and relative humidity, were measured for the time period from September 17 to September 20, 2002, and the results are presented in Figs. 2–4. The peak of solar radiation is at noon, and the peaks of ambient temperature and relativity humidity are at 13:00 and 7:00, respectively. Ambient temperature and relative humidity varied from 21 °C to 32 °C and from 35% to 85%, respectively, during the observation, and wind speed was about 1 m/s.

3.2. Results and discussion

Under the environmental parameters illustrated in Figs. 2–4, water evaporation and transient variations of temperature and moisture content in the soil bed (35 cm high and with a porosity of $\phi = 38.1\%$) were numerically studied.

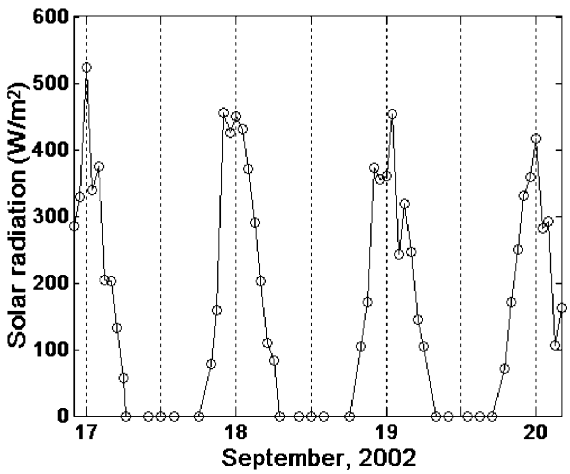

Fig. 2. Solar radiation.

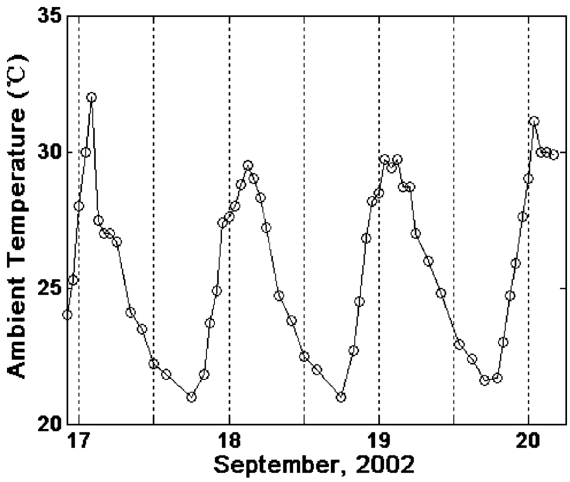

Fig. 3. Atmospheric temperature.

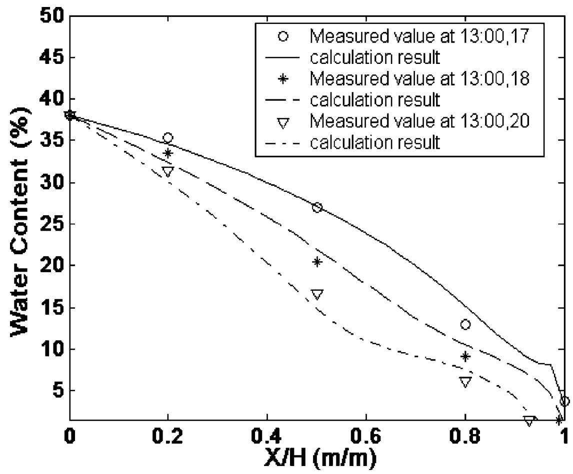

Fig. 5. Distribution of soil water content.

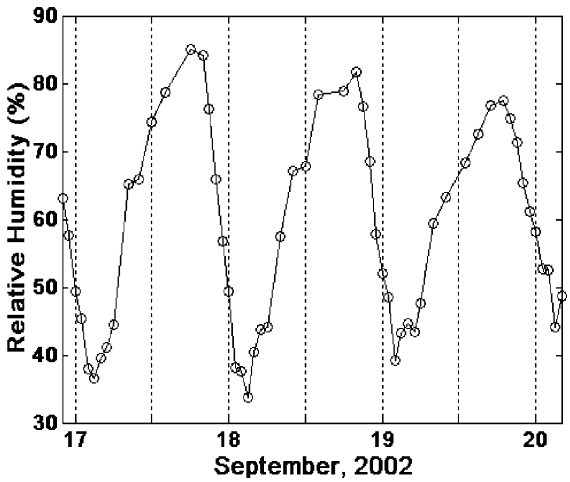

Fig. 4. Atmospheric relativity humidity.

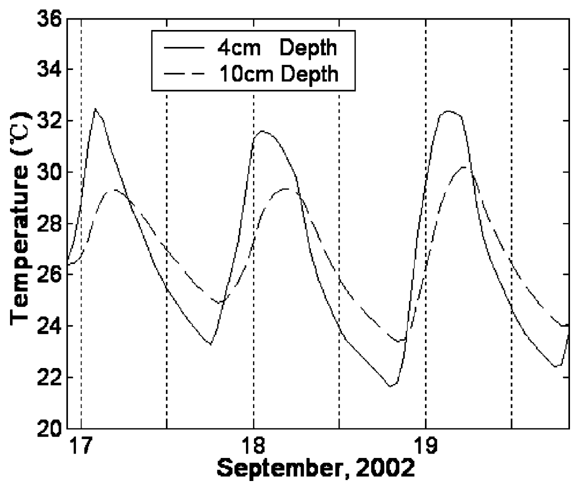

Fig. 6. Soil temperature variation.

3.2.1. Soil water content and temperature

As mentioned above, we have assumed an impermeable bottom boundary of the soil bed. Attention is paid to how the soil water content changes in the daily natural environmental conditions. Fig. 5 shows that, with evaporation, water content decreases gradually and soil surface became dry on September 18. With soil drying, diffusion resistance of water vapor increases. As a result, it can be seen from Fig. 5 that dry surface layer (DSL) develops slowly and its thickness increased only 10 mm from September 18 to September 20, 2002.

Under natural conditions, soil temperature exhibits periodical variation, and the daily temperature peak is at about 13:00 pm and the minimal surface soil temperature is almost at 6:00 am (see Fig. 6). With the increase in depth, the period of soil temperature variation increases and the amplitude of temperature fluctuation drops remarkably. From Eqs. (12) and (15), it may be observed that liquid water transport is not only controlled by moisture content gradient and gravity, but also influenced by soil temperature and temperature gradients. Hydraulic conductivity depends on both water content and soil temperature, see modified Eq. (22). With the increase in soil temperature, hydraulic conductivity increases while soil water capacity decreases. As a result, soil water content decreases with soil surface evaporation during daytime and slightly increases during nighttime (see Fig. 7), which validated the model presumed by Jury and Miller [17]. Figs. 6 and 7 also show that the higher the diurnal variation extent of soil temperature, the more remarkable the diurnal fluctuation of water content.

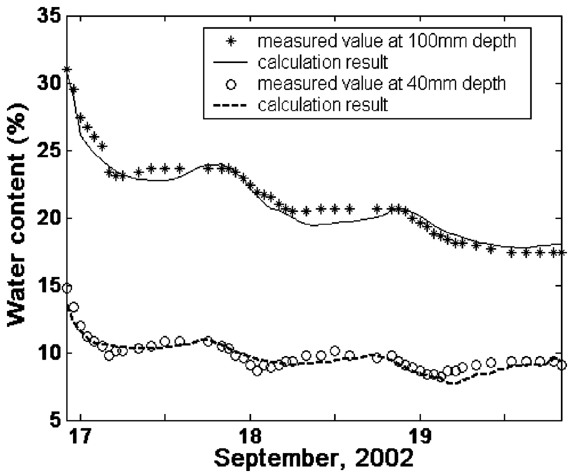

Fig. 7. Soil water content variation.

With the occurrence of dry surface layer, evaporation front moves down from soil upper surface to the interface between dry layer and wet layer, and evaporation intensity decreases gradually. As a result the absorbed heat from evaporation reduces and soil temperature increases obviously. For example, it can be seen from Figs. 2 and 3 that both solar radiation and ambient temperature in September 17 are lower than that in September 20. However, the temperature of soil with dry surface layer of 12 mm thickness at noon of September 20 was higher than that at noon of September 17, as shown in Fig. 8. Furthermore the temperature gradient of upper soil layer in September 20 was also larger than that in September 17, 2002.

3.2.2. Soil water evaporation

Evaporation from the surface of soil leads to considerable loss of water in agriculture, which also causes the problem of soil salinization, and influences heat and moisture transport in soil. Evaporating process is involved with soil physical property and atmosphere parameters [21], such as moisture, vapor pressure, temperature gradient, solar radiation and location of evaporation front. Fig. 9 shows that under natural condition, evaporation rate varies periodically and the peak appears at noon. Along with the development of dry surface region, evaporation resistance increases because of increased distance of vapor diffusion to atmosphere.

The vapor-diffusion motion in soil plays an important role to mass, momentum and energy transports. The phase change of water in unsaturated soil, which could be either evaporation or condensation, differs by situation. As known from Eq. (11), moving direction of water vapor is controlled by temperature gradients

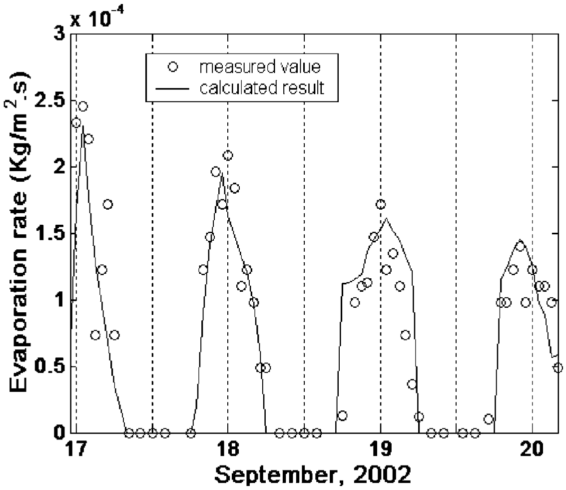

Fig. 9. Evaporation rate from soil surface.

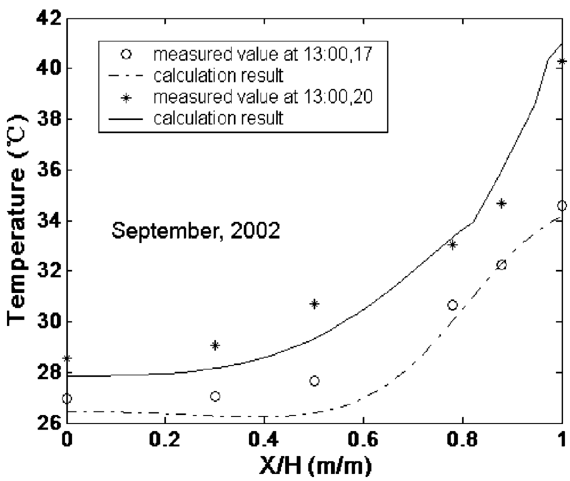

Fig. 8. Distribution of soil temperature.

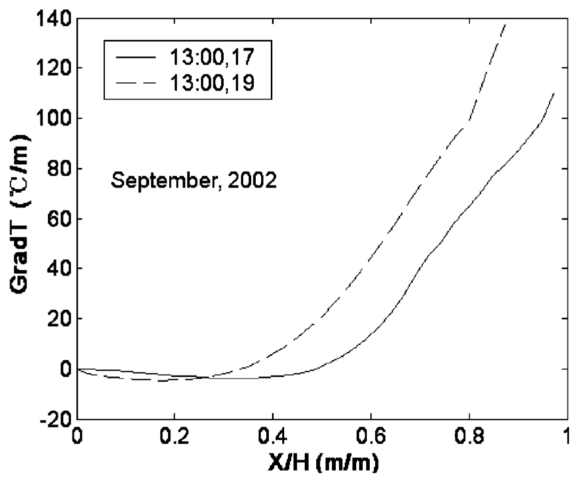

Fig. 10. Soil temperature gradient at noon.

and water content. Compared with the effect of water content gradient in the situation we have investigated, temperature gradient, the first term on the right-hand side in Eq. (11), is a dominant factor for the direction of water vapor flow. For the case of positive temperature gradient during daytime (see Fig. 10), water vapor will move downwards and condensation occurs (Fig. 11), but evaporation happens with a reverse temperature gradient at night (see Fig. 12). Thus a condensation process with a negative evaporation rate and an evaporation process can be observed in Figs. 11 and 13, respectively.

As discussed above, whether condensation or evaporation occurs, it usually depends on the direction of heat flux, not on the temperature level in the wet soil. Figs. 14 and 15 indicate that in the case of diurnal temperature variation, evaporation and condensation process alternatively occurs in unsaturated wet soil. Furthermore, with the increase in depth, the diurnal fluctuation of soil temperature reduces remarkably. Thus it can be observed from Fig. 15 that phase change intensity is sensitively related to temperature gradient: the larger the temperature gradient, the more intensive the evaporation or condensation rate.

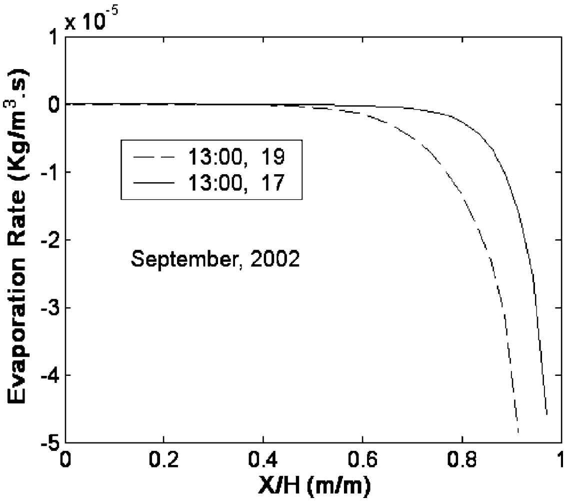

Fig. 11. Vapor condensation rate in soil.

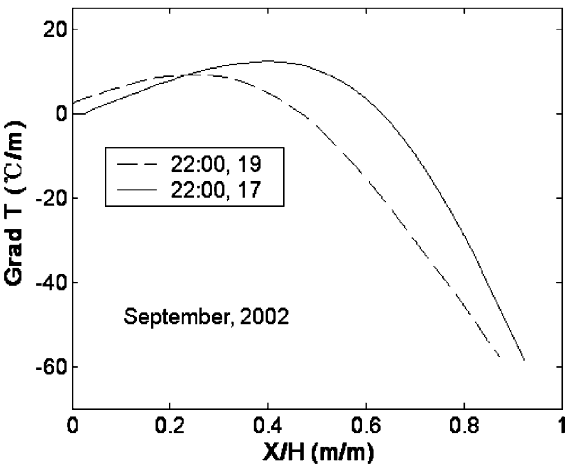

Fig. 12. Soil temperature gradient at night.

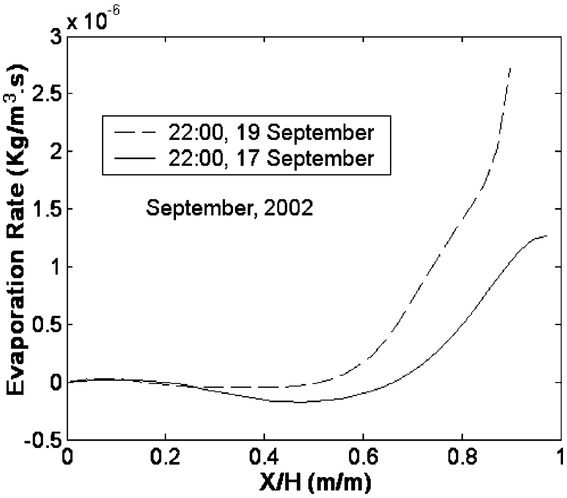

Fig. 13. Water evaporation rate in soil.

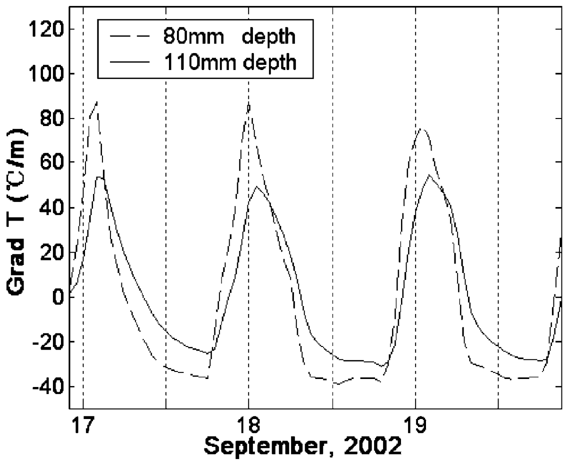

Fig. 14. Soil temperature gradient under natural condition.

4. Experiment

4.1. Site and equipment

In order to investigate the effect of temperature on moisture transport, two experiments were conducted.

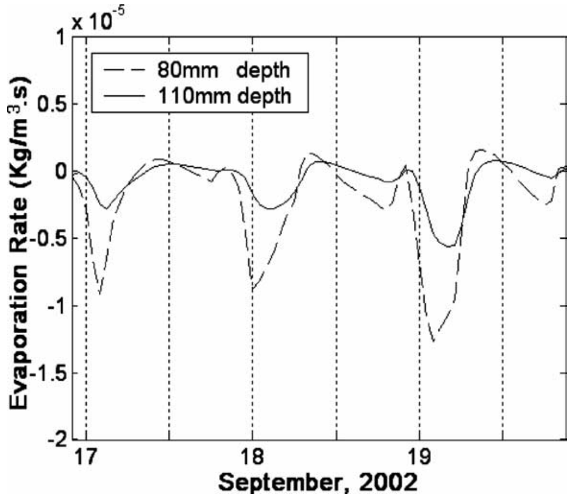

Fig. 15. Evaporation (condensation) rate in soil under natural condition.

One was conducted under natural environmental condition on the campus of Huazhong University of Science and Technology, located in Wuhan, China, during the period of September 17 to September 20, 2002. A cylindrical sand soil bed was used. In this bed, soil surface was exposed to atmosphere. Another was conducted in laboratory by using a closed-loop wind tunnel, in which air temperature, relative humidity and wind speed can be controlled at constant values.

4.2. Measurement and experimental uncertainty analysis

During the period of observation, air temperature, relative humidity, wind speed and solar radiation were measured at an intervals of 15 min by thermometer, temperature-humidity probes (HMP-35A), hot-wire anemometer (AVH 545T) and solar radiation meter, respectively. The measured values were averaged every hour for analysis. Surface soil water content was measured by drying soil in an oven. Temperature and volumetric water content in soil were measured by thermocouples (0.32 mm in diameter) and water content probes (MP-406), respectively. Additionally the thickness of dry soil layer was measured by excavating the shallow soil layer.

All the parameters, including temperature, relative humidity, soil temperature, soil water content and solar radiation, are directly measured in this experiment. The uncertainties of air temperature, relative humidity, soil temperature and soil water content are estimated as 2%, 3%, 1.5% and 2.5%, respectively, based on the instrument manuals. In order to reduce the uncertainty of solar radiation measurement, two solar radiation meters are used and measured values are averaged every 15 min.

4.3. Results and analysis

4.3.1. Soil temperature and water evaporation under natural condition

Under natural conditions, water evaporation rate, transient variations of temperature and moisture in soil and environmental parameters such as air temperature, relativity humidity and wind speed, were recorded and have been discussed previously.

Potential evaporation rate is the maximum flux at which water can be vaporized from a free water surface in atmosphere. Fig. 16 shows an example of the variation of actual and potential rates of evaporation with time. Two important points deserve our special concern: (1) the difference between actual soil evaporation rate and potential evaporation rates were small on September 18, when the soil evaporation is at the falling stage, and the difference increased obviously on September 19 with the occurrence of dry soil layer and (2) the actual evaporation rate in the early morning was equal to or larger than the potential rate throughout the observation period, even though soil drying was on-going.

The comparison between the calculated soil surface temperature and the measured soil surface temperature is shown in Fig. 17. The measured soil surface temperature is a typical one for those with a dry surface layer (DSL), and some similar change can be found in Ref. [22]. Fig. 17 shows that the calculation and the experiment agree reasonably well.

4.3.2. Variations of temperature and water content

In order to investigate the relationship between temperature variation and moisture content variation in unsaturated porous soil, observation was made in wind tunnel. Our objective is to study the influence of temperature on moisture transport in unsaturated porous soil.

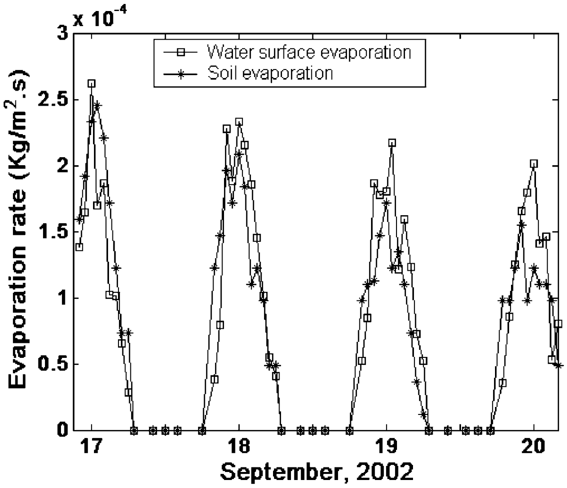

Fig. 16. Potential and actual soil evaporation rate.

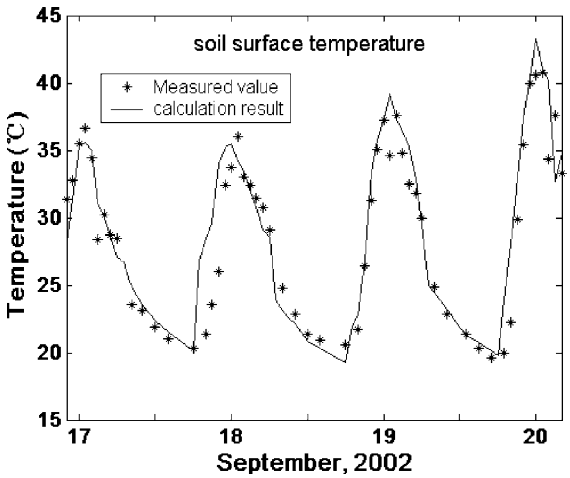

Fig. 17. Soil surface temperature.

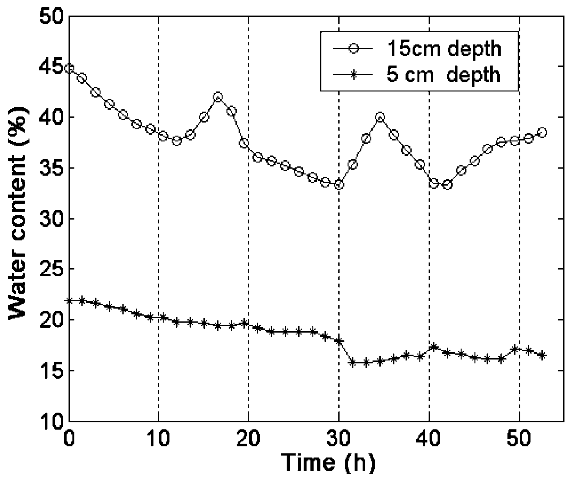

Fig. 19. Soil water content variation.

Figs. 18 and 19 indicate two important points: (1) Soil temperatures at different depths have similar variation while soil column was intermittently heated at bottom; (2) In shallow layer with low water content, the effect of temperature on moisture content was weak. In contrast, temperature effect was fairly obvious in deeper layer with relatively higher water content. The maximum fluctuations of soil temperature and water content at 15 cm depth were 30 °C and 6%, respectively. However, at the depth of 5 cm temperature fluctuation reached 15 °C while the maximum water content fluctuation was only 1%. This can be explained similarly that soil water capacity is controlled by both temperature and water content, see Eq. (22). In shallow layer with low water content, soil water capacity change would be small even though the soil temperature fluctuation is high, thus temperature effect on moisture transport is not obvious.

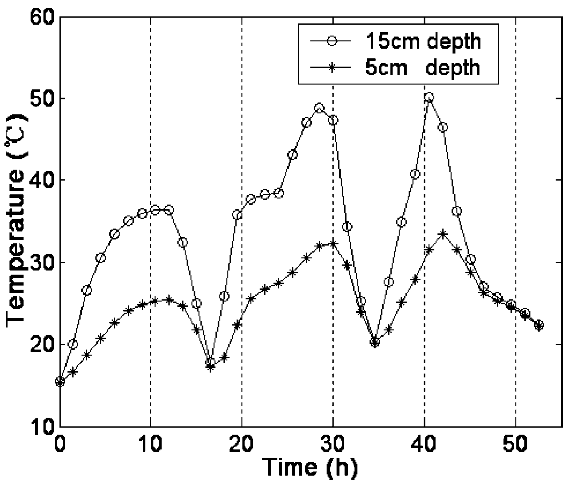

Fig. 18. Soil temperature variation.

5. Conclusions

The governing equations presented in this paper are suitable for describing simultaneous heat, moisture and gas migrations in porous soil with dry surface layer due to its theoretical completeness. The one-dimensional numerical solution of the model involves the estimations of soil water content, temperature and water evaporation under natural condition, which might be useful for agricultural application, especially in arid-semiarid areas, where water evaporation in soil is very strong and is important factor for soil salinization. Good agreement between the simulated results and the measured data validates the model and the numerical method.

Dry surface layer (DSL) has a significant effect on heat and mass transfer in soil. With the development of DSL, evaporation front moves from the upper soil surface to the interface between the wet and dry layers, while soil temperature increases remarkably and water evaporation rate reduces significantly due to the increase of vapor diffusion resistance to atmosphere. Furthermore the calculated and measured results indicate that temperature and temperature gradient play an important role in moisture transport in soil. The influence of temperature on liquid water transfer is strong and the effect of temperature gradient on vapor diffusion is obvious.

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