Finite difference modeling of heat distribution in multilayer soils with time-spatial hydrothermal properties

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ABSTRACT

In this study, the heat distribution throughout the profile of unsaturated multilayered soil is determined using finite difference method while its thermal diffusivity varies with time and depth. First, the input parameters such as water content, dry density and sand content of the soil profile are provided. These data are coupled with the theoretical approaches to estimate thermal properties of soil such as thermal conductivity and thermal diffusivity of multilayered soil. Second, finite difference method is used to model heat distributions in soil profile taking into account the initial and boundary conditions. A continuity of heat flux between each layer is performed as a condition in the numerical model. A comparison of estimated temperature within time throughout the profile with the thermal probe measurements shows a satisfactory capacity of the numerical model. Finally, different cases of nonhomogeneous and homogeneous soil show that thermal response of homogeneous and nonhomogeneous soils are almost similar at average value of thermal diffusivity where hydrothermal characteristics of each soil layer (such as water content, dry density, and soil texture) are required to calculate this average value.

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1. Introduction

A shallow geothermal energy system is considered as a significant source of thermal energy suitable for heating and cooling of individual buildings. It is a renewable source of energy as heat transfers continuously in the soil due to temperature gradient between the bottom of the earth and the air [1].

Since 1980, researchers have widely considered soils for heating or cooling purposes of seasonal thermal energy storage applications, e.g., Refs. [2–8]. Thermal energy can be stored for long time in the underground soil [9]. The development of ground coupled heat pump (GCHP) has been promoted since environmental problems exist [10–12]. Several advantages of using underground soil as heat source for GCHP compared with the traditional air source heat pump, encourage researchers to understand more about thermal behavior of soil [13–16].

Soil temperature is considered to be an important parameter in geothermal energy applications such as the passive heating and cooling of buildings including performance, dimensions, and

installation cost in many geothermal applications [17–20]. In shallow grounds, just 20 m below the surface, the soil maintains nearly constant temperature ranging between 10 and 20, depending on the region [21]. In this context, for the seasonal heat storage (studied in this work) such as earth to air heat exchanger (EAHEs), and GSHPs, the soil temperature is considered to be nearly constant for the depths higher than 20 m [22–24].

Hydrothermal properties of soil have been widely studied in geothermal issues. Soil thermal properties are important inputs for models of soil temperature, but those thermal properties vary while the water content in soil changes. Many researchers have studied the dependence of thermal diffusivity of soil with other soil characteristics as well as degree of saturation [25–30]. However, conventional schemes consider constant thermal properties to give hydrothermal response of soil [31–33]. Ozgener et al. [34] developed a model to predict daily soil temperatures using a sinusoidal function of time and depth. In their work, transient heat flow principles were used with assumptions of one dimensional heat flow, homogeneous soil, and constant thermal diffusivity. Such hypothesis cannot take into account a comprehensive understanding of the effects of thermal properties due to the fact that those properties vary with soil physical characteristics such as soil texture and water content.

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Furthermore, the commercial computer codes taking in consideration the hydrothermal properties variation are not able to attribute different hydrothermal properties of different soil layers for a same profile. It is necessary to develop a code which can calculate the variation of hydrothermal properties of multilayers for one soil profile while applying heat flux continuity at the interface of each layer of soil profile.

In this study, the input parameters of an in-situ soil such as water content, dry density, and sand content of the soil profile are initially measured for different layers with time and depth. Using theoretical approaches, thermal properties of multilayered soil such as thermal conductivity and thermal diffusivity are also determined for the different time and depth. Finally, the finite difference model is applied on this multilayered profile to capture its temperature variation with time and depth.

2. Numerical model

In this section, heat distributions throughout a multilayer profile of soil are obtained by considering variable thermal diffusivity. Fig. 1 shows a multilayer profile of soil with different soil texture and soil density in each layer i . Moreover, soil degree of saturation varies with time and depth throughout the profile.

Therefore, thermal diffusivity of soil changes with time and depth for which the transient heat distribution in soil is governed by:

$$\frac{\partial T}{\partial t} = \alpha(z, t) \left(\frac{\partial^2 T}{\partial z^2} \right) + \frac{\partial \alpha(z, t)}{\partial z} \frac{\partial T}{\partial z} \quad (1)$$

where $\alpha(z, t)$ is the thermal diffusivity of soil which varies with time and space.

Fig. 1. Node arrangement in the finite difference formulation.

The model is subjected to the boundary conditions:

$$\frac{\partial T}{\partial z}(z = L_\infty) = 0 \quad (2)$$

which shows there is no heat flux at large depth.

$$T(z = 0, t) = T_s(t) \quad (3)$$

where $T_s(t)$ is the measured surface temperature.

Moreover, at the interface between two layers, two boundary conditions must be satisfied. The continuity of temperature:

$$T_i(z = L_i, t) = T_{i+1}(z = L_i, t) \quad (4)$$

and the continuity of the heat flux:

$$-\lambda_i(z = L_i, t) \frac{dT_i}{dz}(z = L_i) = -\lambda_{i+1}(z = L_i, t) \frac{dT_{i+1}}{dz}(z = L_i) \quad (5)$$

Using finite difference method, a series of nodes within the substrate which span from the surface to a defined depth is considered as shown in Fig. 1. For each node p , there exists a thermal diffusivity value $\alpha(p)$. At a general node p , the differential terms in Eq. (1) can be approximated using the following expressions.

$$\frac{\partial T}{\partial t} = \frac{T_p^{k+1} - T_p^k}{\Delta t} \quad (6)$$

$$\frac{\partial^2 T}{\partial z^2} = \frac{T_{p+1}^k - 2T_p^k + T_{p-1}^k}{\Delta z^2} \quad (7)$$

$$\frac{\partial \alpha(z, t)}{\partial z} \frac{\partial T}{\partial z} = \frac{\alpha_{p+1}^k - \alpha_{p-1}^k}{2\Delta z} \frac{T_{p+1}^k - T_{p-1}^k}{2\Delta z} \quad (8)$$

The temporal temperature derivative Eq. (6) is evaluated as the difference between the future time step $k + 1$, and the present time step k , so that an explicit scheme is obtained. Considering that thermal diffusivities at each time k and depth p are known parameters; thus, in the finite difference scheme described by Eqs. (7) and (8), thermal diffusivities as well as spatial temperature derivatives are evaluated at time k (we have not used a distributed lag model).

The temperature T at time $k + 1$ is computed in each iteration until the difference of temperature between two iterations becomes less than 10^{-4} . Normally for the convergence, it is enough to have $\text{Max}(\alpha(z, t) dt/dz^2) < 0.5$ by considering thermal diffusivities at each time k , and depth p .

The continuity conditions are applied at the interface of each soil layer i where the soil textures and dry density of soil changed.

3. Input parameters of the numerical model

This section is dedicated to the model input parameters such as degree of saturation, thermal diffusivity and temperature within depth and time.

3.1. Field description

Since September 2012, a field located in Illkirch, Alsace in France was instrumented with five temperature probes TS₂₀, TS₂₁, TS₂₂, TS₂₃, and TH₅₀ and one humidity probe H₅. The field is composed of three in-situ soils: vegetarian soil, wind-blown sandy soil and soil backfill (see Fig. 2).

Fig. 2. Multilayered soil with different sand content x_s and dry density γ_d in each layer. TS_{20} , TS_{21} , TS_{22} , TS_{23} , and TH_{50} are thermal probes while H_s is humidity probe.

3.2. Profile of water content and degree of saturation of multilayered soil

A core sampling from the soil profile is performed as it is shown in Fig. 3. Then, water content of each sample is measured gravimetrically by over drying them at 60 °C. This initial water content profile is called $W_0(z)$. Soil at groundwater level in the field ($z = 4$ m) is considered to be in saturated state ($S_r = 100\%$). Soil water content between the deepest in-situ measured data ($z = 0.87$ m) and the ground water level ($z = 4$ m) is considered to vary exponentially until it arrives at the saturation degree at the underground water level.

Water content of soil was also measured by a humidity probe H_s at the depth of 1.11 m (Fig. 2). The value of water content is shown as a red spot in Fig. 3 which agrees well with the one measured by core sampling at the same position and time.

Humidity probe H_s measured the water content of soil during the year (see Fig. 4). A coefficient Γ which is given as:

$$\Gamma(t) = \frac{W(t)}{W(z = z_{HS})} \quad (9)$$

where $W(t)$ is the water content measured by humidity probe H_s during one year and $W_0(z = z_{HS})$ is the initial water content at the depth of the probe H_s measured at $t = 0$. By multiplying the coefficient $\Gamma(t)$ and the initial water content profile $W_0(z)$, the water content throughout the soil profile and within time $W(z,t)$ is given:

Fig. 3. Initial water content profile $W_0(z)$ performed by core sampling.

Fig. 4. Measured water content by humidity probe w_{HS} during one year.

$$W(z, t) = \Gamma(t) \cdot W_0(z) \quad (10)$$

Fig. 5 shows the maximum and minimum bounds of water content profiles during one year, which correspond to the profile of 1 June 2013 and that of 15 August 2013. It is worthwhile noting that underground water level increases from depth 4 m to 3.6 m during this period.

The profile of dry density $\gamma_d(z)$ is also taken as it is described in Fig. 6. Dry density of soil between the deepest in-situ measured data ($z = 0.87$ m) and the ground water level ($z = 4$ m) is considered to be constant and equals to $\gamma_d = 16 \text{ kN m}^{-3}$.

Knowing dry density and water content profile, the time-space degree of saturation of soil $S_r(z,t)$ can be easily determined by the following equation:

$$S_r(z, t) = W(z, t) \cdot G_s / \left(\left(\frac{G_s \cdot \gamma_w}{\gamma_d(z)} \right) - 1 \right) \quad (11)$$

where G_s is the soil specific gravity taken constant equal to 2.65 and γ_w is the unit weight of water taken constant equal to 10 kN/m^3 .

Fig. 5. Maximum and minimum bounds of water content profiles of multilayered soil corresponding to the profile of 1 June 2013 and that of 15 August 2013.

Fig. 6. Initial estimated dry density performed by core sampling.

3.3. Measurements of thermal properties of multilayered soil

For the layers of vegetarian and wind-blown sandy soils, a handheld device named KD2 Pro [35] is used to measure the thermal properties of the soil. The moist soil of each layer is then compacted into a column (50 mm inner diameter and 100 mm high) to reach a desired dry density. The compacted samples are sealed and placed in a constant temperature room ($20 \pm 1^\circ\text{C}$) for 24 h before measurements to ensure thermal equilibrium between water and soil. Soil thermal properties (λ and C) are determined with the algorithm for the dual needle probe based on the line heat source analysis given in Carslaw and Jaeger [36] and Kluitenberg et al. [37]. A (30 mm long, 1.3 mm diameter) dual needle SH-1 sensor is inserted carefully into each soil column, and current is applied to the sensors to produce a heat pulse. Heat-pulse measurements are repeated two times on each soil column.

Thermal properties of the soil backfill are measured with another method called stationary guarded-hot-plate method [38]. This method is more appropriate for the soil backfill as the size of the largest soil constituent is more than the distance between the dual needles of KD2 Pro device which is 5 mm. The plate size is $40\text{ cm} \times 60\text{ cm}$, the thickness of soil samples is less than 8 cm to ensure one dimensional thermal transfer. Thermal conductivity λ is then determined by the heat flow, the mean temperature difference between the sample surfaces and the dimensions of the samples. The same test apparatus was used to measure the soil thermal capacity C with the help of an insulation cover on the soil samples.

Thermal diffusivity α of soil is taken from two other thermal parameters, i.e., thermal conductivity λ and thermal capacity C of soil:

$$\alpha = \frac{\lambda}{C} \quad (12)$$

Figs. 7–9 show measured thermal properties (λ and α) of the three soil layers: vegetarian soil, wind-blown sandy soil, and soil backfill, respectively. The error bars of 10% in the experimental data are also reported in these figures. The thermal conductivity increases with respect to the degree of saturation while the thermal diffusivity has a maximum value around the degree of saturation of 50% for all the studied soils. Such a behavior was also observed by other researchers for the sandy soil [39].

Fig. 7. Comparison of the theoretical models and the measurements of vegetarian soil thermal properties: a) thermal conductivity λ , and b) thermal diffusivity α , performed with KD2 Pro method.

3.4. Prediction of thermal properties of multilayered soil

In this section, theoretical relationships for each parameter, thermal conductivity λ and thermal diffusivity, α are proposed.

Using the concept of normalized thermal conductivity [28], thermal conductivity is defined as:

$$\lambda = (\lambda_{\text{sat}} - \lambda_{\text{dry}})K_\lambda + \lambda_{\text{dry}} \quad (13)$$

where λ_{dry} and λ_{sat} are the thermal conductivity of dry and saturated soils ($\text{Wm}^{-1}\text{K}^{-1}$), respectively.

The measured thermal conductivity values from eight soils from China [40] are used to calibrate the proposed model. A set of linear relationships is proposed for the fitting functions adapted to those data [41,42]:

$$\begin{aligned} \lambda_{\text{sat}} &= ax_s + b\gamma_d \\ \lambda_{\text{dry}} &= cx_s + d\gamma_d \end{aligned} \quad (14)$$

where a and b are fitted parameters for saturated thermal conductivity equal to $0.53\text{ W m}^{-1}\text{K}^{-1}$ and $0.1\text{ W m}^2\text{K}^{-1}(\text{kN})^{-1}$, respectively, and where c and d are fitted parameters for unsaturated thermal conductivity equal to $0.087\text{ W m}^{-1}\text{K}^{-1}$ and $0.019\text{ W m}^2\text{K}^{-1}(\text{kN})^{-1}$, respectively (Least Square Regression Co-efficient R^2 higher than 0.8 for all fitted parameters).

To relate normalized thermal conductivity K_λ to the degree of saturation S_r , the equation proposed by Coté and Konrad (2005) [43] is used:

Fig. 8. Comparison of the theoretical models and the measurements of windblown sandy soil thermal properties: a) thermal conductivity λ , and b) thermal diffusivity α , performed with KD2 Pro method.

$$K_{\lambda} = \frac{\kappa \cdot Sr}{1 + (\kappa - 1) \cdot Sr} \quad (15)$$

where κ is a texture dependent parameter which varies linearly with sand content [41,42]:

$$\lambda_{sat} = ex_s + f \quad (16)$$

where e and f are constant parameters fitted with literature data [40] which are 4.4 and 0.4, respectively.

The following equation for soil thermal diffusivity is proposed with regard to the degree of saturation by considering the saturated thermal diffusivity (α_{sat}) and the dry thermal diffusivity (α_{dry}):

$$\alpha = (\alpha_{sat} - \alpha_{dry}) * (\beta * Sr - (1 - \beta) * Sr^2) + \alpha_{dry} \quad (17)$$

where β is a constant parameter taken equal to 4.

Results from Abu-Hamdeh (2003) and Ochsner et al. (2001) [29,39] show that there is no effect of initial dry density on the variation of thermal diffusivity. Therefore, a linear relationship is

Fig. 9. Comparison of the theoretical models and the measurements of soil backfill thermal properties: a) thermal conductivity λ , and b) thermal diffusivity α , performed with hot-plate method.

proposed for saturated and dry thermal diffusivities (α_{sat} and α_{dry}) with respect to sand content x_s :

$$\begin{aligned} \alpha_{sat} &= gx_s + h \\ \alpha_{dry} &= ix_s + j \end{aligned} \quad (18)$$

where g , h , i and j are constant parameters of values $4.5 \times 10^{-7} \text{ m}^2\text{s}^{-1}$, $3.5 \times 10^{-7} \text{ m}^2\text{s}^{-1}$, $2 \times 10^{-8} \text{ m}^2\text{s}^{-1}$ and $1.8 \times 10^{-7} \text{ m}^2\text{s}^{-1}$, respectively.

Applying the proposed theoretical approaches on each soil layer and using the related soil parameters in Table 1, the predicted thermal properties are compared with the laboratory measurements. Figs. 7–9 show respectively the predicted thermal properties (λ and α) of the three soil layers: vegetarian soil, wind-blown sandy soil, and soil backfill. It can be globally noted that the models agree well with the experimental measurements.

Based on Eqs. (13)–(18), α and λ in each time and depth needed for the modeling of heat transfer in a multi-layered soil are then related to the space-time variations of degree of saturations $S_r(z,t)$ by considering the sand content and the dry density of each layer.

3.5. Temperature measurements in the field of multilayered soil

As shown in Fig. 2, the monitored field is equipped with 5 temperature probes TS₂₀, TS₂₁, TS₂₂, TS₂₃, and TH₅₀. Temperature probe TH₅₀ is placed at the soil surface to measure the ambient temperature. The temperature probes TS₂₀ and TS₂₃ are placed at the same depth of 1.11 m. Temperature probes TS₂₁ and TS₂₂ are

Table 1

Sand content x_s and dry density γ_d of each layer at profile of multilayered soil.

Soil no.	Soil type	Sand content (%)	Dry density (kN m ⁻³)
1	Vegetarian soil	35	13.5
2	Soil backfill	80	16
3	Wind-blown sand	50	14.2

Fig. 10. Measured temperatures at the surface and different depths of soil during one year.

placed at the depths of 1.26 m and 0.61 m, respectively. Fig. 10 shows the temperature variations measured by the temperature probes at the surface of soil and throughout the profile during a year.

Fig. 11 shows the temperatures measured by different probes at $t = 0$ with depth ($T_0(z)$). A second degree polynomial presented in Fig. 11 is used as an initial condition $T_0(z)$ in the following simulations.

4. Numerical modeling

In this section, the results of the finite difference model are presented within time for the nonhomogeneous profile, then for the homogeneous profile.

4.1. Simulation of thermal response of multilayered soil

Using all the input parameters of the model, i.e., water content, dry density and sand content throughout the profile, the thermal conductivity as the thermal diffusivity were determined for each layer within time. Applying initial and boundary conditions as well as continuity conditions at interface of each layer, the finite difference modeling is performed to simulate thermal response of the profile of the multilayered soil.

In order to converge the finite difference calculations ($\text{Max}(\alpha(z,t)dt/dz^2) < 0.5$), for a Δt equal to 90 min (1 h30) based on the experimental acquisition, the minimum Δz should be selected higher than 0.07 m. It was taken 0.1 m for the studied geometry.

Fig. 12 shows the comparison of the model with the observation by thermal probes TS₂₀, TS₂₁, TS₂₂, and TS₂₃. Generally, good agreements are obtained between the simulation results and the

Fig. 11. Initial temperature variation at different depths at $t = 0$ ($T_0(z)$).

Fig. 12. Comparison of the model with the observation by thermal probes a) TS₂₀–TS₂₃, b) TS₂₁, and c) TS₂₂ during one year.

Fig. 13. Comparison of the model with the thermal probe measurements on four seasons throughout the profile.

experimental data. However, a small lag is observed between those simulation and experimental results. This is due to the continuum effects of heat transfer phenomenon by which the initial temperature of the problem is subjected to the effect of the temperature field history existed before the initial condition.

Fig. 13 shows the comparison of the model with the observations carried out on four seasons. It can be noted that the adiabatic condition (see Eq. (2)) at the depth of 4 m just above the ground water level is obtained. Generally, the results show that the model can estimate well thermal response of the multi-layered soil particularly for the last seasons with the average error less than 5% mainly due to the less time lagging for this period.

4.2. Comparison between nonhomogeneous and homogeneous soil cases

Until now, the modeling of the nonhomogeneous multilayered soil with different characteristics in each layer was studied. It is essential to compare the nonhomogeneous soil with the homogeneous one to evaluate the effects of non-homogeneity on the soil thermal response.

Here, the same soil profile shown in Fig. 2 is considered as a homogeneous soil with similar thermal characteristics throughout the profile. Three cases are considered: (I) thermal diffusivity of the soil profile is taken as the maximum value of the variable thermal diffusivity in depth during the period of one year, i.e., $\alpha_{\max} = \max(\alpha_{p,k})$, (II) thermal diffusivity of the soil profile is taken as the minimum value of the variable thermal diffusivity in depth during the period of one year, i.e., $\alpha_{\min} = \min(\alpha_{p,k})$, and (III) thermal diffusivity of the soil is taken as the average value, i.e., $\alpha_{\text{average}} = (\sum_{p=1}^n \sum_{k=1}^m \alpha_{p,k}) / N$ where $N = p \times k$ is the total number of nodes p and k .

Fig. 14a shows the distribution of degree of saturation throughout the profile of nonhomogeneous soil in its minimum value ($S_{r_{\min}}$) in 15 August 23 and maximum value in 1 June 2013 ($S_{r_{\max}}$) associated to the previously presented water content profile in Fig. 5 based on Eq. (11). The distribution of thermal diffusivity associated for $S_{r_{\min}}$ and $S_{r_{\max}}$ is shown in Fig. 14b. Thermal diffusivity of nonhomogeneous soil at $S_{r_{\min}}$ and $S_{r_{\max}}$ varies between the so-called values of $\alpha_{\min}$ and $\alpha_{\max}$. Thermal diffusivity of soil profile at $S_{r_{\max}}$ is generally smaller than that of $S_{r_{\min}}$. Such phenomenon is due to the fact that the degree of saturation of the

Fig. 14. a) Variation of degree of saturation at minimum and maximum values of S_r . b) Variation of thermal diffusivity throughout the nonhomogeneous soil profile at maximum and minimum values of S_r .

profile varies between 0.5 and 1.0. In this range of degree of saturation, the thermal diffusivity decreases with respect to the degree of saturation (after $S_r = 0.5$) for the three soil types (see Figs. 7b, 8b and 9b). For the degree of saturation less than 0.5, thermal diffusivity of soil increases with degree of saturation while there are small ranges of thermal diffusivity when thermal diffusivity at $S_{r_{\max}}$ is larger than that of $S_{r_{\min}}$.

Figs. 15 and 16 show temperature distribution throughout the homogeneous and nonhomogeneous soil profile on 1 December 2012 when the temperature is close to 0° (temperature below zero was not considered in this study because of the complexity of heat transfer phenomena in the frozen soils) and 15 June 2013 when there is a maximum value of temperature. The experimental observations were also reported in these figures. Figs. 15a and 16a show that temperature distribution of the nonhomogeneous soil profile with variable thermal diffusivity varies between the temperature distributions of homogeneous soil at minimum and maximum values of thermal diffusivity. It shows the effect of non-homogeneity on thermal response of soil. Figs. 15b and 16b show that the temperature distribution of the nonhomogeneous soil with

Fig. 15. Temperature distribution of nonhomogeneous soil (model and observation) compared with a) soil maximum and minimum thermal diffusivity and b) soil average thermal diffusivity, on 15 June 2013.

Fig. 16. Temperature distribution of nonhomogeneous soil (model and observation) compared with a) soil maximum and minimum thermal diffusivity and b) soil average thermal diffusivity, on 1 December 2012.

variable thermal diffusivity becomes almost the same value as the temperature distribution of the homogeneous soil with the average value of thermal diffusivity. But, it should be mentioned that knowing the average value of thermal diffusivity of the homogeneous soil requires thermal diffusivity values of each layer in the entire profile.

5. Conclusion

This study deals with a finite difference modeling of heat transfer in unsaturated multilayered soil while hydrothermal properties of soil differ in each layer.

A profile of thermal diffusivity within time is assigned to each node in finite difference formulation. The numerical model requires easily accessible input parameters such as water content, dry density, and sand content of the soil profile. It is coupled with the reliable theoretical approaches which estimate thermal properties of soil such as thermal conductivity and thermal diffusivity.

A comparison of the numerical results with the results of in-situ thermal probe measurements shows that the model can well estimate heat distribution within a multilayered soil.

Moreover, a study is performed on the effect of non-homogeneity on soil thermal response. The soil with an average value of thermal diffusivity produces a thermal response close to nonhomogeneous profile. However, it is worthwhile noting that knowing the average value of thermal diffusivity requires thermal characteristics of each layer in the nonhomogeneous profile.

Such a study on the thermal response of multilayered soil can be considered on the study of the performance of geothermal borehole heat exchangers installed in soils with variable hydrothermal properties.

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