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Research Paper

A consolidation model for estimating the settlement of warm permafrost

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ABSTRACT

On the basis of the Biot's consolidation theory, a thermo-hydro-mechanical (THM) model was presented for computing the settlement of warm frozen soils under an external load. The proposed model allowed the unfrozen water in the soils to migrate according to the effective stress principle. In addition, the pressure potential due to loading was considered as the only driving force for the migration of the unfrozen water. The model is validated against an *in situ* long-term plate loading test that was conducted in a warm permafrost region on the Qinghai–Tibetan Plateau. The computed ground temperature and settlement were consistent with the measured data. Using the model, we analyze the variations of the pore-water pressures at different depths and discuss the influences of the seepage on the deformation, ground temperature and effective stress.

1. Introduction

In saturated frozen soils, most of the unfrozen water becomes ice. When subjected to an external load, the soils perform rheological deformation, which may be detrimental to the stability of infrastructures on permafrost and has been merely considered as creep by most researchers [1–3]. Several calculation methods have been developed to describe creep processes in frozen soils [2,4–12]. However, a considerable amount of unfrozen water still exists in the soil matrix, particularly at the subzero temperatures near 0 °C [5,12,13]. Moreover, the hydraulic conductivity of warm frozen soils is not equal to zero [14,15]. Thus, In addition to the movement of soil particles and the flow of ice, the migration of the unfrozen water around the solid particles could be an important factor that results in the displacement of frozen soil.

Frozen soils may follow the effective stress principle to produce a consolidation deformation [16]. At the temperatures near the freezing point, Sayles [17] indicated that the consolidation of frozen soils could account for more than 1/3 of the total soil deformation. Therefore, a consolidation model considering the displacement of warm and ice-rich frozen soils subjected to external loads and varying temperature is critical to estimate the settlement of warm and rich-rich permafrost. But unfortunately, few studies have considered consolidation in this situation.

Although the ground temperature varies seasonally, the displacement of the frozen soils beneath permafrost table under a load is mainly characterized by settlement rather than frost heave [11]. Thus, the pressure potential that results from loading can be regarded as the only force driving the migration of unfrozen water when analyzing the deformation mechanisms of frozen soils under an external load.

In this paper, we establish a mathematical model based on Biot's consolidation theory to compute the settlement of warm frozen soils when subjected to loadings. Then, we apply the proposed model to simulate a long-term plate loading experiment in a warm permafrost region on the Qinghai Tibetan Plateau using the Comsol Multiphysics software. The simulation results are validated against the experimental data. The variations of the simulated pore-water pressure in the permafrost are analyzed, and the influences of seepage on deformation, ground temperature and effective stress are discussed.

2. Mathematical model

Based on Biot's consolidation theory, the following assumptions have been adopted to develop a coupled thermo-hydro-mechanical model for warm frozen soils. (1) Soil is homogeneous, continuous, isotropic, and saturated. (2) Pressure thawing of ice is ignored. (3) Deformation of permafrost obeys the principle of effective stress. (4) The migration of water due to a temperature gradient is not considered.

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Nomenclature

C	heat capacity
C_f	heat capacity of the frozen soil
C_u	heat capacity of the thawed soil
C_w	heat capacity of water, 4.18 kJ/(kg K)
λ	thermal conductivity
λ_f	thermal conductivity of the frozen soil
λ_u	thermal conductivity of the thawed soil
T	temperature
T_b	lower limit temperature of the phase change
T_p	freezing point
t	time
ρ	soil density
ρ_f	water density, 998 kg/m^3
$\vec{V}$	seepage velocity of the pore-water
L	latent heat of the ice-water phase change, 334.56 kJ/kg
W_0	total water content
W_i	ice content
W_u	unfrozen water content
γ	test parameter
p_f	pore-water pressure
k	permeability
K_p	hydraulic conductivity of the frozen soil
K_{-1}	hydraulic conductivity of the frozen soil at −1 °C
K_0	hydraulic conductivity of the thawed soil
β	test parameter
χ_f	compressive coefficient, 5×10^{-10} Pa^{-1}
μ	viscosity coefficient of water, 1.781×10^{-3} Pa s
g	gravitational acceleration, 9.8 N/kg
ε_V	volume strain of soil
S	storage coefficient
n_0	initial porosity
n	porosity
α_B	Biot's coefficient
σ	total stress
σ'	effective stress
ε	total strain
E_1	elastic moduli for the Maxwell body
E_2	elastic moduli for the Kelvin body
η_1	viscosity coefficients for the Maxwell body
η_2	viscosity coefficients for the Kelvin body
v	Poisson ratio
T_0	mean annual temperature of the ground surface
α	warming rate of the air temperature
A	fluctuation amplitude of the surface temperature
$n\pi$	initial phase of the surface temperature variation
a_i	test constant, $i = 1, 2$
b_i	test constant, $i = 1, 2$
∇D	relative height

2.1. Heat transfer equation

Based on the energy conservation theory, the heat transfer equation considering the heat conduction and heat convection can be expressed as follows [18]:

$$\rho C \frac{\partial T}{\partial t} = div(\lambda grad(T)) + C_w \rho_f div(\vec{V} \cdot T) \tag{1}$$

In this study, it is assumed that the phase change of ice to water occurs at the temperatures of $T_b \leqslant T \leqslant T_p$. When the phase change occurs, the latent heat is released or absorbed, which influences the heat transfer enormously. This influence can be considered using the sensible heat capacity method. The heat capacity and thermal conductivity vary with ground temperature in the interval of the phase change or are held constant.

$$C = \begin{cases} C_u & T > T_p \\ C_f + \frac{C_u - C_f}{T_p - T_b}(T - T_b) + \frac{L}{1+W_0}\frac{\partial W_i}{\partial T} & T_b \leqslant T \leqslant T_p \\ C_f & T < T_b \end{cases} \tag{2}$$

$$\lambda = \begin{cases} \lambda_u & T > T_p \\ \lambda_f + \frac{\lambda_u - \lambda_f}{T_p - T_b}(T - T_b) & T_b \leqslant T \leqslant T_p \\ \lambda_f & T < T_b \end{cases} \tag{3}$$

At temperatures below the freezing point, the quantity of the unfrozen water increases as the temperature increases. Once the soil thaws, the quantity of the unfrozen water becomes a constant [19].

$$W_u = \begin{cases} W_0 (T_p/T)^{\gamma} & T \leqslant T_p \\ W_0 & T > T_p \end{cases} \tag{4}$$

Because of the mutual transformation between unfrozen water and ice, their rates of change are equal $\left(\frac{\partial W_i}{\partial T} = \frac{\partial W_u}{\partial T}\right)$.

2.2. Water migration equation

Darcy's law requires a linear correlation between hydraulic gradient and discharge. Some researchers have found that this correlation can also be met in frozen soils [14,15]. Saturated frozen soils contain three components: soil particles, ice and unfrozen water. When subjected to an external load, the internal stress can be attributed to the following three pressures: soil skeleton, ice and pore-water pressures. To simplify the internal stress calculations, the ice can be regarded as a part of the solid matrix. Thus, the internal stress in frozen soils consists of the matrix stress and pore-water pressure.

By improving Biot's consolidation equations according to Van der Knaap [19] and Verruijt [20], the seepage equation can be written as follows:

$$\rho_f S \frac{\partial p_f}{\partial t} + \nabla \cdot \rho_f \left[-\frac{k}{\mu} (\nabla p_f + g \nabla D) \right] = -\rho_f \alpha_B \frac{\partial \varepsilon_V}{\partial t} \tag{5}$$

Biot's coefficient, α_B, is equal to 1 [21]. When the soil is saturated, the storage coefficient S (MPa^{-1}) is the product of the porosity n and the compressive coefficient of the fluid χ_f, $S = n\chi_f$. When the soils are compressed, n is reduced and can be obtained as follows:

$$n = \frac{(n_0 - \varepsilon_V)}{(1 - \varepsilon_V)} \tag{6}$$

Seepage in frozen soils is closely related to soil temperature, which can be described as follows [15,22]:

$$K_p = \begin{cases} \frac{K_{-1}}{(-T+T_p)^{\beta}} & T < T_p \\ \frac{K_0}{1+\varepsilon_V}\left(1 + \frac{\varepsilon_V}{n_0}\right)^3 & T \geqslant T_p \end{cases} \tag{7}$$

When the soil temperature is higher than the freezing point, the hydraulic conductivity changes with the volumetric strain. Below the freezing point, the hydraulic conductivity is considered as a function of only soil temperature because the volumetric strain is negligibly small.