Pergamon

Int. J. Heat Mass Transfer. Vol. 38, No. 12, pp. 2297–2303, 1995

0017-9310/95 $9.50+0.00

0017-9310(94)00283-5

Evaluation of heat and moisture transfer properties in a frozen–unfrozen water–soil system

ASHOK K. SINGH† and D. R. CHAUDHARY

Thermal Physics Laboratory, Department of Physics, University of Rajasthan, Jaipur 302 004, India

(*Received in final form* 10 *August* 1994)

Abstract—An experimental study of heat and mass transfer in a moist unsaturated sand has been carried out by maintaining the two ends of a sample column at temperatures of −5 and 30°C, respectively. Measurements of temperature have been made as a function of distance from the cold end and time, along with effective thermal conductivity and liquid water distribution at the end of the experiment. Comparison of the measured thermal conductivity values shows close agreement with calculated values from an effective continuous model (Singh *et al.*). Evaluation of thermal and moisture diffusion coefficients has been done from measured steady-state temperature and moisture profiles. A dryout condition has also been determined from the evaluation of heat and moisture transfer properties.

1. INTRODUCTION

Heat and mass transfer in soils is a complicated problem because of the presence of the solid, liquid and gaseous phases. During the freezing process the problem is also complicated because the freezing of soil generally produces significant changes in its structural, mechanical and thermal properties. The thermal status of freezing soils which are in a delicate balance and disturbance of this region could have serious engineering and ecological implications. Knowledge of thermal behaviour is essential in ensuring that such a disturbance does not occur, and also for the energy balance at the soil surface. The process of simultaneous heat and mass transfer in the freezing process in soils is of practical interest in many other engineering applications. The heat and mass transfer properties of soil systems are also essential for evaluating the effectiveness of insulation of energy storage devices, buildings, and storage of solar and geothermal energy in soils. These properties are also useful for drying processes, heat transfer calculations from buried pipelines and underground space utilization. They include heat dissipation from underground electrical power cables, heat transfer to and from buried heat pump coils, recovery of geothermal energy and waste heat rejection from power plants.

For schemes of a medium to large scale, buried pipelines and ground-based reservoirs are appropriate means of storage and transportation. Unfortunately, these devices usually need to be surrounded by bulky and relatively cheap thermal insulation. This suggests that sand/soil can act as a natural and robust thermal insulation. In determining the long-term thermal insulation properties of sand/soil, the different properties, i.e. diffusion coefficients, thermal properties with distribution of moisture under the effect of a temperature gradient and dryout possibilities under different conditions, have to be known.

Thermal gradients in a moist soil can induce moisture movement which, in turn, can result in a decrease in the thermal conductivity of the unfrozen soil and an increase in the thermal conductivity of the freezing side. In particular, the layered structure is associated with considerable moisture migration to the freezing front, as may occur in a frost-susceptible soil. They are considered to be coupled processes interacting together. In consequence, changes occur in the properties of both the freezing layer and underlying unfreezing zone.

The long range objective of this study is to determine experimentally the effect of moisture migration on freezing point depression and related phenomena on the cooling of soil with sufficient precision, and to provide the basis for the development of mathematical models which take these complications into proper account. Heat and mass transport properties of moist porous media are basically important in solving such problems. However, there are few data and still fewer convenient means available for the measurement of these properties at the moment. An evaluation of these properties with the distribution of moisture content and its effect on the effective thermal conductivity (ETC) and the diffusion coefficients has been made in the present investigation, along with the conditions for the existence of a dryout region under experimental conditions.

† Present Address: Headquarters, Snow and Avalanche Study Establishment (SASE), Manali, H. P. 175 131, India.

NOMENCLATURE

a	constant
b	slope of density temperature curve
D	diffusion coefficient
h	latent heat of evaporation
J	moisture flux
K	moisture diffusion coefficient
Le	Lewis number
M	moisture content by weight at saturation
m	moisture content by weight at unsaturation
q	power per unit length supplied to the probe heater
T	temperature [°C]
t	time
W	moisture content
x	coordinate (distance).

Greek symbols

α	thermal diffusivity
β	new thermal diffusivity
λ	thermal conductivity
ρ	density
ε	porosity
ν	mass condensation/evaporation per unit volume
ω	mass content of water vapour in saturated air at 760 mm of Hg
ψ	volume fraction
ψ_T	non-dimensional steady-state temperature
ψ_m	non-dimensional steady-state moisture
ζ	phase dispersion
ϕ	ratio of energy stored in vapour enthalpy to that stored as internal energy.

Subscripts

a	air
c	cold
d	dry
e	effective
ECM	effective continuous media
g	gas
H	hot
i	initial
j	1 for frozen and 2 for unfrozen
ma	moist air
S	saturation
s	solid
s.s.	steady-state
v	vapour
w	moisture (water)
θ	thermal
θv	thermal vapour
θw	thermal moisture.

2. MEASUREMENT AND ANALYSIS

Temperature and water content profiles in a vertical soil column, with the cold side temperature below the freezing point of water, and subjected to temperature gradient, were measured as a function of time and distance from the cold end. The moisture content profile was measured at the steady state. The soil sample was dune sand of particle size 150–177 μm, porosity 0.42 in dry state and density 1620 kg m^{-3}. The sample had a predetermined moisture content of 14.47% and was in a PVC pipe of length 27.0 cm and diameter 15.0 cm. The moisture content was measured by a gravimetric method using a precise electric balance: accurate to 0.1 mg. This oven drying technique is the most widely used method for measuring soil moisture content and is the standard for the calibration of all other soil moisture determination methods [1]. The error associated with this method can be up to 10% [2] and consists of drying a soil sample in an oven at 105°C until a constant weight is obtained.

The experimental arrangement is similar to that illustrated in Fig. 1. Controlled liquids of constant temperatures, i.e. -5°C and 30°C, were circulated through hollow copper circulation plates placed at both the ends of the soil column. Thirteen thermal

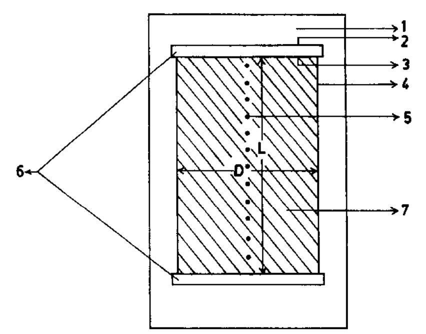

Fig. 1. Schematic diagram of experimental arrangement.

conductivity probes were placed at intervals of 2.0 cm throughout the length of the column. Temperature recording was achieved throughout the length of column by an inbuilt copper–constantan thermocouple within each probe and a Therm 5502 (microprocessor-based data recording system, German made), which has a facility of recording temperatures at 20 points within an accuracy of $\pm$0.02°C.

The effective thermal conductivity at each probe position i.e. with the distribution of moisture and temperature profile was measured using the expression [3]:

$$\lambda_e = \frac{q}{4\pi(T_2 - T_1)} \ln\left(\frac{t_2}{t_1}\right) \quad (1)$$

where T_2 and T_1 are two temperatures at times t_2 and t_1, respectively. By knowing the value of the slope between $\ln(t)$, the temperature rise, and q, power per unit length supplied to the probe heater, one can determine the thermal conductivity λ_e. At the end of the experiment, the probes were removed and a small amount of the sample was extracted from the vicinity of the probe position with the help of hollow brass tubes. The determination of the water content was achieved by weighing a sample specimen with a precise electrical balance before and after drying in an oven at 105°C. The calculation of the ETC was accomplished using temperature-dependent expressions [4].

The moisture concentration (W) and temperature (T) in an unsaturated porous medium has been described as [5, 6]:

$$\frac{\partial W}{\partial t} = D_\theta \nabla^2 T + K\nabla^2 W \quad (2)$$

and

$$\frac{\partial T_j}{\partial t} = \beta_j \nabla^2 T_j \quad (3)$$

where

$$\beta_j = \alpha_j \frac{(1+(\varepsilon_g b_j h_j/(\rho c)_j)(D_j/\alpha_j))}{(1+\varepsilon_g b_j h_j/(\rho c)_j)}.$$

The term β_j is a rather new diffusivity term which is dependent on the thermal diffusivity of the material (α_j), the latent heat of evaporation (h_j), the slope of density temperature curve (b_j), the volumetric heat capacity of the material $(\rho c)_j$, and it reveals the character of the system. A brief account of the development of equation (3) is given below.

At any point within the system, the rate of change in vapour density ($\varepsilon_g \rho_v$) equals the diffusion rate to that point minus/plus the condensation/evaporation (v), i.e.

$$\frac{\partial(\varepsilon_g \rho_v)}{\partial t} = D_j \frac{\partial^2(\varepsilon_g \rho_v)}{\partial x^2} - v_j. \quad (3a)$$

Here subscript j is 1 for the frozen region and j is 2 for the unfrozen region. The change in internal energy at any point is expressed as

$$\frac{\partial(\rho c)_j T_j}{\partial t} = \lambda_{ej} \frac{\partial^2 T_j}{\partial x^2} + h_j v_j. \quad (3b)$$

The empirical relation for vapour density with temperature can be written as [7]

$$\rho_{vj} = a_j + b_j T_j. \quad (3c)$$

Using equation (3a), (3b) with (3c) one can obtain equation (3).

For one-dimensional flow of heat the moisture concentration flux from equation (2) is

$$J = -D_\theta \frac{\partial T}{\partial x} - K\frac{\partial W}{\partial x} \quad (4)$$

where $D_\theta = D_{\theta w} + D_{\theta v}$ and $K = K_w + K_v$ are thermal and moisture diffusion coefficients, respectively. Under the condition of steady-state flow, the net flux of moisture must be zero, i.e.

$$\frac{D_\theta}{K} = -\frac{(\mathrm{d}W/\mathrm{d}x)}{(\mathrm{d}T/\mathrm{d}x)}. \quad (5)$$

From equation (5) the evaluation of the ratio (D_θ/K) can be obtained under the condition of one-dimensional heat flow from the steady-state measurements of temperature and moisture profiles. The thermal diffusion coefficient from the definition itself can be expressed as

$$D_\theta = \frac{\rho_a}{\rho_d} D_s \frac{\partial \omega}{\partial t}. \quad (6)$$

The air density (ρ_a) can be expressed as

$$\rho_a = 1.2929\left(\frac{273.13}{T+273.13}\right).$$

The vapour diffusivity through the porous material (D_s), is equal to approximately one-fifth of the diffusion coefficient of water vapour through air, which is expressed as

$$D_j = 2.29 \times 10^{-5}\left[\frac{T+273.13}{273.13}\right]^{1.75} \quad [\mathrm{m^2\,s^{-1}}].$$

The mass content of water vapour (ω) in saturated humid air under 760 mm of Hg with variation of temperature was derived by Yu [8] and is expressed as

$$\frac{\mathrm{d}\omega}{\mathrm{d}t} = \frac{1.8041 \times 10^6 \exp(g)}{[760 - 0.378 \times \exp(g)]^2 (T+227.02)^2} \quad [\mathrm{K^{-1}}]. \quad (7)$$

The parameter g is

$$g(T) = 18.304 - \frac{3816.4}{T+227.02}.$$

Here one obtains the concept of dryout in a moist porous medium when the moisture profile reaches its steady state [9] by writing equations (2) and (3) in dimensionless form and using Green's theorem under steady-state conditions. The non-dimensional steady-state temperature (ψ_T) and moisture (ψ_M) profiles are related by $\psi_M = -\psi_T + \langle\psi_T\rangle$, irrespective of the thermal boundary conditions with impermeable boundaries. One then has

$$(\psi_M)_{s.s.} = \frac{K}{D_\theta \nabla T}((W)_{s.s.} - W_i) = -(\psi_T)_{s.s.} + \langle(\psi_T)_{s.s.}\rangle$$

$$(W)_{s.s.} = W_i - \frac{D_\theta \nabla T}{K}((\psi_T)_{s.s.} - \langle(\psi_T)_{s.s.}\rangle). \quad (7a)$$

From this, one defines the wet region ($W > 0$) as

$$(\psi_T)_{s.s.} < \frac{KW_i}{D_\theta \nabla T} + \langle (\psi_T)_{s.s.} \rangle. \quad (7b)$$

Using the Laplace transform technique, the solution at steady state is

$$(\psi_T)_{s.s.} = 1 - X \quad \text{with} \quad \langle (\psi_T)_{s.s.} \rangle = 1/2 \quad (8)$$

$$(\psi_M)_{s.s.} = -(1-X) + \langle (1-X) \rangle$$

$$= -(1-X) + 1/2 = X - 1/2. \quad (9)$$

In these expressions $X = 2x$, then equation (7a) shall be

$$W = W_i + (2x - 1/2) D_\theta \nabla T / K \quad (10)$$

and the wet region is given by equation (7b)

$$\frac{1}{4} = \frac{KW_i}{2D_\theta \nabla T} < x. \quad (11)$$

When the left-hand side of equation (11) is set equal to zero, one obtains the upper bound of W_i for which dryout is possible.

$$(W_i)_{dryout} = \frac{D_\theta \nabla T}{2K}. \quad (12)$$

The effective thermal conductivity (ETC) has been calculated using the expressions developed by Singh *et al.* [4]. This model has been tested for different types of porous materials saturated with various liquids and for materials having a wide variation in porosity. The ETC of moist soil is expressed as

$$\lambda_e = \lambda_{ECM} \left[1 + 3.844 \left(\frac{\lambda_s - \lambda_{ECM}}{\lambda_s + 2\lambda_{ECM}} \right) \zeta_s^{2/3} \right]. \quad (13)$$

Here $\zeta_s = \psi_s - 0.5$, where ψ_s, ζ_s, λ_s and λ_{ECM} denote the volume fraction of the solid phase, the solid phase dispersion, the thermal conductivity of solid phase and the thermal conductivity of the effective continuous medium, which is expressed for a moist porous medium as

$$\lambda_{ECM} = 1.092(\lambda_{ma}\lambda_s)^{1/2}. \quad (14)$$

The thermal conductivity of moist air (λ_{ma}) is expressed as (for ψ_{ma}, volume fraction of moisture in pore space, $0 < \psi_{ma}^{2/3} < 0.4$)

$$\lambda_{ma} = \lambda_a \left(1 + 3.844 \frac{\lambda_w - \lambda_a}{\lambda_w + 2\lambda_a} \psi_{ma}^{2/3} \right). \quad (15)$$

When ψ_{ma} lies between 0.4 and 1.0, the thermal conductivity of moist air becomes

$$\lambda_{ma} = \lambda_w \left(1 + 3.844 \frac{\lambda_a - \lambda_w}{\lambda_a + 2\lambda_w} (1 - \psi_{ma}^{2/3}) \right). \quad (16)$$

The thermal conductivities of water (λ_w) and air (λ_a) are expressed as

$$\lambda_w = 0.55 + 2.34 \times 10^{-3}\, T - 1.1 \times 10^{-5}\, T^2$$

and

$$\lambda_a = 0.0237 + 6.41 \times 10^{-5}\, T.$$

The volume fraction of moisture in the pore space will be

$$\psi_{ma} = \frac{\psi_m}{\psi_a}$$

where $\psi_m = (m/M)\psi_a$. Here m and M represent the varying moisture content and the moisture content at saturation by weight percent, respectively. In the case of a frozen system, λ_w will be replaced by the thermal conductivity of ice (λ_{ice}).

3. RESULTS AND DISCUSSION

The results of the study are presented in Figs. 2–6. In Fig. 2 the transient temperature measurements are presented. This figure presents temperature recordings for various times after the temperature at the upper end of the column is lowered suddenly to -5°C (below the freezing point of water) and the bottom end of the column is at 30°C, which is approximately room temperature. From Fig. 2 it can be observed that the soil temperature decreases and approaches a linear distribution, because of the initial uniform moisture content, and then approaches a steady-state non-linear profile. It might be due to the fact that the

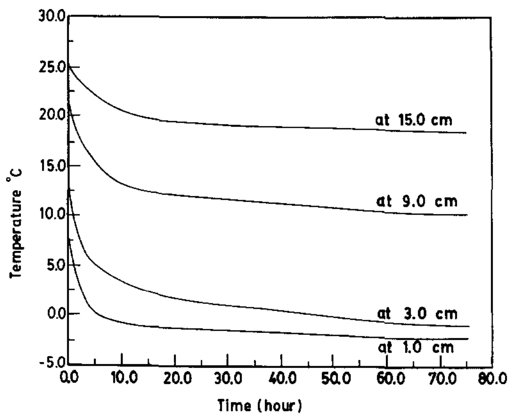

Fig. 2. Variation of non-dimensional temperature $[(T-T_C)/(T_H-T_C)]$ with non-dimensional distance (x/L) from cold end ($L = 27.0$ cm, $T_c = -5$°C and $T_H = 30$°C).

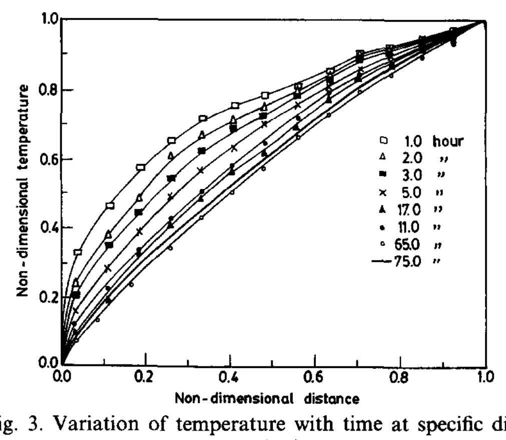

Fig. 3. Variation of temperature with time at specific distances (cm).

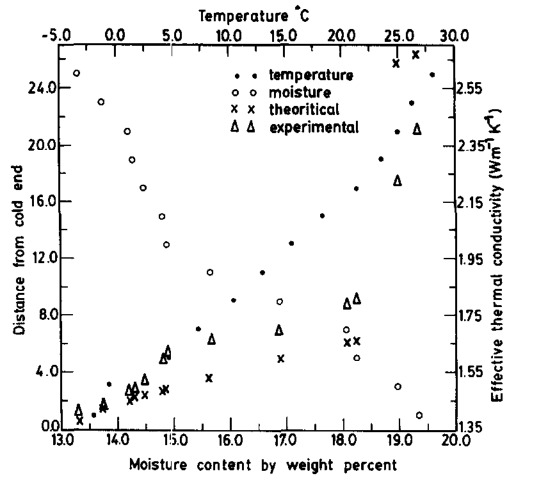

Fig. 4. Variation of effective thermal conductivity (ETC) with moisture content and distribution of steady-state temperature and moisture with axial distance from cold end (cm).

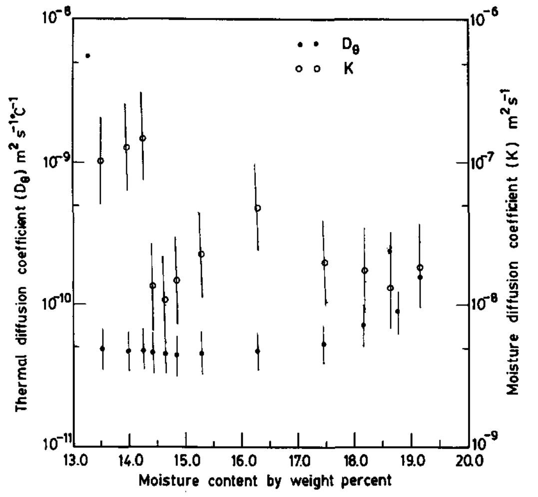

Fig. 6. Variation of thermal and moisture diffusion coefficients with moisture content.

effective thermal conductivity of moist porous materials primarily depends upon moisture content and composition [10, 11]. Also if the effective thermal conductivity of the moist material is to remain uniform, then a linear temperature distribution would be expected. As the heat and moisture migration is a coupled process, therefore the steady-state temperature profile will be non-linear, because of moisture redistribution resulting from the applied temperature difference. Since the slope of the temperature profile decreases with increase in axial distance from the cold end, this profile shows the characteristics of the material, whose ETC decreases with x/L. Transient temperature measurements reveal that the temperature adjusts rapidly compared with the moisture and that a quasi-steady state exists during the time of moisture redistribution.

In Fig. 3 the variation of temperature with time at specific distances is presented. This figure presents the effectiveness of time-dependent cooling of the material with the axial distance from the cold end. In Fig. 4 the steady-state temperature and moisture profiles are shown as a function of axial distance from the cold end. These steady-state temperature and moisture profiles have been used for the evaluation of diffusion coefficients. Also on this figure are plotted the measured and calculated values of the effective thermal conductivity. The Singh *et al.* model [4] was used to calculate the ETC values. A reasonable agreement has been found between the measured and calculated ETC values. A sharp increase in the ETC has been found at the interface that might be due to the freezing of soil. At the interface, the moisture content within the soil freezes, which increases the ETC of soil because freezing water has a thermal conductivity which is four times greater than that of free water. In the calculations, the thermal conductivity of ice was taken to be 2.20 W m^{-1} K^{-1}.

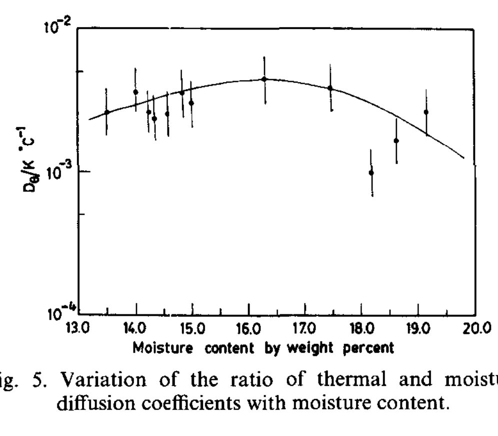

Fig. 5. Variation of the ratio of thermal and moisture diffusion coefficients with moisture content.

An analysis of the proposed diffusivity term β_j and some points of interest are discussed here. The new diffusivity term is

$$\beta_j = \alpha_j \frac{(1+\phi_j/Le_j)}{(1+\phi_j)}$$

where $Le_j = \alpha_j/D_j$ represents the Lewis number, which is of basic importance in determining the behaviour of the system. The term ϕ, which also affects the magnitude of the difference between α_j and β_j, is the ratio of energy stored in the vapour enthalpy to that stored in the internal energy

$$\phi_j = \frac{\varepsilon_g b_j h_j}{(\rho c)_j}.$$

The vapour density temperature slopes for the frozen and unfrozen regions are $b_1 = 1.94 \times 10^{-4}$ and $b_2 = 6.18 \times 10^{-4}$, respectively. The magnitude of Le_j presents the existence of evaporation or condensation. When $Le_j > 1$ evaporation will occur in the system, which will increase the thermal diffusivity of the system. The ratio of the energy flux in the vapour enthalpy to that in the internal energy is

$$\frac{\phi_i}{Le_j} = \frac{D_j b_j h_j \varepsilon_g}{\alpha_j(\rho c)_j} = \frac{D_j b_j h_j \varepsilon_g}{\lambda_{ej}}.$$

In Fig. 5 variation of the ratio of thermal and moisture diffusion coefficient (D_θ/K) with moisture content is presented. The ratio of D_θ/K is a function of both composition and soil moisture content. Variation of the ratio can also be seen by combining the above two with moisture content individually. These results also verify that the ratio should not be more than 10^{-2}, as proposed by Luikov [12] from the theoretical analysis. It can be concluded from the graph that the ratio becomes greater with the increase of moisture content and thereafter decreases. The ratio is small for a low moisture content. As the moisture content increases, the ratio also increases up to a certain value of moisture content, which depends on soil type. Further increase in water content results in a decrease in the ratio of the diffusion coefficients. The decrease in D_θ/K at higher water content is mainly because of the decrease in porosity of soil, which reduces the amount of vapour transfer. The experimental data presented by Luikov [12] indicated that D_θ/K is approximately zero when moisture content reaches saturation. In the dry region of the soil, the ratio should also decrease to a small value (i.e. zero), since the mass transfer of water vapour is not possible due to the absence of liquid water. Figure 4 shows a small dip in the moisture content distribution between 0.7 and 0.85 (x/L). This dip affects the $\mathrm{d}W/\mathrm{d}x$ values in the region and hence causes the aforementioned D_θ/K valley, which may be caused by measurement error. The ratio of thermal and moisture diffusion coefficients has been found to be in the range of 10^{-3}–10^{-2}°C^{-1}. Based on theoretical analysis and experimental data, Luikov has proposed that, for the coupled heat and mass transfer process in porous media, the ratio of diffusion coefficients should be less than 1.0×10^{-2}°C.

The variation of individual thermal and moisture diffusion coefficients with moisture content has been plotted in Fig. 6. The moisture diffusion coefficient exhibits a broad minimum, because it is inversely proportional to D_θ/K and D_v varies monotonically with W/W_S. For the sand sample, the thermal diffusion coefficient has been found to be in the range 10^{-10}–10^{-11} m^2 s^{-1}°C^{-1} and the moisture diffusion coefficient has been found to be in the range of 10^{-8}–10^{-6} m^2 s^{-1}(10^{-8}–10^{-7} for unfrozen soil and 10^{-7}–10^{-6} for frozen soil). The order of these values is in agreement with results of Wang and Yu [13].

The analysis of the dryout condition presents a method for determining the existence of a dryout region by knowing only the boundary conditions and the ratio of diffusion coefficients. It has been found that, for the present thermal boundary conditions, dryout is possible only when the initial moisture content is less than 3.0%. Thus the existence of a dryout region will depend upon the initial moisture content. The extent of the dryout region increases with a decrease in initial moisture content. Dryout starts at a time when the temperature field has already reached a steady state. It can be observed that the moisture content decreases near the hot surface of the slab and increases near the cold surface. After a certain time, the moisture content at the hot surface reaches the value zero. At that moment, dryout starts and the dryout region expands into the interior of the slab, an asymptotic steady state is finally reached in which the moisture content increases linearly with increasing depth of the slab. The depth to which dryout penetrates increases with larger values of thermal/mass diffusion parameters. The total moisture content in the slab must remain unchanged. As such the local moisture content value at the cold side increases with increase in value of the mass diffusion parameter.

4. CONCLUSION

Ratios of diffusion coefficients, D_θ/K, were experimentally determined by measuring the one-dimensional temperature and moisture distribution in the temperature range of −5–30°C. The ratio of D_θ/K was found to be function of the moisture content. The ratio was found to be small for low moisture content values, increases with moisture content and thereafter decreases. The ratio D_θ/K was found to be in the range of 10^{-3}–10^{-2}°C. For the soil used in this study, the moisture diffusion coefficient was estimated to be in the range of 10^{-8}–10^{-6} m^2 s^{-1} (10^{-8}–10^{-7} for unfrozen region and 10^{-7}–10^{-6} for frozen region).

The transient thermal response of the sand is much quicker than its moisture response, the thermal response soon settles down to being quasi-steady. The steady-state moisture and temperature profiles were found to be functions of initial moisture. In general, an increase in initial moisture content leads to uniform moisture distribution, thus a uniform thermal conductivity and linear temperature profile. Since both D_θ and K are physical properties, they should be independent of time. Therefore, the diffusion coefficient determined at any specific instant (e.g. quasi-steady state) must be applicable through the process.

The results show that the effective thermal conductivity model is adequate for a system of porous media under different conditions. This analysis establishes a method of predicting the existence of a dryout region for different boundary conditions using the diffusion coefficients.

The moisture diffusion coefficient K is very sensitive to moisture content. Small errors in the measurement of the moisture content give rise to large errors in the K value.

Acknowledgements—The authors are thankful to Dr R. Singh and Mr A. K. Shrotriya for discussion and suggestions during preparation of the manuscript. Financial assistances from DNES, New Delhi and TWAS, Trieste (Italy) are gratefully acknowledged.

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