Quantification of subsurface thermal regimes beneath evaporating porous surfaces

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ARTICLE INFO

Article history:
Received 10 June 2010
Received in revised form 12 April 2011
Accepted 26 April 2011

Keywords:
Evaporation
Porous media
Infrared thermography
Heat and mass transfer

ABSTRACT

The spatial distribution and dynamics of non-uniform evaporative mass fluxes from porous media are quantified remotely from induced surface thermal field obtained by spatially resolved infrared thermometry. Conversion of surface thermal signatures to the estimation of evaporation fluxes hinges on knowledge of the unobservable characteristic thermal decay depth. We considered quasi-static evaporation fluxes combined with measured surface thermal fields within a surface energy balance model to obtain a general analytical approximation for the thermal decay depth. This approximation was experimentally evaluated using side view IR imagery of temperature fields beneath an evaporative surface. The analytical approximation reveals dependency of thermal decay depth on the magnitude of evaporative flux which was also sensitive to convection and radiation intensity. Results indicate that typical values of evapo-thermal decay depth for a wide range of natural soil surfaces is in the range of a few centimeters (<5 cm). The solution enables remote estimation of non-uniform evaporative fluxes from surface temperature fields [1].

1. Introduction

Radiative and thermal fluxes on terrestrial surfaces and their partitioning to various components of the energy balance are key drivers for many land–atmosphere exchange processes especially the hydrologic cycle. Quantification of the resulting highly dynamic and spatially variable latent and sensible heat fluxes is one of the central challenges for modern hydrology and remote sensing [2]. The interest in remote estimation of spatially and temporally variable evaporative fluxes from porous surfaces is not limited to hydrology, it is of interest to wood, paper and food processing, waste isolation, construction and material engineering, biomedical and many more applications [3–6].

Evaporation rate from a porous medium often exhibits complex dynamics determined by interactions between internal transport mechanisms and media properties. Early stages of evaporation from an initially-saturated homogeneous porous medium are often marked by high and relatively constant drying rate and the formation of a drying front receding into the medium. Capillary flow along hydraulically connected pathways extending from the drying front to the evaporative surface sustains high evaporation rates [7]. At a certain drying front depth, hydraulic continuity is disrupted resulting in a marked reduction in evaporation flux to lower values sustained by vapor diffusion [8].

The presence of textural contrasts may give rise to additional form of hydraulic coupling within the porous medium in which internal capillary gradients induce lateral flows from coarse textured to fine textured regions. In such heterogeneous porous media, an evaporation drying front may propagate exclusively in the coarse domain leaving the fine-textured domain saturated and fully coupled with the atmosphere. Lehmann and Or [9] formulated the behavior of such coupled systems based on the intrinsic capillary and transport properties of the porous media forming the texturally-contrasting domains.

Not surprising, the resulting evaporation fluxes from texturally-heterogeneous systems are spatially non-uniform reflecting internal lateral exchanges and higher evaporation rates from fine textured regions supporting evaporation rates imposed by external conditions. Such surface flux variability would also be reflected by the spatially variable surface temperature which, in turn, would facilitate thermally-based estimation of evaporation rates and their spatial distribution. Shahraeeni and Or [1] exploited texturally-induced evaporative flux variations in a cylindrical geometry to develop a physically-based method for mapping surface temperature field to evaporation fluxes using surface energy balance equation:

\[
\dot{e}(\bar{r}, t) = \frac{\Delta z}{\rho_v \Delta L_{fr}} \left\{ \sigma [T^4_e(\bar{r}, t) - T^4(\bar{r}, t)] + h[T_e(\bar{r}, t) - T(\bar{r}, t)] - \rho C_p \frac{DT(\bar{r}, t)}{Dt} + k_T \nabla^2 T(\bar{r}, t) \right\}. \tag{1}
\]
Eq. (1) expressed for the evaporation domain extended to the depth $\Delta z$ was used to provide the spatial and temporal evaporative fluxes from temperature fields obtained by infrared imagery. Certain physical properties such as: $\rho$ bulk medium density; $c_p$ medium heat capacity; and $k_T$ medium thermal conductivity are dependent on the water saturation in the evaporation zone. The penetration depth of surface evaporation $\Delta z$ was assumed to be a medium property lumping heat transfer beneath the evaporative surface. Using common soil surface parameters and evaporation rates support the observational grander and Hanks [10] predicting a narrow range for $\Delta z$ in soils between 20 and 30 mm, a value that was used in the work of Shahraeeni and Or [1] predicting the integral form of the inversion approach based on Eq. (1) termed MTERF approximation, implements average surface temperatures of regions (patches) with similar evaporative fluxes as depicted in Fig. 1. It is worth mentioning that Eq. (1) reflects laboratory conditions without direct exposure to sun radiation, where shortwave radiation is negligible; however the general method could be modified to include short wave radiation on natural surfaces [1].

Infrared thermography (IRT) offers distinct advantages for obtaining spatially and temporally resolved surface thermal information. Advances in IRT led to a wide range of noncontact techniques for thermal detection of various processes and surface properties [11,12]. Hydrological applications of IRT include the study of water movement in porous media [13,14], mapping surface temperature and circulation patterns in the lakes [15], and the modeling of heat dispersion in runoff water [16]. A similar approach is often used for larger spatial scales to remotely sense land–atmosphere exchanges at regional and global scales using IRT measured skin temperature to estimate surface moisture availability, evaporation rate, sensible heat flux and thermal inertia [17–20].

Due to challenges presented by atmospheric attenuation and filtering at large scales, a combination of infrared and microwave radiations and empirical algorithms are routinely used in land-surface hydrology and hydrometeorology fields [21–25]. Howington [26] recently developed an adaptive hydrology model (ADH) to represent different processes affecting an infrared image of the soil. Lin et al. [27] used infrared remote-sensing-based field scale soil moisture simulator to initialize soil moisture dynamics. Moran et al. [28] used IR data for partitioning between different contributors of evaporapotranspiration. Gaikovich and Troitsky [29] studied dynamics of heat and mass transfer through an interface between air and water and developed a method of remote monitoring based on the joint solution of thermal radiation and heat conductivity equations to analyze data on heat and mass exchange dynamics at the air–water interface. Entekhabi et al. [30] developed a physically-based inverse algorithm for retrieving soil moisture and temperature profiles based on remotely sensed observations of multispectral irradiance. IRT has also been used to detect fractures, cracks and other deficiencies in the materials [31]. Recently, hydraulically active fracture systems were identified based on a combination of infrared thermography, fracture mapping and regional tectonic and hydrologic knowledge [14,32].

A main challenge for remote estimation of mass exchange from evaporative surface is the estimation of unobservable variables beneath the surface required for energy balance closure (i.e., temperature profile and heat flux). At the field scale, inverse simulation based on SVAT models that implement experimental data for the estimation of penetration depth of evaporation have been implemented [33–35]. However, a general solution of the closure problem requires a physically-based approach and control experiments (lab scale) for testing model predictions. The objectives of this study were:

1. To formulate a model for energy and mass exchanges linking measured evaporative fluxes from porous surfaces with the resulting temperature fields.
2. To develop an analytical approximation for characteristic thermal decay depth based on porous media properties and external inputs.
3. To validate the model using experimental data from temperature field beneath an evaporative surface.

Following this introduction, we present in part (2) theoretical considerations including the solution of energy balance equation for thermal fields beneath an evaporative surface. Part (3) presents experimental methods with a brief introduction to the infrared

**Nomenclature**

- $u_{in}, C_m$: coefficients of Fourier solution
- $B$: domain characteristic length (m)
- $B^r$: dimensionless domain characteristic length ($-$)
- $b$: evaporative patch characteristic length (m)
- $c_p$: heat capacity (J/kg K)
- $D$: material derivative ($\frac{\partial}{\partial t} + \mathbf{v} \cdot \nabla$)
- $e$: evaporation rate (mm/day)
- $h$: heat transfer coefficient (W/m² K)
- $h_L$: heat transfer coefficient averaged over the whole surface (W/m² K)
- $k_a$: thermal conductivity of air (W/m K)
- $k_T$: thermal conductivity of partially saturated soil (W/m K)
- $L$: domain length (m)
- $\Delta E_{ev}$: latent heat of vaporization for water (kJ/kg)
- $N_{U_{ext}}$: extended Nusselt number ($-) = \left( \frac{h}{k_T} \right) T^r_L$
- $N_{U_k}$: averaged Nusselt number ($-) = \left( \frac{h}{k_a} \right)$
- $\Pr$: Prandtl number ($-) = \left( \frac{\nu}{\alpha} \right)$
- $r$: coordination vector (m)
- $Re$: Reynolds number ($-) = \left( \frac{\rho U L}{\mu} \right)$
- $T$: surface averaged temperature (K)
- $T_v$: volume averaged temperature (K)
- $t$: time (s)
- $T_{aw}$: ambient air temperature (K) (=298 K)
- $U$: air velocity (m/s)
- $\chi, z$: coordination variables (m)
- $x^*, z^*$: dimensionless coordination variables (m)

**Greek symbols**

- $\Delta z$: the penetration depth of surface evaporation (m)
- $\Delta T_0$: maximum temperature difference between surface and ambient air (K)
- $\varepsilon$: emissivity ($-$)
- $\lambda_c$: decay constant (1/m)
- $\mu$: air viscosity (Pa s)
- $\rho$: averaged partially saturated soil density (kg/m³)
- $\rho_w$: water density (kg/m³)
- $\rho_a$: air density (kg/m³)
- $\nu$: Laplacian of temperature (K/m²)
- $\sigma$: Stefan – Boltzmann constant (W/m² K⁴)
- $s$: superscript of dimensionless parameters
thermography followed by experimental setup. Part (4) presents results of the measured temperature field and the calculated evaporation patterns for the method verification comparing temperature-based mass loss estimates with direct mass balance results and with the analytical solutions, followed by summary and conclusions in part (5).

2. Theoretical considerations

Fig. 2 illustrates a schematic of the problem considering side view of a finite Cartesian domain \(|x| < B, -\infty < z < 0\) with heterogeneous porous media properties which resulting in a patchy evaporative pattern \(\hat{\varepsilon}(x)\) from the top surface \((z=0)\). Under steady state condition, one can formulate heat transfer due to surface evaporation as:

\[
\nabla^2 T = 0 \text{ for } |x| < B, -\infty < z < 0
\]

\[
k \left. \frac{\partial T}{\partial z} \right|_{z=0} = h(T_\infty - T) + \sigma\varepsilon(T_\infty^4 - T^4) - \rho_w \Delta E_v \varepsilon(x)
\]

\[
\left. \frac{\partial T}{\partial x} \right|_{x=0} = \left. \frac{\partial T}{\partial z} \right|_{z=-\infty} = 0
\]

The only nontrivial surface boundary condition for conservation of energy is to equate vertical heat conduction with convection and radiation terms plus energy flux due to the heterogeneous evaporation field. For the other three boundaries, homogeneous Neumann boundary conditions \(\partial T/\partial x|_{x=} = 0\) and \(\partial T|_{z=\infty} = 0\) are assumed. To derive an analytical solution for this problem, one should first linearize the radiation term. Expanding \(T\) around \(T_\infty\), \(T_\infty^4 - T^4\) could be expressed as \(4T_\infty^3(T_\infty - T)\) which is in good approximation for common range of temperatures between 0 and 60 °C in terrestrial systems. Fig. 3 compares linearized radiation term and nonlinear one for the ambient temperature \(T_\infty\) of 298 K.

Physical properties of heat conductivity \(k\), convection coefficient \(h\), and emissivity \(\varepsilon\) depend on the texture and phase distribution in the porous medium as well as on the air flow regime applied. These are known to vary across the domain, however as a first approximation the use of averaged values facilitate derivation of an analytical solution. Separation of variable yields analytical solution of (2) as:

\[
T(x,z) = \sum_{m=0}^{\infty} C_m \exp \left( \frac{m\pi z}{B} \right) \cos \left( \frac{m\pi x}{B} \right),
\]

where:

\[
C_m = \left\{ \begin{array}{ll}
\frac{\varepsilon_0 a_m}{h + 4\varepsilon_0 \kappa_k} & m = 0 \\
\frac{a_m}{v_3} & m \neq 0 \\
\end{array} \right.
\]

\[
am_m = \frac{2\rho_w \Delta E_v \sin \left( \frac{m\pi h}{B} \right)}{v_3}
\]

and \(a_m\) are cosine Fourier coefficients of the mass loss distribution on the evaporative plane.

For a spatially-distributed evaporation rate \(\dot{\varepsilon}(x)\), Eq. (3) describes the resulting temperature distribution underneath of the evaporative surface. For example, considering evaporation from a uniform patch with constant \(\varepsilon_0\) over \(-b < x < b\), \(a_0 = \frac{2\rho_w \Delta E_v \varepsilon_0}{v_3}\)

and \(a_m = \frac{2\rho_w \Delta E_v \varepsilon_0}{v_3} \sin \left( \frac{mb\pi}{B} \right)\), yields the following temperature field (surface and depth):
\[ T(x,z) = T_\infty - \frac{\rho_w \Delta E_\varepsilon \varepsilon \varepsilon_b}{B(h + 4\sigma \varepsilon T_\infty^4)} - 2\rho_w \Delta E_\varepsilon \varepsilon \varepsilon_b B \sum_{m=1}^\infty \frac{1}{m\pi k_l + B(h + 4\sigma \varepsilon T_\infty^4)} \sin\left(\frac{m\pi b}{B}\right) \cos\left(\frac{m\pi x}{B}\right). \]

The parameters and physical properties can be estimated based on the particular porous medium and boundary conditions. For thermal conductivity \( (k_T) \), Chen [36] proposed empirical expression for \( k_T \) as a function of medium porosity \( (n) \) and saturation level \( (S) \):

\[ k_T(n, S) = 7.5^{1.061} \cdot 0.61^{0.9922} \cdot 0.0022 \cdot 0.78n. \]  

The averaged value of thermal conductivity \( (k_T) \) predicted by (5) over a wide range of saturations \( (0.1 \leq S \leq 1) \) and porosities \( (0.2 \leq n \leq 0.8) \) is about 1.8 W/m K which is used in subsequent analysis. The thermal convection coefficient depends on air flow regime over the surface which may be deduced from correlations based on the Nusselt number and on flow characteristics. For parallel laminar flow over a flat plate considering Reynolds number less than \( 5 \times 10^5 \), the mean Nusselt number is [37]:

\[ \text{Nu}_L = \frac{hL}{k_0} = 0.664\text{Re}^{1/3}\text{Pr}. \]  

For an air flow velocity \( u = 1.5 \text{ m/s} \) and \( T_\infty = 298 \text{ K} \) over an evaporative surface with characteristic length of 1 m \( (\text{Re} \approx 96'000 \text{ and } \text{Pr} = 0.7296) \), the mean Nusselt number is about 185 and the corresponding thermal convection coefficient is about 4.63 W/m\(^2\) K.

Surface emissivity \( (\varepsilon) \) also depends on the phase distribution in the porous medium. Zhang [38] provides emissivity data for a collection of natural and artificial materials. An averaged value of 0.925 is considered for the partially saturated sand with the grain size distribution in the range of 100 \( \mu \text{m} \) to 1 mm. Considering the aforementioned physical properties, Fig. 4 shows predicted

Fig. 2. Definition sketch for heat and mass transfer during evaporation from a texturally heterogeneous medium with associated non-uniform evaporation flux pattern from the top surface.

Fig. 3. Comparison of the linearized radiation term with the original nonlinear one with \( T_\infty \) equal to 298 K (room temperature).

Fig. 4. Temperature field beneath an evaporating surface calculated by Eq. (4) for a finite medium \( B = 1 \text{ m} \) with heat conductivity \( k_T = 1.8 \text{ W/m K} \), emissivity \( \varepsilon = 0.925 \) and \( \varepsilon = 13 \text{ mm/day from } b = 10 \text{ cm which is exposed to an air flow characterized by } T_1 = 25 \text{ ºC and } V_1 = 1.5 \text{ m/s.} \)
temperature field resulted from Eq. (4) for a finite domain $B = 1$ m with an evaporative patch $e(x) = 13$ mm/day in $|x| < 10$ cm of a semi-infinite domain $-B < x < B$ and $-\infty < z < 0$.

The exponential decay of temperature depression in the vertical direction emerges from Eq. (4). Dimensional analysis of Eq. (2) shows decaying modes of radiation, convection and evaporation with decaying constants of $4 \sigma z T_\infty^4 / k_\xi$, $h / k_1$ and $-\rho_w \Delta E e e / k_1 T_\infty$, respectively, resulting in a decay constant of the entire process as:

$$
\lambda_c = \frac{\text{const.}}{k_1} \left( h + 4 \sigma z T_\infty^4 - \frac{\rho_w \Delta E e e}{k_1 T_\infty} \right).
$$

(7)

Thus, defining the depth of evaporation zone (thermal decay depth) as the length over which temperature difference decays to less than 1% of the ambient temperature, Eq. (4) in geometry similar to Fig. 4 yields an estimate as:

$$
\Delta h = \frac{\text{const.} \cdot k_1}{h + 4 \sigma z T_\infty^4 - \frac{\rho_w \Delta E e e}{k_1 T_\infty}}.
$$

(8)

Eq. (8) provides a general estimate for the thermal decay depth and thus is the key to employing a lumped transient form of the energy conservation equation. Based on the distribution of Fig. 4, $\text{const.}$ would be calculated equal to 0.15. We consider evaporation as a quasi-static process in which at each time step steady state solution (3) yields temperature field based on the instantaneous evaporation rate of that time step. Therefore the penetration depth of evaporation ($\Delta z$) would be changed by the change of the evaporation rate in the course of a drying experiment. A consequence of the uniform evaporation field is that a correspondingly uniform surface temperature field with $T_s$ as the mean temperature of the evaporative surface ($T_s < T_\infty$). Considering “evaporation zone” as the volume with thickness $\Delta z$ in which temperature changes along the vertical direction from surface temperature ($T_s$) to 0.99$T_\infty$ (the ambient temperature), mean temperature of the evaporation zone ($T$) could be nearly linearly related to $T_s$, as $T = 2T_s - T_s$. Based on the mean temperature of the evaporative surface, transient energy balance can be written as:

$$
-\frac{k_1 dT_s}{dz} \bigg|_{z = -\Delta z} + \sigma e (T_\infty^4 - T_s^4) + h(T_\infty - T_s) - \rho_w \Delta E e e \frac{d(T_s)}{dT_s} = \frac{\rho c_p \Delta h}{2} \frac{dT_s}{dT_s}.
$$

(9)

Dimensional analysis of Eq. (9) provides more insight regarding the relative importance of different terms. Using $b$ (length scale of the evaporation zone), $T_\infty$ (ambient temperature) and $\Delta T_0 = \rho_w \Delta E e e (h + 4 \sigma z T_\infty^3)^{-1}$ (maximum temperature difference to the ambient achievable on the top surface due to the maximum evaporation rate $e_{\text{max}}$) and $t_0 = \rho c_p (h + 4 \sigma z T_\infty^3)^{-1}$ (time scale of the evaporation process) as references, the nondimensional form of Eq. (9) is:

$$
\tilde{e}^* - \tilde{T}_s^* = \frac{1}{2k_1 N_u \text{ext}} \left( \frac{dT_s}{dz} \bigg|_{z = -\Delta z} - \frac{dT_s}{dT_s} \right) = 0.
$$

(10)

The asterisks quantities are nondimensional defined as:

$$
T^* = \frac{T_s - T_\infty}{\Delta T_0}, \quad \Delta T_0 = \frac{\rho_w \Delta E e e e_{\text{max}}}{h + 4 \sigma z T_\infty^3},
$$

$$
\tilde{e}^* = \frac{e}{e_{\text{max}}}, \quad \tilde{T}_s^* = \frac{T_s}{T_\infty}, \quad \tilde{z}^* = \frac{z}{b}, \quad \tilde{t}^* = \frac{t}{t_0}, \quad t_0 = \frac{\rho c_p b}{h + 4 \sigma z T_\infty^3},
$$

$$
N_u \text{ext} = \frac{b(h + 4 \sigma z T_\infty^3)}{k_1}, \quad k^* = \frac{k_s}{k_1}.
$$

(11)

We implicitly assume that $\tilde{e}^* \Delta T_0 / T_\infty = 0$ since evaporative temperature depression ($\Delta T_0$) is negligibly compared with the ambient absolute temperature ($T_\infty$). From Eq. (10) it is clear that the contribution of conductive and storage terms are controlled by the dimensionless group $N_u \text{ext}$ designating as the extended Nusselt number considering radiation effect. Qualitatively, high Nusselt number diminishes the roles of conduction and storage in the energy balance equation. Higher convection or radiation flux on the surface, as well as higher evaporation patch length, increase Nusselt number values, whereas medium thermal conductivity can increase the importance of storage and conduction terms in the partitioning of energy in the evaporation zone.

Eq. (10) results in evaporation rate according to:

$$
\tilde{e}^* = \tilde{T}_s^* + \frac{1}{2k_1 N_u \text{ext}} \left( \frac{dT_s}{dz} \bigg|_{z = -\Delta z} + \frac{dT_s}{dT_s} \right).
$$

(12)

Since $\Delta z$ is defined as the depth over which temperature reaches 99% of its final value, the first term in the parenthesis could be ignored resulting in the instantaneous evaporation rate estimate in the dimensional form as:

$$
\dot{e}(t) = \frac{1}{\rho_w \Delta E e e} \left( h + 4 \sigma z T_\infty^3 (T_\infty - T_s) - \rho c_p k_1 \frac{dT_s}{dT_s} \right).
$$

(13)

The importance of Eq. (13) is its independence of unobservable subsurface thermal information. Thus temporal changes in surface temperature enable estimate of associated instantaneous evaporation rate. Furthermore, ignoring the relatively small storage term results in a very simple form:

$$
\dot{e}(t) = \frac{h + 4 \sigma z T_\infty^3}{\rho_w \Delta E e e} (T_\infty - T_s).
$$

(14)

Eq. (14) provides the simplest estimate of uniform evaporation rate as a function of surface temperature which is now solely based on external controlling parameters and surface temperature. Away from the boundary within an evaporative patch, Eq. (14) could be used to predict the evaporation rate. However, near a patch boundary the lateral heat exchange between domains with different evaporation patterns takes place which is considered by extending Eq. (10):

$$
\dot{e}^* - \tilde{T}_s^* = \frac{1}{2k_1 N_u \text{ext}} \left( \frac{dT_s}{dz} \bigg|_{z = -\Delta z} + \frac{dT_s}{dT_s} \right) = 0.
$$

(15)

A similar reasoning as previously enables the elimination of vertical component of the conductive flux, however the lateral conduction flux must be calculated to deduce evaporation rate from temperature measurements. Assuming quasi-static evaporation rate enables the use of temperature distribution (4) for a steady state uniform evaporative flux as a first estimate of $dT_s / dx'$. Nondimensional $dT_s / dx'$ from Eq. (4) yields:

$$
\frac{dT_s}{dx'} \bigg|_{x' = 1} + \frac{2 \epsilon^*}{B} \sum_{m=1}^{\infty} \frac{\sin^2 \left( \frac{m \pi}{B} \right)}{m \pi} \exp \left( \frac{m \pi \epsilon^*}{B} \right) = 0.
$$

(16)

Integrating (16) along vertical depth of the evaporation zone provides the mean thermal flux conducted laterally through the boundary $x' = \pm 1$ which based on the averaged surface temperature $T_s$ results in:
Changes in evaporation rate from a surface patch based on surface temperature gradient coefficient in Eq. (17) as a function of dimensionless length of the domain \( B^* \) for different values of \( k^* N_{\text{ext}} \).
Ambient temperature and relative humidity were measured using an anemometer placed about 5 cm above the evaporating surface. A blower with three different speeds blows air over the evaporative surface. IR image of the evaporation front was captured with a high resolution camera. Data from the experiments included velocity, temperature, and relative humidity measurements. The data loggers measured velocity, temperature, and relative humidity of the ambient air. Two different experiments with different boundary conditions were conducted during which the evaporation demand varied across a wide range of values (from 1 to 16 mm/day). Measured evaporation rates versus temperature difference between the evaporative surface and the ambient air are presented in Fig. 8.

The slope of the line in Fig. 8a is 5.61 mm/day K, considering recorded room temperature of 298 K, gives h equal to 4.34 W/m² K which in turn results in u = 1.32 m/s over the surface from Eq. (6). This is in agreement with independently measured velocity of 1.5 m/s by the hot wire anemometer. In Fig. 8b the slope decreases to 3.21 mm/day K which in similar room temperature resulted in \( h = 2.45 \) W/m² K, and \( u = 0.42 \) m/s. The recorded mean air velocity by anemometer in this case was 0.5 m/s.

4.2. Measured and predicted thermal field below an evaporative surface

For evaporation from uniform surface as the system approached steady state conditions (as indicated by high concentration of data points in Fig. 8), Eq. (4) predicts temperature profile below the surface for sufficiently large B (characterizing the entire domain) for an evaporative patch with \( b = 9 \) cm enabling assumption of uniform temperature field \( T(x, z) \) in Eq. (4) around \( x = 0 \). Fig. 9a depicts a comparison between predictions by Eq. (4) and experimental data for high demand (strong convective flow) corresponding to the steady state evaporation condition of Fig. 8a. To capture the curvature of the experimental temperature profile, we used a constant evaporation rate \( \varepsilon_0 \) in Eq. (4) assumed to vary between 1.1 mm/day and 2.1 mm/day for high evaporative demand experiment (Fig. 8a). Fig. 9b is the same data for the case of low demand – mild convective flow corresponding to the case of Fig. 8b. In this case the constant evaporation rate assumed to vary between 2.7 mm/day and 0.9 mm/day in Eq. (4) which results in different curves of Fig. 9b.

Temperature profile evolves in time by the change of evaporation rate and ultimately would be equal to the ambient temperature when evaporation stops on the top surface of the medium. Since Eq. (4) assumes steady state heat transfer in Eq. (2), deviation from steady state assumption causes error in the prediction of the temperature field beneath the evaporative surface.

4.3. Effective depth of thermal perturbation below an evaporative surface

Although the depth of evaporative thermal perturbation zone \( \Delta z \) is implicitly included in Eq. (14), it is instructive to check the validity of a direct approximation by Eq. (8) and study potential changes in \( \Delta z \) during the course of an evaporation experiment.

For evaporation from homogeneous surfaces (columns filled with coarse sand), we used Eq. (14) to predict temporal dynamics of the evaporation rate capitalizing on the linear relation with variations of temperature difference between evaporating surface and ambient air. Two different experiments with different boundary conditions have been conducted during which the evaporation demand varied across a wide range of values (from 1 to 16 mm/day). Measured evaporation rates versus temperature difference between the evaporative surface and the ambient air are presented in Fig. 8.

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4. Results and discussion

4.1. Evaporation rate and surface temperature depression temporal dynamics

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For evaporation from uniform surface as the system approached steady state conditions (as indicated by high concentration of data points in Fig. 8), Eq. (4) predicts temperature profile below the surface for sufficiently large B (characterizing the entire domain) for an evaporative patch with \( b = 9 \) cm enabling assumption of uniform temperature field \( T(x, z) \) in Eq. (4) around \( x = 0 \). Fig. 9a depicts a comparison between predictions by Eq. (4) and experimental data for high demand (strong convective flow) corresponding to the steady state evaporation condition of Fig. 8a. To capture the curvature of the experimental temperature profile, we used a constant evaporation rate \( \varepsilon_0 \) in Eq. (4) assumed to vary between 1.1 mm/day and 2.1 mm/day for high evaporative demand experiment (Fig. 8a). Fig. 9b is the same data for the case of low demand – mild convective flow corresponding to the case of Fig. 8b. In this case the constant evaporation rate assumed to vary between 2.7 mm/day and 0.9 mm/day in Eq. (4) which results in different curves of Fig. 9b.

Temperature profile evolves in time by the change of evaporation rate and ultimately would be equal to the ambient temperature when evaporation stops on the top surface of the medium. Since Eq. (4) assumes steady state heat transfer in Eq. (2), deviation from steady state assumption causes error in the prediction of the temperature field beneath the evaporative surface.
reaches 99% of ambient temperature. The other estimate of \( D_z \) is obtained by inserting the balance recorded mass loss rate \( e \) in Eq. (8) to calculate \( D_z \). The magnitude and dynamics of these two depth estimates are in good agreement (with some discrepancies between thermal and weighted measurements as also seen in the results of Fig. 12) essentially confirming the utility of Eq. (8) to predict \( D_z \).

4.4. Linking heterogeneous evaporation fluxes with surface thermal signatures

To examine the validity of Eq. (4) for describing thermo-evaporative processes from the heterogeneous surfaces, we have used the experimental results from heterogeneous column filled with fine and coarse sands focusing on the surface and subsurface thermal fields with the border between two domains in the middle of the PIR window. Considering for simplicity negligible thermal conduction through side walls of the column, we may represent the system by \( b = 9 \) cm and \( B = 18 \) cm using Eqs. (2) and (4). The results in Fig. 11 compare analytical solution of Eq. (4) with the experimental data for the time at which the drying front enters the fine part of the heterogeneous surface.

We selected this stage in the experiment when conditions postulated for the analytical solution in Eq. (4) are satisfied. At this stage according to [9], only a small fraction of evaporation takes place from the coarse domain of the column (mostly as vapor transport from a deep drying front with the minimal surface thermal signature), and measured mass loss by the balance could be attributed primarily to evaporation from saturated and capillary connected fine region of the column. Based on balance measurements (direct mass loss), the evaporation rate from fine-textured surface was \( e = 6 \) mm/day which was then introduced into Eq. (4) and provided the temperature distribution seen from side view (Fig. 11). Considering normal experimental heterogeneities, the resulting consistency in the shape and intensity of the surface and subsurface temperature field provides experimental support for the solution in Eq. (4).

Following experimental testing of the key theoretical elements, we may now use Eq. (19) to describe spatially variable evaporative...
fluxes from the heterogeneous column based on the surface temperature signatures. These are summarized in Fig. 12 comparing spatially resolved temperature-based evaporation rate calculations with the total evaporation rate from the column obtained from balance recorded data. Temperature based calculation for the entire surface (green curve) is the weighted average of the evaporation rate from the coarse and fine parts of the surface which are shown separately in blue line for the fine and red line for the coarse part of the surface (these are calculated using Eq. (19)). The temperature-based results are in good agreement with the balance recorded data (for the entire surface) providing a benchmark for the accuracy of the proposed method in the integral sense.

5. Conclusion

A physically based model for mass and energy exchange during evaporation from homogeneous and heterogeneous porous surfaces was developed and tested. We considered quasi static evaporation rate to derive analytical solutions for the lateral and vertical evaporative temperature depression lengths as functions of evaporative flux for different media properties and boundary conditions. This enabled consideration of inaccessible thermal information beneath an evaporative surface and resolves a closure issue for the surface energy balance equation. The solution plays a key role in resolving the energy exchange processes such as phase change, conduction, convection, radiation and storage and helps identify their relative contributions. We devised a system for direct experimental testing of subsurface temperature distribution using side view IR imagery beneath the evaporative surfaces and successfully compared the measured and predicted surface and subsurface thermal fields. Results confirm that the length scales over which temperature depression of evaporation decays are relatively small (a few centimeters) and their analytical prediction was in good agreement with the experimental data.

Approximations based on measurable mean surface temperature of patches with similar evaporation rates provide enable quantification of spatially distributed evaporation fluxes and their contribution to overall surface evaporation. This result enables remote detection by IR imagery and quantification of the evaporative fluxes from spatially heterogeneous and patchy surfaces. We examined the validity of such spatially resolved detection (in the integral sense) using evaporation from sand-filled columns placed on a balance and obtained good agreement between thermally-decuded and measured mass loss. The relative contributions of the various parts of the column were consistent with the theory and with independent (lumped) mass loss measurements. The study provides the necessary theoretical basis for the application of surface energy balance for evaporative surface considering lateral and vertical conduction and thus enables spatially resolved quantification of evaporation fluxes from hydraulically interacting heterogeneous surfaces. The results may be extended to the plot and field scales using IR remote sensing and could consider special cases with precipitation, crust formation, and additional effects on the surface evaporation.

Acknowledgements

The authors gratefully acknowledge funding by the Swiss National Science Foundation project 200021-113676/1 and funding by the German Research Foundation DFG of project (FOR 1083).
Multi-Scale interfaces in Unsaturated Soil (MUSIS). The generous assistance of Daniel Breitenstein and Dr. Peter Lehmann in various aspects of the study is sincerely acknowledged which also should be extended to Mr. Fabian Rudy for the graphical and illustration supports.

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