

Ground-Coupled Heat and Moisture Transfer from Buildings Part 1: Analysis and Modeling

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Ground-Coupled Heat and Moisture Transfer From Buildings

Part 1 – Analysis and Modeling

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ABSTRACT

Ground-heat transfer is tightly coupled with soil-moisture transfer. The coupling is threefold: heat is transferred by thermal conduction and by moisture transfer; the thermal properties of soil are strong functions of the moisture content; and moisture phase change includes latent heat effects and changes in thermal and hydraulic properties. A heat and moisture transfer model was developed to study the ground-coupled heat and moisture transfer from buildings. The model also includes detailed considerations of the atmospheric boundary conditions, including precipitation. Solutions for the soil temperature distribution are obtained using a finite element procedure. The model compared well with the seasonal variation of measured ground temperatures.

NOMENCLATURE

C_p = specific heat capacity [J/kg K]
 $C_{T\ell}, C_{\psi\ell}$ = thermal and matric liquid capacitance [K^{-1}, m^{-1}]
 $C_{Tm}, C_{\psi m}$ = thermal and matric moisture capacitance [K^{-1}, m^{-1}]
 $C_{Tv}, C_{\psi v}$ = thermal and matric vapor capacitance [K^{-1}, m^{-1}]
 $C_{TT}, C_{T\psi}$ = thermal and matric heat capacitance [$J s/m^4, J/m^3 K$]
 D_a = molecular diffusivity of water vapor in air [m^2/s]
 D_h, D_v = heat and vapor diffusivities in air
 D_m = neutral stability momentum transfer coef. [m/s]
 $D_{Tm}, D_{\psi m}$ = thermal and matric moisture diffusivities [$m^2/s K, m/s$]
 $D_{Tv}, D_{\psi v}$ = thermal and matric vapor diffusivities [m/s, $m^2/s K$]
 $D_{\psi T}$ = matric potential heat diffusivity [W/m^2]
 e = specific internally stored energy
 E = evaporation rate [s^{-1}]
 f = vapor diffusion correction factor
 g = acceleration due to gravity [m/s^2]
 h = convective heat transfer coefficient [$W/m^2 K$]
 h_{fg} = latent heat of water vaporization [J/kg]
 k = thermal conductivity [W/m K]
 k^* = thermal conductivity of soil with no moisture movement [W/m K]
 K = soil hydraulic conductivity [m/s]

\dot{m} = mass flow rate [$kg/m^2 S$]
 P = pressure [Pa]
 q = heat flux [W/m^2]
 R_v = gas constant for water vapor [461.5 J/mole K]
 T = temperature [K]
 $(\nabla T)_p/\nabla T$ = ratio of microscopic temperature gradient in the pores to the macroscopic temperature gradient over the soil sample
 t = time [s]
 u_ℓ = bulk liquid velocity [m/s]
 u_* = frictional velocity [m/s]
 W = heat of wetting
 x = volumetric fraction
 z = vertical displacement, positive upward [m]

Greek Letters

β = time-weighting factor used in the transient solution
 ϕ = relative humidity
 Φ = total soil potential for liquid flow [m]
 η = soil porosity
 ρ = density [kg/m^3]
 θ_ℓ = volumetric liquid content [m^3/m^3]
 θ_k = critical moisture content below which liquid water is no longer continuous
 ξ = temperature gradient ratio term used to calculate soil thermal conductivity
 ψ = matric liquid potential [m]

Subscripts

a = air
amb = ambient conditions
conv = convection
 ℓ = liquid water
m = moisture
lw = long wave
o = reference
p = gas-filled pores
s = solid surface
sens = sensible heat
sw = short wave
v = water vapor
vs = saturated water vapor

1. Introduction

The heat transfer between a building and the surrounding soil is complicated by many unknowns, such as the soil's physical properties and complex physical processes, many of which involve moisture considerations. For example, heat is transferred by thermal conduction and moisture transfer; the thermal properties of soil are strong functions of the moisture content; and moisture phase change includes latent heat effects and changes in the soil's thermal and hydraulic properties. This paper outlines the development of a model to compute building ground-coupled heat and moisture transfer [1]. In a companion paper, the model is applied toward determining the heat transfer from building slabs and basements.

There have been numerous works on building ground-coupled heat transfer. Some of the important numerical solutions to the problem are by Kusuda and Achenbach [2], Wang [3], Mitalas [4], and Bahnfleth [5]. The only work to consider heat and moisture transfer was by Shen [6].

2. Heat Transfer Paths in Soil

Heat transfer in soil occurs by conduction through the soil grains, liquid, and gases; latent heat transfer through evaporation-condensation cycles; sensible heat transfer by vapor and liquid diffusion and convection; and radiation in the gas-filled pores [7,8]. Conduction through the solid soil particles is the dominant mode of heat transfer under most circumstances. The presence of moisture in the soil provides additional transport mechanisms. For example, in the gas-filled pores of unsaturated soils, liquid water evaporates on the warm side, absorbing the latent heat of vaporization. Diffusion occurs because of the vapor-pressure gradient, and the vapor condenses on the other side of the pore, releasing the latent heat of vaporization. This process increases the overall thermal conductivity because the effective thermal conductivity of the vapor-distillation cycles is larger than that of the gas-filled pores alone. Forced convection from infiltration of liquid at the ground surface can be significant for a short time after a large amount of precipitation.

3. Hydraulic and Thermal Properties of Soil

There are three major soil properties that govern the transport of heat and mass in soils. These are the soil water potential, the hydraulic conductivity, and the thermal conductivity. In the model presented in this paper, we assume the total soil water potential to be the sum of the gravitational and matric potentials as presented in Eq. (1), where z is taken as positive upwards. The soil matric potential is a measure of the attractive force of the capillary

and adsorptive actions of the soil matrix. The potential is often expressed as an equivalent head of water and therefore has the dimension of length.

$$\Phi = \psi + z \quad (1)$$

The relationship between the matric potential and the soil volumetric moisture content is given by the soil moisture retention curve shown in Fig. 1 for a loamy sand [9], and Yolo light clay [10]. Note that the sand and clay are saturated at moisture contents of 0.395 and 0.495. The flatness of the sandy soil curve between moisture contents of 0.1 and 0.4 indicates that the moisture drains more rapidly from the sandy soil over this range relative to the clay soil, which has a higher attraction to moisture. The temperature dependence of the matric potential is approximated through the temperature dependence of the soil water surface tension. Hysteresis between wetting and drying is not included in this model.

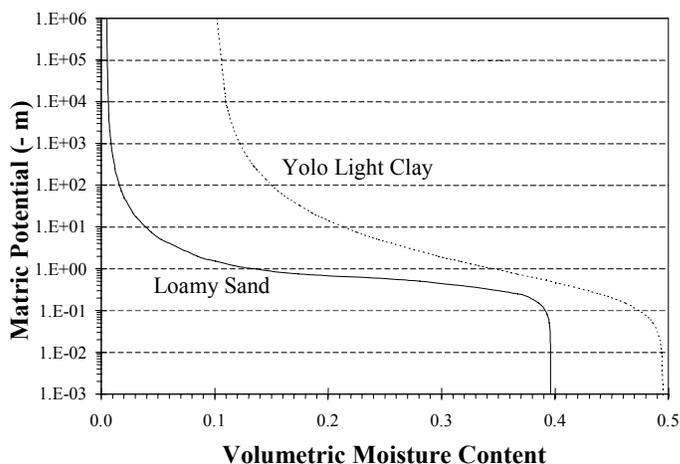


Figure 1. Soil moisture retention curves for a loamy sand and Yolo light clay

The hydraulic conductivity in unsaturated soil is a function of soil and fluid properties, moisture content, and temperature. Curves of the hydraulic conductivity of a loamy sand and of Yolo light clay are shown in Fig. 2. Sands will have a hydraulic conductivity about three orders of magnitude greater than clays due to their larger grain size. The temperature dependence of the hydraulic conductivity is approximated through the temperature dependence of the soil water kinematic viscosity. The hydraulic conductivity of frozen soil is approximated by the relationship for unfrozen soil using the moisture content that remains unfrozen.

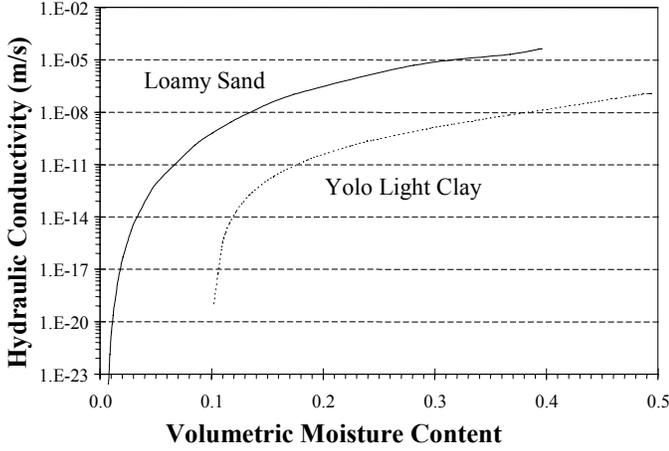


Figure 2. Hydraulic conductivity of loamy sand and Yolo light clay

Soil thermal conductivity is a function of the moisture content; temperature; and the size, shape, orientation, packing, and type of grains that make up the soil matrix. The thermal conductivity is very dependent on the moisture content, as changes in the moisture content can affect the thermal conductivity by almost a factor of ten. In this model, the soil thermal conductivity is approximated by a method developed by de Vries [11]. It is assumed that the soil consists of a continuous medium with an even distribution of ellipsoidal-shaped grains. The thermal conductivity is then given by Eq. (2). This equation represents an average of the thermal conductivities of water w , pores p , and n types of soil grains (including ice), weighted by the volumetric contents x and the ratio of the average temperature gradient in the constituent and the average temperature gradient of the medium ξ . Water is considered to be the continuous medium, except at very low moisture contents, where air is used.

$$k = \frac{x_w k_w + \xi_p x_p k_p + \sum_{i=1}^n \xi_i x_i k_i}{x_w + \xi_p x_p + \sum_{i=1}^n \xi_i x_i} \quad (2)$$

The main driving force for the vapor diffusion in the soil pores is the vapor density gradient resulting from temperature gradient. Because temperature is the main driving force, this heat transfer is treated as an effective heat conduction, and the thermal conductivity of the gas-filled pores is estimated as the sum of the thermal conductivity of the air k_a and the effective thermal conductivity of the vapor diffusion k_v .

$$k_p = k_a + k_v \quad (3)$$

The effective thermal conductivity of vapor distillation cycles for a single pore can be expressed as

$$k_v = \phi h_{fg} D_a \frac{d\rho_{vs}}{dT} \quad (4)$$

The effective thermal conductivity of Bighorn sandy loam as a function of moisture content with and without the vapor diffusion term is shown in Fig. 3 for two different temperatures. As the temperature increases, the contribution of the vapor diffusion term increases. The thermal conductivity of frozen soil relative to the unfrozen value decreases for low moisture contents and increases for high moisture contents [12].

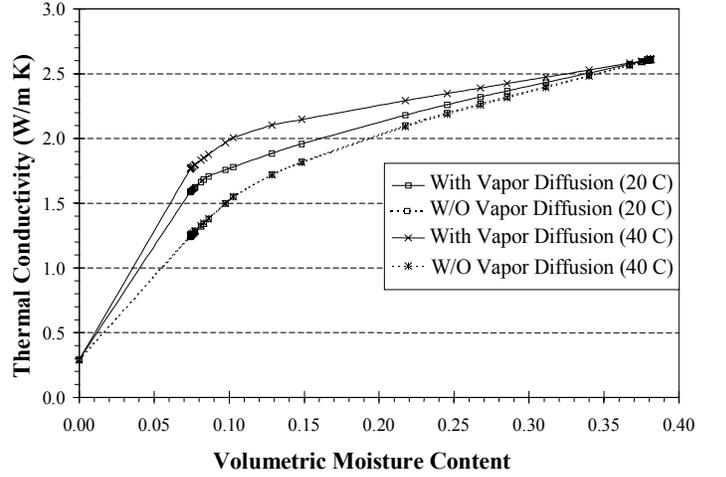


Figure 3. Effective thermal conductivity of sandy loam

4. Coupled Heat and Moisture Transfer Model

The heat and mass transfer equations developed in this work follow an approach based on physical models of the processes that occur in the soil [7,13]. The moisture transfer equations are cast in terms of the matric potential (ψ -based equations) because this approach can handle saturation conditions. The main simplifying assumptions made in the derivations are that the soil is assumed to be homogenous and isotropic within each defined unit of soil.

4.1. Liquid Transfer

Darcy's law can be extended to unsaturated soil by using the gradient of the total potential, Φ , and defining the hydraulic conductivity, K (m/s), as a function of the soil water matric potential.

$$\mathbf{u}_\ell = -K(\psi)\nabla\Phi \quad (5)$$

Substitution of Eq. (1) into Eq. (5) yields

$$\mathbf{u}_\ell = -K\nabla\psi - K\hat{k} \quad (6)$$

By applying the continuity equation to the liquid moisture content of a control volume of soil, the conservation of liquid can be written as

$$\frac{\partial \theta_\ell}{\partial t} = -\nabla \cdot \mathbf{u}_\ell - E \quad (7)$$

The liquid content is a function of the matric potential and temperature; therefore, the time derivative of the liquid content can be expanded by the chain rule to give

$$C_{\psi\ell} \frac{\partial \psi}{\partial t} + C_{T\ell} \frac{\partial T}{\partial t} = \nabla \cdot [K \nabla \psi] + \frac{\partial K}{\partial z} - E \quad (8)$$

4.2. Vapor Transfer

The vapor diffusion in a gas-filled pore can be approximated by modifying Fick's law of diffusion, assuming a uniform and constant total pressure, P , and that the vapor behaves as an ideal gas [8,14].

$$\dot{\mathbf{m}}_v = -D_a \left(\nabla \rho_v + \frac{\rho_v}{T} (\nabla T)_p \right) \quad (9)$$

The molecular diffusivity of water vapor in air, D_a , is given by Eq. (10), and $(\nabla T)_p$ is the temperature gradient across a single gas-filled pore [8].

$$D_a = c \left(\frac{P_o}{P} \right) \left(\frac{T}{T_o} \right)^n \quad (10)$$

The reference pressure is $P_o = 1.01325 \times 10^5$ Pa, the reference temperature is $T_o = 273.15$ K, $c = 2.17 \times 10^{-5}$ m²/s, and $n = 1.88$. The vapor density in the pores can be expressed as the product of the relative humidity, ϕ , and the saturated vapor density, ρ_{vs} .

$$\rho_v = \rho_{vs} \phi \quad (11)$$

Assuming that the soil liquid and vapor are in thermodynamic equilibrium, and in the absence of solutes, the relative humidity in the soil pores may be written as [15]

$$\phi = \exp \left(\frac{\psi g}{R_v T} \right) \quad (12)$$

From Eq.'s (11) and (12) the vapor diffusion equation can be written as

$$\dot{\mathbf{m}}_v = -D_a \phi \rho_{vs} \left[\frac{g}{R_v T} \nabla \psi + \left(\frac{1}{\rho_{vs}} \frac{\partial \rho_{vs}}{\partial T} + \frac{1}{T} \right. \right. \\ \left. \left. - \frac{\psi g}{R_v T^2} + \frac{g}{R_v T} \frac{\partial \psi}{\partial T} \Big|_{\theta} \right) (\nabla T)_p \right] \quad (13)$$

Only the first of the four temperature gradient coefficient terms is significant for most situations. The last two terms are important for $\psi < -10^4$ m [1]. When applying the vapor mass flux given by Eq. (13) to a porous soil system, the effects of the reduced cross section for diffusion, the tortuosity, and the interactions between the vapor and liquid phases must be addressed. A vapor diffusion correction factor, $f(\theta_\ell)$, is added to account for these effects [7].

$$f(\theta_\ell) = \eta, \quad \text{for } \theta_\ell \leq \theta_k \\ f(\theta_\ell) = \theta_a + \theta_a \theta_\ell / (\eta - \theta_k), \quad \text{for } \theta_\ell > \theta_k \quad (14)$$

The critical moisture content, θ_k , is that below which the hydraulic conductivity falls to a value much lower than its value at saturation. Deru [1] presents a method of determining this value based on the relative humidity of the soil moisture. The vapor mass flux of Eq. (13), divided by the liquid density for consistency with the liquid-transfer equation can be written for the soil system as

$$\frac{\dot{\mathbf{m}}_v}{\rho_\ell} = -D_{\psi v} \nabla \psi - D_{Tv} \nabla T \quad (15)$$

where the matric and thermal vapor diffusivities are

$$D_{\psi v} = f(\theta_\ell) D_a \frac{\rho_{vs}}{\rho_\ell} \frac{\phi g}{R_v T} \quad (16)$$

$$D_{Tv} = f(\theta_\ell) D_a \phi \frac{\rho_{vs}}{\rho_\ell} \left(\frac{1}{\rho_{vs}} \frac{\partial \rho_{vs}}{\partial T} \right. \\ \left. - \frac{\psi g}{R_v T^2} + \frac{g}{R_v T} \frac{\partial \psi}{\partial T} \Big|_{\theta} \right) \frac{(\nabla T)_p}{(\nabla T)} \quad (17)$$

The temperature gradients across the gas-filled pores are higher than those across the system because of the lower thermal conductivity of the gas-filled pores, which is accounted for by the ratio of the average temperature gradient across the pores and the temperature gradient across the system.

The vapor content is expressed as an equivalent liquid content. Assuming thermodynamic equilibrium between the liquid and vapor, we can write

$$\theta_v = \frac{(\eta - \theta_\ell) \rho_v}{\rho_\ell} \quad (18)$$

Expanding $(\partial \theta_v / \partial t)$, and using the chain rule to include the dependence on moisture and temperature yields

$$\frac{\partial \theta_v}{\partial t} = C_{\psi v} \frac{\partial \psi}{\partial t} + C_{Tv} \frac{\partial T}{\partial t} \quad (19)$$

where the matric and thermal vapor capacitances are

$$C_{\psi v} = \frac{\rho_{vs}\phi}{\rho_\ell} \left[\frac{(\eta - \theta_\ell)g}{R_v T} - \left(\frac{\partial \theta_\ell}{\partial \psi} \right)_T \right] \quad (20)$$

$$C_{Tv} = \frac{\rho_{vs}\phi}{\rho_\ell} \left[\frac{(\eta - \theta_\ell)}{\rho_{vs}} \left(\frac{\partial \rho_{vs}}{\partial T} \right) - \left(\frac{\partial \theta_\ell}{\partial T} \right)_\psi \right] \quad (21)$$

The conservation of vapor content can be written as

$$\frac{\partial \theta_v}{\partial t} = -\nabla \cdot (\dot{\mathbf{m}}_v / \rho_\ell) + E \quad (22)$$

The governing equation for vapor transfer can now be derived by substitution of Eq.'s (15) and (19) into (22).

$$C_{\psi v} \frac{\partial \psi}{\partial t} + C_{Tv} \frac{\partial T}{\partial t} = \nabla \cdot [D_{\psi v} \nabla \psi] + \nabla \cdot [D_{Tv} \nabla T] + E \quad (23)$$

4.3. Total Moisture Transfer

The moisture transfer equation is then found by combining the equations for liquid and vapor transport, Eq.'s (8) and (23).

$$C_{\psi m} \frac{\partial \psi}{\partial t} + C_{Tm} \frac{\partial T}{\partial t} = \nabla \cdot [D_{\psi m} \nabla \psi] + \nabla \cdot [D_{Tm} \nabla T] + \frac{\partial K}{\partial z} \quad (24)$$

4.4. Heat Transfer

By applying the conservation of energy on the soil system we can write

$$\frac{\partial(\rho e)}{\partial t} = -\nabla \cdot \mathbf{q} + q''' \quad (25)$$

Neglecting the heat capacity of the air, the thermal energy of the soil system relative to a reference temperature, T_o , can be written as [16]

$$\begin{aligned} \rho e = & \rho_s (1 - \eta) C_{p,s} (T - T_o) + \rho_\ell \theta_\ell C_{p,\ell} (T - T_o) \\ & + \rho_\ell \theta_v C_{p,v} (T - T_o) + \rho_\ell \theta_v h_{fg} (T_o) \\ & - \rho_\ell \int_0^{\theta_\ell} W(\theta) d\theta \end{aligned} \quad (26)$$

The heat of wetting, W , is the energy released when the water molecules adsorb on the surface of the soil grains. It is a function of the initial moisture content and surface area. It is only significant for wetting dry clay soils and is neglected in this work.

The heat flux in the soil can be represented by

$$\mathbf{q} = -k^* \nabla T + h_{fg} (T_o) \dot{\mathbf{m}}_v + C_{p,\ell} (T - T_o) \dot{\mathbf{m}}_\ell + C_{p,v} (T - T_o) \dot{\mathbf{m}}_v \quad (27)$$

The thermal conductivity k^* represents the pure heat conduction through the soil system with no moisture movement. Note that this is different from the thermal conductivity k measured in a soil system and calculated by the de Vries method, which includes the effect of latent heat transfer by vapor distillation.

The total volumetric heat capacity of the soil system and the latent heat of vaporization at temperature T can be defined as

$$C = (1 - \eta) \rho_s C_{p,s} + \theta_\ell \rho_\ell C_{p,\ell} + \theta_v \rho_\ell C_{p,v} \quad (28)$$

$$h_{fg}(T) = h_{fg}(T_o) - C_{p,\ell} (T - T_o) + C_{p,v} (T - T_o) \quad (29)$$

With the substitution of Eq.'s (32) and (33) and substitutions for the liquid and vapor mass fluxes the final version of the governing equation for soil heat transfer is

$$C_{TT} \frac{\partial T}{\partial t} + C_{\psi T} \frac{\partial \psi}{\partial t} = \nabla \cdot (k \nabla T) + \nabla \cdot (D_{\psi T} \nabla \psi) + C_{p,\ell} \dot{\mathbf{m}}_\ell \cdot \nabla T + q''' \quad (30)$$

The heat and mass transfer equations are coupled by the temperature and moisture terms occurring in both equations. The first term on the right side of Eq. (30) represents the Fourier heat conduction defined by the thermal conductivity k^* and the effective heat conduction by thermally driven vapor diffusion. The effective thermal conductivity, k , is the value calculated by the de Vries method. The second term is the heat transfer by moisture diffusion due to the moisture gradient. The third term is the sensible heat transfer by bulk liquid flow. In a subsequent section of the paper the relative magnitudes of these terms will be compared. The first term on the left hand side of Eq. (30) relates the change in the thermal energy stored to changes in temperature and the second term relates the change in thermal energy stored to changes in the matric potential.

5. Surface Energy Balance

An energy balance at the ground surface is defined by

$$q_g + q_{sens} + q_{rad} + q_{conv} + q_v = 0 \quad (31)$$

The first term in Eq. (31) represents heat conduction in the ground due to gradients of temperature and matric potential.

$$q_g = -D_{TT} \frac{\partial T}{\partial n} - D_{\psi T} \frac{\partial \psi}{\partial n} \quad (32)$$

The sensible heat, q_{sens} , is divided into two sources: that due to liquid transfer across the boundary, q_ℓ , and another term, q_h , to account for other miscellaneous heattransfer sources. The temperature of the liquid, T_ℓ , at the atmospheric boundary is assumed to be equal to the ambient dry-bulb temperature and the reference temperature is T_o .

$$q_{\text{sens}} = q_\ell + q_h \quad (33)$$

$$q_\ell = u_\ell \rho_\ell C_p (T_\ell - T_o) \quad (34)$$

The radiation term, q_{rad} , can be divided into absorbed short-wave radiation from the sun, q_{sw} , and the long-wave radiation exchange between the ground and the atmosphere, q_{lw} , and is discussed further in [1]. The long-wave radiation exchange model includes the effects of cloud cover following Martin and Berdahl [17].

The heat and mass transfer processes from the ground to the air behave similarly and, thus, the governing equations for q_{conv} and q_v have a similar form. The basic equations for convective heat and vapor transfer are

$$q_{\text{conv}} = \rho_a C_{p,a} D_h (T_{\text{amb}} - T_s) \quad (35)$$

$$q_v = h_{fg} D_v (\rho_{v,\text{amb}} - \rho_{v,s}) \quad (36)$$

The convection and evaporation depend on the wind speed, the surface conditions, and the gradients of temperature and vapor density, and are discussed in more detail in [1].

6. Surface Moisture Balance

The moisture balance at the ground surface is

$$\dot{m}_{m,g} + \dot{m}_\ell + \dot{m}_v = 0 \quad (37)$$

The ground moisture transfer is represented by Eq. (38). This term includes the liquid transfer due to the gradient of the total potential, Φ , and vapor transfer due to gradients of the matric potential and temperature.

$$\frac{\dot{m}_{m,g}}{\rho_\ell} = -K \frac{\sigma \Psi}{\partial n} - D_{\psi v} \frac{\sigma \Psi}{\partial n} - D_{Tm} \frac{\sigma T}{\partial n} \quad (38)$$

The liquid transfer at the surface is assumed to be in the form of precipitation. Moisture is assumed to accumulate up to the point of saturation, and any additional moisture is assumed to run off. The main form of vapor transfer is by evaporation from the surface.

$$\dot{m}_v = D_v (\rho_{v,\text{amb}} - \rho_{v,s}) \quad (39)$$

Using aerodynamic resistance alone has been shown to over-predict the evaporation from bare soil [18] and from plant canopies [19]; therefore, a surface resistance is added in series, as recommended by Camillo and Gurney [18].

7. Finite Element Formulation

Because of the complexity and nonlinearity of the heat and moisture transfer equations, an analytic solution would be extremely difficult to obtain. A numerical solution is obtained using a finite element method (FEM) formulation. The heat transfer equation, Eq. (30), and the moisture transfer equation, Eq. (24), are transformed into the following equations for the FEM.

$$\mathbf{D}_{TT} \mathbf{T} + \mathbf{D}_{\psi T} \boldsymbol{\Psi} + \mathbf{C}_{TT} \dot{\mathbf{T}} + \mathbf{C}_{\psi T} \dot{\boldsymbol{\Psi}} + \mathbf{f}_T = 0 \quad (40)$$

$$\mathbf{D}_{Tm} \mathbf{T} + \mathbf{D}_{\psi m} \boldsymbol{\Psi} + \mathbf{C}_{Tm} \dot{\mathbf{T}} + \mathbf{C}_{\psi m} \dot{\boldsymbol{\Psi}} + \mathbf{f}_\psi = 0 \quad (41)$$

The two sets of equations (40) and (41) can be combined into one system assembled by alternating equations for T and ψ for each node to reduce the bandwidth of the coefficient matrices as shown below.

$$\mathbf{D}\mathbf{U} + \mathbf{C}\dot{\mathbf{U}} + \mathbf{f} = 0 \quad (42)$$

A single-step algorithm for the transient analysis is presented in Eq. (43), where the subscript k denotes the time step and β is a time-weighting function, which determines whether the method is explicit or implicit [20]. All of the analyses completed in this work applied a time-weighting function of 0.5, which is analogous to a Crank-Nicolson implicit routine. The forcing function is averaged assuming a linear variation in time as in Eq. (44).

$$[\mathbf{C} + \beta \Delta t \mathbf{D}] \mathbf{U}_{k+1} = [\mathbf{C} - (1 - \beta) \Delta t \mathbf{D}] \mathbf{U}_k - \Delta t \bar{\mathbf{f}} \quad (43)$$

$$\bar{\mathbf{f}} = \mathbf{f}_k + \beta(\mathbf{f}_{k+1} - \mathbf{f}_k) \quad (44)$$

Equation (42) is nonlinear because the coefficient matrices are functions of the dependent variables. The solution of the moisture transfer equation is also subject to mass balance errors due to the highly nonlinear relationship between the moisture content and the matric potential [21,22]. According to Celia, et al. [22] diagonalizing or lumping the capacitance matrix is necessary to ensure a non-oscillatory and mass conservative solution. The nonlinear equations are solved using a modified Picard iteration method. A modified skyline storage technique is used to reduce the storage requirements. The solution of Eq. (43) is carried out using a Gaussian elimination routine.

8. Verification and Validation

The model was successfully verified against the analytic solution to a two-dimensional heat conduction problem of a rectangular region initialized at $T = 0.0^\circ\text{C}$ and boundary conditions of $T = 0.0^\circ\text{C}$ for three sides and $T = 100^\circ\text{C}$ for the fourth side [1]. The program was also verified, with excellent agreement, against the analytical solution of an isothermal moisture infiltration problem presented by Philip [23].

To validate the heat and moisture transfer models and the atmospheric boundary condition models, they were compared with field data. Weather data and ground temperatures at depths to 1 m deep were measured from October 29, 1998, to June 7, 1999, in an open field on the Colorado State University campus. A one-dimensional ground temperature simulation was conducted to a depth of 3.0 m, with Solar Village clay [24]. The simulation period extended from December 28 to June 8. This represented a time period with good data and minor snow cover. The mesh varied from 0.74 cm at the surface and increased with depth into the ground. The upper boundary was modeled using the measured weather data. Assumptions about the surface conditions were: ground cover height = 0.1 m, ground infrared emissivity = 0.9, and ground albedo = 0.23, which is representative of average crop cover. The precipitation data was obtained from a nearby weather station. Prior to December 28, a large snowfall had just melted and so the matric potential initial

condition was set at $\psi(z) = -0.1 + z*0.05$ to simulate near-saturated soil. The lower boundary condition was set with a temperature $T = 10^\circ\text{C}$ and matric potential $\psi = -0.25\text{ m}$.

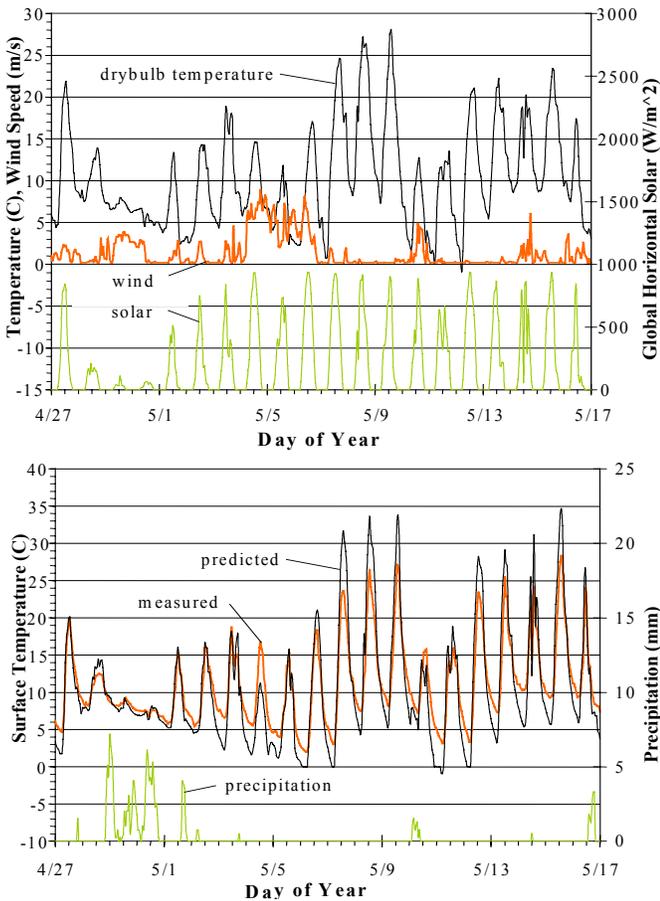


Figure 4. Atmospheric conditions and a comparison of the measured and computed surface temperatures

The weather conditions, along with the measured and predicted surface temperatures for the period between April 27 and May 17, are shown in Fig. 4. Predicting the surface temperature is very sensitive to the atmospheric boundary conditions and is often too high or too low. The over-predicted surface temperatures during the day seem to be caused by a combination of the incident solar radiation term and the convection term. During the period from April 28 to May 2, a total of 12.8 cm of precipitation was measured at the weather station. Opaque cloud cover was assumed for one hour before, during, and for two hours after any precipitation.

Figures 5 and 6 show the measured and predicted temperatures along with the root mean square error at depths of $z = -0.34\text{ m}$ and -0.95 m . The curves show the decrease in the temperature fluctuation with depth and the seasonal warming of the soil. The predicted values at -0.34 m are very close to the measured temperatures with the root mean square error less than 1.0°C . The values at -0.95 m are also

good, but the error is slightly larger. The error is the least during the spring and summer. If the precipitation is not included in the model, the calculated ground temperatures are larger than the measured temperatures, and the error increases in time. Incorporating the precipitation allows for a more accurate computation of the soil moisture, and thus the soil thermal conductivity.

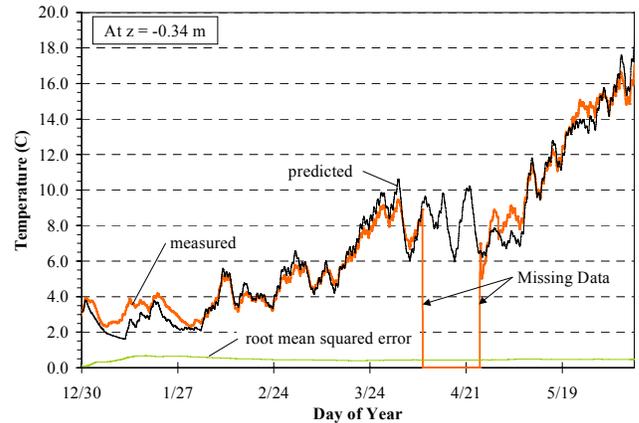


Figure 5. Comparison of measured and predicted soil temperatures at $z = -0.34\text{ m}$

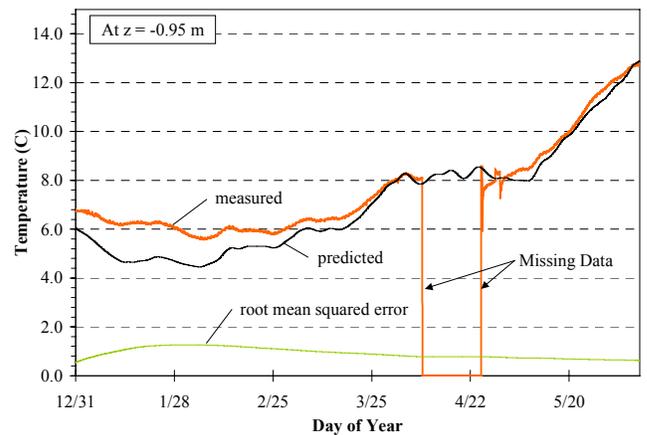


Figure 6. Comparison of measured and computed soil temperatures at $z = -0.95\text{ m}$

Figures 7 and 8 show the magnitude of the four heat flux terms: heat conduction, vapor driven by temperature gradients, vapor driven by matric potential gradients, and liquid convection at depths of $z = -0.35\text{ m}$ and -0.95 m . A positive value represents heat flow into the ground from the atmosphere. At $z = -0.35\text{ m}$, the maximum heat flux into the ground was about 40 W/m^2 , while at $z = -0.95\text{ m}$, the maximum heat flux was about 6 W/m^2 . The conduction heat transfer is usually the dominant term. Note the seasonal change in sign of the conduction term on May 10, from negative to positive heat flow. After precipitation, the heat flux by liquid convection, which is normally very small, dominates for a short time. The heat transfer by

vapor distillation is small at these depths, because of the relatively cold soil temperatures, and high moisture contents. The heat transfer of vapor driven by matric potential gradients was two orders of magnitude smaller than the heat transfer by vapor driven by temperature gradients, and is not plotted here.

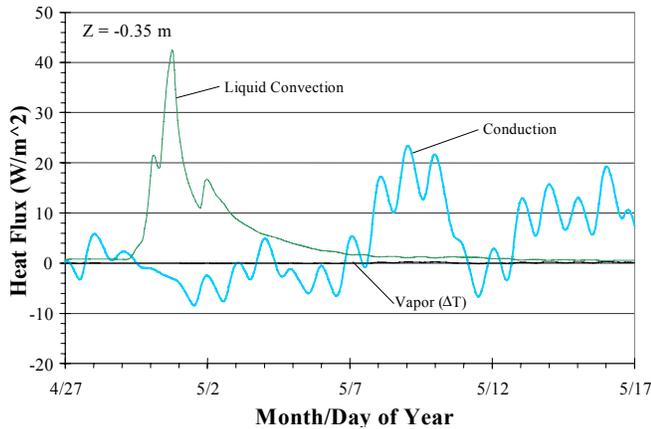


Figure 7. Conduction, vapor distillation, and liquid convection heat flux terms at $z = -0.35$ m.

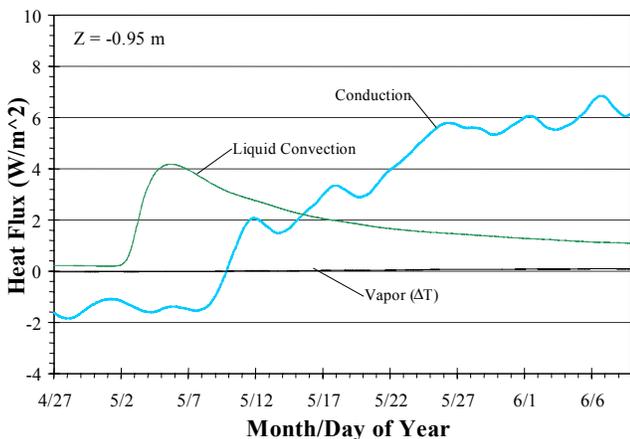


Figure 8. Conduction, vapor distillation, and liquid convection heat flux terms at $z = -0.95$ m.

9. Summary and Conclusions

A heat and moisture transfer model was developed to study the coupled heat and moisture transfer in soils. Using a matric potential formulation, the soil properties are expressed as functions of soil type and moisture content. The model includes a detailed treatment of the surface heat and moisture balances and includes the effects of vaporization in the soil and surface precipitation. The program compared well with measured ground temperatures for a one-dimensional field test case. The conduction heat transfer was usually the dominant term. After a period of precipitation, the heat flux by liquid convection, which was normally small, dominates for a short time.

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