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A moving-pump model for water migration in unsaturated freezing soil

Jiazuo Zhou [a], Changfu Wei [a,*], Dongqing Li [b], Houzhen Wei [a]

[a] State Key Laboratory of Geomechanics and Geotechnical Engineering, Institute of Rock and Soil Mechanics, Chinese Academy of Sciences, Wuhan 430071, China
[b] State Key Laboratory of Frozen Soil Engineering, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou 730000, China

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ABSTRACT

This paper presents a new approach for simulating the water migration in freezing soils, in which the pore water migration and heat transfer are characterized using an imaginary pump attached with a small imaginary reservoir. The pump moves with the freezing front as temperature decreases, sucks the liquid water from the unfrozen zone and then stores it in the frozen zone. The reservoir is used to gather the sucked water and store it in the form of pore ice through phase change. Explicit governing equations are developed for describing the water migration, crystallization and/or heat transfer in the soil, the pump and the reservoir. The proposed model is numerically implemented into a commercial code. Compared to the previous approaches used to simulate the soil freezing processes, application of the new approach avoids remeshing and recalculating the moving boundaries, and this feature can drastically simplify the numerical implementation of the theoretical model. The new approach is used to analyze the one-dimensional freezing process in soils. The simulated results are compared with the experimental data available in the literature and the simulations based on other approaches, showing that the new approach is capable of effectively simulating the freezing process of soils.

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1. Introduction

Any freezing process in soil is accompanied by both heat and pore water transfers, and these processes occur in a coupled manner. During a freezing process, a temperature gradient forms in the soil, driving the heat to flow from the higher-temperature zone toward the lower-temperature zone and the pore water to migrate from the unfrozen zone to the frozen zone. The pore water migration driven by temperature can influence the heat conduction process due to the effect of convection and latent heat of phase change, while the heat conduction may induce phase change and in turn change the hydraulic conductivity of the soil (Harlan, 1973; O'Neill and Miller, 1985; Taylor and Luthin, 1978). In addition, both heat and mass transfer can change the physical and mechanical properties of the soil. In analyzing the problems related to soil freezing, it is crucial to properly characterize both heat and pore water transferring processes.

If a fully saturated soil with a sufficient water supply begins to freeze, the soil water will constantly migrate to the frozen zone from the unfrozen region due to the effect of cryosuction, resulting in an increase in the water content of the frozen zone, while the water content in the unfrozen zone remains practically unchanged. Therefore, in analyzing the freezing process of a fully saturated soil with a sufficient water supply, only the water increase in the frozen zone and consequently the total frost heave are of concern, and the problem can be solved by ignoring the effect of the water content variation and the skeletal deformation in the unfrozen zone (Xu and Deng, 1991; Zhou et al., 2011). In the freezing process of an unsaturated soil, however, the water migration is more complex, and this is the case especially for a closed system, i.e., the soil without a water supply. In this case, the water content increases in the frozen zone while decreases in the unfrozen zone. The problem is complicated by the movement of the interface between the frozen and the unfrozen zone with temperature.

To simulate the processes of heat and water transfer in unsaturated freezing soil, Chen et al. (1990) and Hu et al. (1992) developed the governing equations of water migration in the frozen and unfrozen zones, respectively. It is remarkable, however, that in applying these equations, a boundary condition has to be introduced to ensure the flow continuity between the frozen and unfrozen zones. As such, when the interface moves with the freezing front, the boundary condition of these two equations also vary with temperature. Hence, if a finite element or finite difference procedure is adopted in the simulation, it is necessary to remesh and recalculate the moving boundary after each time step. In addition to this complexity, the high nonlinearity of the governing equations and the coupling of heat and water transfer make the simulation procedure rather difficult (Black, 1995a; Chen et al., 1990; Hu et al., 1992; O'Neill and Miller, 1985; Taylor and Luthin, 1978; Zhou and Zhou, 2010).

In this paper a new approach is presented to simulating the one-dimensional pore water migration in freezing unsaturated soils. In this approach, the frozen fringe can be envisioned as a moving pump,

* Corresponding author. Tel.: +86 27 87197239; fax: +86 27 8719 8346.
E-mail address: cfwei@whrsm.ac.cn (C. Wei).

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which sucks pore water from the unfrozen zone and store it as pore ice through phase transition in the frozen zone. To characterize the crystallization of the sucked pore water, a small imaginary reservoir is introduced, which is attached to the moving pump and used for collecting the sucked pore water and related phase transition. During the soil freezing, the pump and the reservoir simultaneously move as the temperature decreases. Governing equations are then developed for the water migration and heat transfer as well as phase transition in the pump and/or in the reservoir. As such, in simulating the soil freezing process, it is not necessary to change the boundary conditions of the solution domain, and the theoretical model can be readily implemented into a commercial code without complex programming.

2. Theory

2.1. The driving force for water migration in freezing soil

In a frozen soil, a certain amount of unfrozen water exists in the vicinity between the surfaces of soil grains and ice grains due to the premelting effect (Wettlaufer and Worster, 1995; Wettlaufer et al., 1996; Xu et al., 1993). At equilibrium, the pressures of the unfrozen pore water and the pore ice can be related to each other through the generalized Clapeyron equation (Black, 1995b), i.e.,

$$\mathrm{d}u_w = \frac{\rho_w}{\rho_i}\mathrm{d}u_i + \frac{\Delta H \rho_w}{T_0}\mathrm{d}T \tag{1}$$

where u_w and u_i are the pressures of unfrozen pore water and pore ice, respectively; ρ_w and ρ_i are the densities of pore water and pore ice, respectively; T_0 (K), T(°C) and ΔH are the freezing point of bulk water, the current temperature and the latent heat of fusion, respectively. Without overburden loading, u_i remains practically unchanged, while u_w decreases linearly with temperature according to Eq. (1).

Strictly, the generalized Clapeyron equation is valid only in the case that unfrozen pore water and pore ice coexist in equilibrium. In any transient process, however, this equilibrium condition cannot be strictly achieved. In the following, it is assumed that the temperature change and the water migration are slow enough compared to the phase change, so that Eq. (1) is valid under the transient condition. Eq. (1) implies that, if ice pressure gradient is neglected, a temperature gradient can induce pore water pressure gradient, driving the pore water to migrate from the higher temperature zone to the lower temperature zone.

Harlan (1973) assumed that the potential of pore water in the frozen soil equals to that in the unsaturated soil with the same liquid water content. This assumption has been validated using the soil–water characteristic curve (SWCC) and the soil freezing characteristic curve (SFCC) of the soil under partially saturated and frozen conditions, respectively (Azmatch et al., 2012; Liu et al., 2011; Spaans and Baker, 1996). Indeed, according to the generalized Clapeyron equation, one can easily see that, if the ice pressure is constant and capillary hysteresis is excluded, there is a one-to-one correspondence between unfrozen water content and temperature in the frozen soil. Hence, both the pore water pressure and the temperature in the frozen soil (with undercooling pore water) can be expressed as a function of unfrozen water content only. Based on the above discussions, it is suggested that, if the ice pressure remains constant and its gradient is neglected, the driving force of water migration in the frozen soil can be expressed as the gradient of unfrozen water content.

2.2. Governing for water migration and heat transfer

Based on the above discussions, if the gradient of ice pressure is negligible, the seepage velocity can generally be expressed as the diffusivity multiplying the gradient of unfrozen water content (Shao et al., 2006). Recalling the assumption that the heat conduction and the water migration are slow enough compared to the phase change (between liquid water and ice), one obtains the governing equation for water migration in frozen soils, which in a form similar to the Richards Equation (Richards, 1931; Taylor and Luthin, 1978):

$$\frac{\partial}{\partial t}\left(\theta_u + \frac{\rho_i}{\rho_w}\theta_i\right) = \frac{\partial}{\partial x}\left(D\frac{\partial \theta_u}{\partial x}\right) \tag{2}$$

where t and x represent the elapsing time and the spatial coordinate, respectively; D is the water diffusivity; θ_u is the specific unfrozen (or liquid) water content; θ_i is the specific ice content. The term in the bracket of the left-hand side is equal to the total specific water content (including both liquid water and ice). Hereinafter both ρ_w and ρ_i are assumed to be constant. Eq. (2) implies that any change in the total specific water content of the frozen soil is solely due to the transfer of unfrozen water.

In general, the migrating process of pore water in a frozen soil is slow, and thus its effect on heat convection is negligible. Hence, one obtains the heat conduction equation as (Taylor and Luthin, 1978):

$$C\frac{\partial T}{\partial t} = \frac{\partial}{\partial x}\left(\lambda\frac{\partial T}{\partial x}\right) + L\rho_i\frac{\partial \theta_i}{\partial t} \tag{3}$$

where C is the volumetric heat capacity, λ is the thermal conductivity of the soil, and L is the latent heat of fusion (per unit mass of water).

In a freezing soil subjected to a temperature gradient, two processes associated with crystallization in the pores may simultaneously occur. Indeed, as the temperature decreases, part of the liquid water at the site may change into ice, while a certain amount of liquid water is sucked from the higher-temperature zone into the frozen zone where it crystallizes. Correspondently, the variation of θ_i (Eq. (3)) can be additively decomposed into two components (Penner and Ueda, 1978; Zhou et al., 2011): one is due to the crystallization of the liquid pore water at the site (denoted as $\mathrm{d}\theta_1$), and the other is due to the crystallization of the sucked liquid water (i.e., the water transferred from other places), which is denoted as $\mathrm{d}\theta_2$. Then, the variation of the specific ice content can be expressed as

$$\mathrm{d}\theta_i = \mathrm{d}\theta_1 + \mathrm{d}\theta_2 \tag{4}$$

where

$$\mathrm{d}\theta_1 = -\frac{\rho_w}{\rho_i}\mathrm{d}\theta_{in} \tag{5}$$

and

$$\frac{\partial \theta_2}{\partial t} = \frac{\rho_w}{\rho_i}q \tag{6}$$

where θ_{in} is the specific content of the liquid water that changes into pore ice at the site, and q is the changing rate of the specific content of the liquid water sucked from the higher-temperature zone. Eq. (6) implies that all the liquid water sucked from the higher-temperature zone changes into pore ice.

Substituting Eqs. (4)–(6) into Eq. (3), one obtains

$$C_e\frac{\partial T}{\partial t} = \frac{\partial}{\partial x}\left(\lambda\frac{\partial T}{\partial x}\right) + L\rho_w q \tag{7}$$

where C_e is the equivalent volumetric heat capacity, defined by

$$C_e = C + L\rho_w\frac{\partial \theta_{in}}{\partial T} \tag{8}$$

Clearly, C_e includes a component related to the latent heat that is released by the phase change at the site (Bonacina et al., 1973).

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