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APPLICATION OF EVAPORATION METHOD USING TWO TENSIOMETERS FOR DETERMINING UNSATURATED HYDRAULIC CONDUCTIVITY BEYOND TENSIOMETRIC RANGE

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Abstract – Accurate determination of unsaturated hydraulic conductivity in low pressure head range is critical for predicting evaporation rate under localized irrigation or root water uptake under drought stress conditions. We proposed a relatively low cost and fast laboratory method to simultaneously determine water retention and hydraulic conductivity functions across a wide range of pressure heads. The method is quite similar to conventional evaporation method using two tensiometers. In addition to tensiometer readings, the proposed method uses water content profile at the end and cumulative evaporation. Experimental results for three soils with different texture showed that inversely optimized hydraulic conductivity functions agreed with K- data measured with the steady state evaporation method, indicating the reliability of the proposed method. The hydraulic conductivity functions fitted for K-data obtained by tensiometer readings with Campbell's K(θ) function also agreed well with the reference K- data for two of the three soils, but largely deviated from those for a soil. This indicates the importance of actual measurement of K(θ) in low pressure head r*ange.*

Keywords: evaporation, drought, unsaturated hydraulic conductivity, WASH_1D

INTRODUCTION

Knowledge of soil hydraulic properties which consists of soil water retention function and the hydraulic conductivity function is a prerequisite for predicting evapotranspiration rate under water-limiting conditions as well as the water flow in soils by solving the Richards equation. In particular, accurate determination of unsaturated hydraulic conductivity in low pressure head range (less than - 800 cm, in this study) is critical for predicting evaporation rate under localized irrigation or root water uptake under drought stress conditions.

Various methods to determine the soil hydraulic properties have been developed since early days of soil physics. Retention data can be measured with equilibrium methods and unsaturated hydraulic conductivity can be measured with steady-state methods [1]. Such equilibrium or steady-state methods are straightforward and accurate as long as equilibrium or steady-state are attained, but the time demand for these methods can be prohibitively large for lower water content. For this reason, various inverse methods under transient flow such as multi-step-outflow methods [2-4] or evaporation methods [5-8] have been developed to fasten duration for the experiments.

In practice, the reliable pressure head range of the multistep-outflow methods is limited to smaller suctions because at lower water content, samples may lose contact with the ceramic plate placed below the soil cores [2,4]. Fujimaki and Inoue (2003) [9] also showed the existence of a hydraulic resistance at the soil–porous plate interface even in low suction and they inferred that the resistance was caused by pore plugging with fine particles transported to the ceramic plate at each stepwise changes in pressure.

Evaporation methods using tensiometers are immune from such resistance problems, and it seems that evaporation methods are more commonly used partly promoted by the availability of commercial devices for that method (e.g. the HYPROP TM sys-tem from UMS, Germany).

Currently, however, there is a lack of simple and quick methods to reliably obtain soil hydraulic conductivity functions in the dry range beyond the pressure head range measurable with tensiometer [10].

Fujimaki and Inoue (2003) [11] presented a flux-controlled steady-state evaporation method for determining unsaturated hydraulic conductivity at low matric pressure head values. This method would give reliable data, since this does not depend on assumptions invoked in direct methods by Schindler (1980) [6] or inverse methods. But this method requires independent measurement of retention function and does not give unsaturated hydraulic conductivity at large water diffusivity range. Fujimaki and Inoue (2003) [12] also presented a combined method of multistep-outflow methods at low suction range (< 160cm) and an evaporation method for high suction range. They used evaporation rate and water content profile at the end, but did not use outflow data for inverse parameter estimation of hydraulic conductivity function for the above reason and therefore, accuracy of hydraulic conductivity function in low and middle suction range is somewhat unreliable.

Schelle et al. (2011) [10] also presented a combined method of multistep-outflow and evaporation method using improved tensiometers which resist cavitation to much lower pressure heads than conventional tensiometers for determining soil hydraulic properties in a wide pressure head range

Likewise, possibilities to extend the measurement range are the inclusion of advanced matric potential sensors in the evaporation method, such as MPS1 sensors (Decagon Devices, 2009), polymer tensiometers [13]. This approach may be sound as long as additional cost for equipment is not a concern.

To extend the measurement range without using expensive non-conventional matric potential sensors, the use of the air-entry value of the conventional tensiometer cup in the evaporation method as an additional measurement point has been proposed [14]. This method has the disadvantage of reliability on the framework of numerical inversion compared with directly measured hydraulic properties. It also requires measurement of air-entry value for each ceramic cup in a separate experiment.

Thus, there is still a lack of low cost and quick methods to reliably determine soil hydraulic conductivity functions in the dry range beyond the workable limit of conventional tensiometers. This lack has lead wide use of estimation methods for hydraulic conductivity function from parameters in retention function presented by Campbell (1974) [15] or other studies [16-22] or inverse determination using observed soil moisture data in fields [23,18,19], which may be less accurate than actually measured hydraulic properties in laboratories.

By simply continuing evaporation even after air-entry into tensiometers, we may obtain steep gradient of soil water content in surface layer at the end and decreasing evaporation rate, which can be used as critical information for inverse parameter estimation of hydraulic conductivity. The objectives of this study, therefore, were i) to propose a relatively low cost and fast method of the extended evaporation method to simultaneously determine water retention and hydraulic conductivity functions across a wide range of pressure heads using such a critical information and ii) to evaluate if extrapolation of hydraulic conductivity function obtained with tensiometer readings only gives satisfactory accuracy.

MATERIAL AND METHODS

Three soils were used: a loamy sand (ACRISOLS) taken from Khaosan-Kwan district, north-east Thailand, a sandy loam (REGOSOLS) taken from Iwami town, western Japan, and a loam (ANDOSOL) from Tsukuba, eastern Japan. Basic characteristics of those soils are listed in Table 1. Before air dry, saturated hydraulic conductivity of each soil was measured with the falling head method and after the falling head experiment, those samples were well leached with distilled water to remove soluble salts until electrical conductivity of leachate became less than 0.04dS/m.

Experimental Setup

As depicted in Fig.1, a 100 cm3 core sampler with 5 cm diameter and 5.1 cm height was inserted into a cylinder excavated on an acrylic pipe with 10 cm diameter and 7.1 cm height. Two tensiometers were inserted from the bottom so that the center of each porous cup were located at the depths of 1.55 and 3.55 cm, respectively. Tensiometers were connected with pressure transducers which were then connected with data loggers. Dummy pipes were mounted on porous cups to avoid convergence flow above the porous cups, which may violate the assumption of one dimensional flow.

A thermocouple was inserted into near the soil surface to keep the temperature constant. The soil column and connected devices except for thermocouple were mounted on an electronic balance to automatically record weight at an interval of 5 minutes.

Procedure

Air-dry soil samples were filled into the core sampler so that bulk density became intended values. Each soil sample was saturated from the bottom with distilled water. When tensiometer reading become steady, soil surface was uncovered to allow evaporation under nearly constant meteorological conditions (25 °C and relative humidity at 30%), except for radiation, which was automatically regulated using a thermostat such that the soil temperature remained constant at 25 °C.

Evaporation was accelerated by blowing air across the soil surface with an electric fan. When suction at the upper tensiometer reached greater than 700 cm, water in the tensiometers were sucked with a syringe. After removing the water in the tensiometers, the soil column was again placed under the evaporative condition until the evaporation rate became lower than 20% of that at the initial stage. After termination, the soil columns were dismantled to obtain the water content profiles. Soil samples at depths of 0-0.5 cm and 0.5-1.0 cm were used to measure water potential using a psychrometer (Decagon WP4) to obtain water retention data beyond tensiometric range. To obtain matric potential, osmotic potential estimated from electrical conductivity of water of the samples were subtracted from water potential.

Steady-State Evaporation Experiment

To check the reliability of the hydraulic conductivity function determined with the presented method and to evaluate if extrapolation of hydraulic conductivity data obtained with tensiometer readings only gives satisfactory accuracy, steady-state evaporation experiments (SEM) for measuring hydraulic conductivity under low pressures were performed. Detailed experimental setup and procedure are presented in Fujimaki and Inoue (2003)[12]. Figure 2 shows measured and fitted water content profiles at the steady state.

DATA ANALYSIS

Retention Function

Average volumetric water content θ at each time was calculated by water balance:

$$\bar{\theta} = \overline{\theta_{end}} + \frac{M - M_{end}}{\rho_w V} \tag{1}$$

where M is mass of bulk soil (g), V is volume (cm^3) of soil and ρ_w is density of water (g cm^{-3}) and subscripts *end* represent values at the end. Retention data were obtained by linking θ and average pressure head of the two tensiometers. Time evolution of cumulative transpiration and tensiometer reading are shown in Fig. 3. Note that the experiments were terminated within 30 hours after starting evaporation. We fitted retention data for Khaosan-Kwan loamy sand with the bimodal retention function:

$$\theta = \frac{s\theta_{sat}}{\{1 + (-\alpha_1\psi)^{n_1}\}^{1-\frac{1}{n_1}}} + \frac{(1-s)\theta_{sat}}{\{1 + (-\alpha_2\psi)^{n_2}\}^{1-\frac{1}{n_2}}} \tag{2}$$

where θ_{sat} is θ at $\psi = 0$, and s, α_1, n_1, α_2, n_2 are fitting parameters. The θ_{sat} was calculated as 0.275 from water balance. Fitted values were s= 0.8, α_1= 0.112, n_1= 3.77, α_2= 0.00119, and n_2 = 1.49. We did not use tensiometer readings lower than -300 cm for this sand, because even small leakage from porous cup due to bubble expansion may sharply drop suction around the cup owing to both low water capacity and water diffusivity of sandy soils in high suction range. Likewise, tensiometer readings lower than -500 cm for Iwami sandy loam were not used.

For Iwami sandy loam and Tsukuba loam, the retention data were fitted with the empirical equation [12]:

$$\theta = \frac{\theta_{sat} - \zeta}{[1 + (-\alpha\psi)^n]^m} + \zeta\{1 - [\frac{\ln(-\psi + 1)}{\ln(-\psi_0 + 1)}]^2\}$$

(3)

where α, ζ, *n*, and *m* are fitting parameters. The parameter ψ_0 is the pressure head where the water content becomes nearly zero (i.e. oven dry). In this study, ψ_0 was set to -10^7 cm, while *m* was handled as an independent fitting parameter.

Figure 4 shows measured and fitted retention data and Table 2 lists fitted parameter values for Iwami sandy loam and Tsukuba loam. These fitted retention function were used for both inverse method described below and the steady-state evaporation method.

Direct Method for Determining Hydraulic Conductivity

Hydraulic conductivity *K* (cm s^{-1}) was calculated by inversely applying Darcy's law with finite difference:

$$K = -\frac{q}{\frac{\psi_2 - \psi_1}{z_2 - z_1} - 1}$$

(4)

where *q* is flux (cm s^{-1}), and *z* is depth (cm).

In a short column whose lower boundary is impermeable, *q* linearly increases with height from the bottom except for very beginning and duration when soil surface is near air-dry [24]. Thus, at the center depth, *q* is half of the evaporation rate, *E* (cm s^{-1}). This method is essentially the same as Schindler's one [6].

Evaporation rate can be calculated by dividing weight reduction rate by area, *A* (cm^2), of the soil and density of water.

$$E = -\frac{1}{\rho_w A}\frac{\partial M}{\partial t}$$

(5)

Cross-sectional area of the tensiometers was subtracted from the area of core (19.63 cm^2). Usually, readings of electronic balance are fluctuated by wind under such experimental conditions. Therefore, weight reduction rate was calculated by fitting the time-weight curve with an appropriate empirical function and differentiating it at each time.

We fitted the time evolution of cumulative evaporation with integral of the logistic function plus a constant:

$$\int_0^t Edt = -\frac{a_e}{c_e}\{\ln[\frac{a_e - b_e}{b_e}\exp(-c_e t)] - \ln[1 + \frac{a_e - b_e}{b_e}\exp(-c_e t)]\} + d_e t \qquad (6)$$

where a_e, b_e, c_e, and d_e are fitting parameters. Figure 3 also shows fitted cumulative evaporation.

Then, evaporation rate at any time is given by the logistic function plus a constant:

$$E = \frac{a_e}{\frac{a_e - b_e}{b_e}\exp(-c_e t)} + d_e \tag{7}$$

Inverse Method for Determining Hydraulic Conductivity

After determining the water retention function, the parameter values in the following function was inversely estimated using the Levenberg-Marquardt's maximum neighborhood method [25] combined with a one-dimensional numerical solution of the water flow equation.

$$K = K_{sat}(\frac{\theta}{\theta_{sat}})^{\omega}$$
$$\omega = b_k\theta + c_k \tag{8}$$

where b_k and c_k are fitting parameters. The term $b_k\theta$ was added to enhance flexibility of the function. If b_k is 0, it becomes the same as widely used Campbell's function [15].

Hydraulic conductivity calculated with the direct method, cumulative evaporation and the water content profile were used in the objective function to be minimized in the algorithm:

$$O(b_k, c_k) = \sum_{j=1}^{3}(\frac{1}{n_j\sigma_j^2})\sum_{i=1}^{n_j}[p_{j,i} - p_{j,i}(b_k, c_k)]^2 \tag{9}$$

where j denotes the different sets of measurements, n_j are the numbers of measurements within particular sets, p_{j,i^*} are the measurements of type j at time t_i, $p_{j,i}$ are the corresponding model predictions using a_k and b_k, and σ_j are the variances of the measurements of data type j. In this study, type $j = 1$ was the logarithm of hydraulic conductivity calculated with the direct method, $j = 2$ was cumulative evaporation, and $j = 3$ was the final water content profile.

The water flow equation including isothermal vapor movement was solved by the finite difference method based on the mass-conservative iterative scheme proposed by Celia et al. (1990) [26] and Fujimaki and Inoue (2003) [12]. Space increments were set at 0.05 cm at the soil surface and 0.1 cm at the bottom and between them were given with nearly geometric progression. Time steps were controlled automatically so that the number of iterations in each time step was around five and the maximum change of ln(ψ) in each time step was less than 0.693 (= ln(2)). The initial conditions were equilibrium pressure head profiles based on initial tensiometer readings. The lower boundary condition was zero flux, while the upper boundary condition was the atmospheric boundary condition where the evaporation rates at each time increment were calculated using bulk transfer equation [12].

The aerodynamic resistance was determined from the evaporation rate during the first five hours using the bulk transfer equation, because the evaporation rate is nearly constant and relative humidity at the soil surface is kept at approximately 1.0 for some time after the start of a run.

The inverse analysis was carried out with WASH_1D code, which we are developing. It is freely distributed with the source code and data files for those experiments under the general public license on the website web site of Arid Land Research Center, Tottori University (http://www.alrc.tottori-u.ac.jp/fujimaki/download/WASH_1D/).

Figure 3 also shows numerical solutions for cumulative evaporation and pressure head at the tensiometer depths using the optimum parameter values. Simulated cumulative evaporation tended to overestimate at the beginning of second stage of evaporation. This might partly be owing to the assumption of uniform relative humidity of air across the soil surface. We observed that soil surface in the upwind dried slightly faster than in downwind. Relative humidity of air just above the soil surface should increase more or less with downwind owing to convective transport of water vapor and it cannot be considered in one-dimensional model. Automatically switching blow-direction alternatively may minimize this effect, but it is beyond the scope of this study. Diverging discrepancy in pressure head at tensiometer depths for the sandy soils may partly be caused by this overshoot. Another reason may be inability of the determined retention function to accurately describe retention curve at high suction range for sandy soils.

Figure 5 shows those for water content profiles. Numerical solutions were in fair agreement with measured ones.

RESULTS AND DISCUSSION

Directly calculated K data and optimized $K(\theta)$ functions are plotted in Fig.6. Obtained parameter values are listed in Table 3. Optimized $K(\theta)$ functions were in fair agreement with K- data obtained with the SEM, indicating the reliability of the method. Root mean square errors for common logarithm of $K(\theta)$ and K-data obtained with the steady-state evaporation method are listed in Table 4. The $K(\theta)$ functions with $b_k = 0$ fitted for K-data calculated with the direct method also agreed well with K- data obtained with the SEM except for Tsukuba loam. If restriction of $b_k = 0$ was lifted, fitted curves more agreed with K-data calculated with the direct method for Tsukuba loam, but they consistently underestimated at very low pressure head range. These results indicate that extrapolation of K-data beyond tensiometric range does not necessarily give reliable $K(\theta)$ function which is applicable to low pressure head range and the importance of measurement of $K(\theta)$ in that range. The proposed method requires temperature controller and room with constant temperature and relative humidity compared with simple evaporation method with tensiometers. But temperature controller is commercially available at a cost of less than 100 USD (e.g. Omron E5CB). Also, room temperature can easy be kept constant with air-conditioner commonly used in households. Keeping relative humidity at lower limit of dehumidifier is quite easy in a narrow room. The use of WP4 psychrometer can be replaced with conventional vapor equilibrium which can be accomplished at very low cost. Therefore, the proposed method essentially just requires additional work of taking water content profile to conventional evaporation method with tensiometers. Measurement of water content profile is quite easy and does not require any additional cost for general soil laboratories.

CONCLUSIONS

We have proposed an inexpensive and fast method of the extended evaporation method to simultaneously determine water retention and hydraulic conductivity functions across a wide range of pressure heads. In addition to tensiometer readings, the proposed method uses water content profile at the end and cumulative evaporation. Inversely optimized hydraulic conductivity functions agreed with K- data measured with the steady state evaporation method for three soils with different texture, which indicates the reliability of the method. The hydraulic conductivity functions fitted for K-data obtained by tensiometer readings with Campbell's $K(\theta)$ function also agreed well with K- data measured with the steady state evaporation method for two of the three soils, but largely deviated from those for a soil. This indicates the importance of actual measurement of $K(\theta)$ in low pressure head range.

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TABLES .

Table1 Basic characteristics of the two soils used in the experiments.

	Particle size distribution (%)			Bulk density	Saturated hydraulic conductivity
	Sand	Silt	Clay	Mg m^{-3}	cm s^{-1}
Khaosan-Kwan loamy sand	88	5	7	1.60	0.00057
Iwami sandy loam	84	7	9	1.23	0.00470
Tsukuba loam	50	44	6	0.78	0.00126

Table 2 Fitted parameter values for Eq.(3).

	θ_{sat}	ζ	α	n
Iwami sandy loam	0.50	0.552	0.0500	1.44
Tsukuba loam	0.64	0.126	0.0183	1.34

Table 3 Inversely determined and fitted parameter value in the $K(\theta)$ function

	Loamy sand		Sandy loam		Loam	
	b_k	c_k	b_k	c_k	b_k	c_k
Proposed method	-20.4	7.72	-4.76	9.07	-15.4	15.20
Fitted with b_k =0	0.0	6.03	0.00	8.12	0.0	9.87
Fitted with $b_k \neq 0$	-20.1	7.57	-9.18	9.94	-38.0	23.10

Table 4 Root Mean Square Errors for common logarithm of $K(\theta)$ and K-data obtained with the steady-state evaporation method

	Loamy sand	Sand loam	Loam	average
Proposed method	0.78	0.46	0.48	0.57
Fitted with b_k =0	0.22	0.20	1.03	0.48
Fitted with $b_k \neq 0$	0.71	0.79	0.77	0.76

FIGURES

Fig. 1. Schematic of the experimental setup.

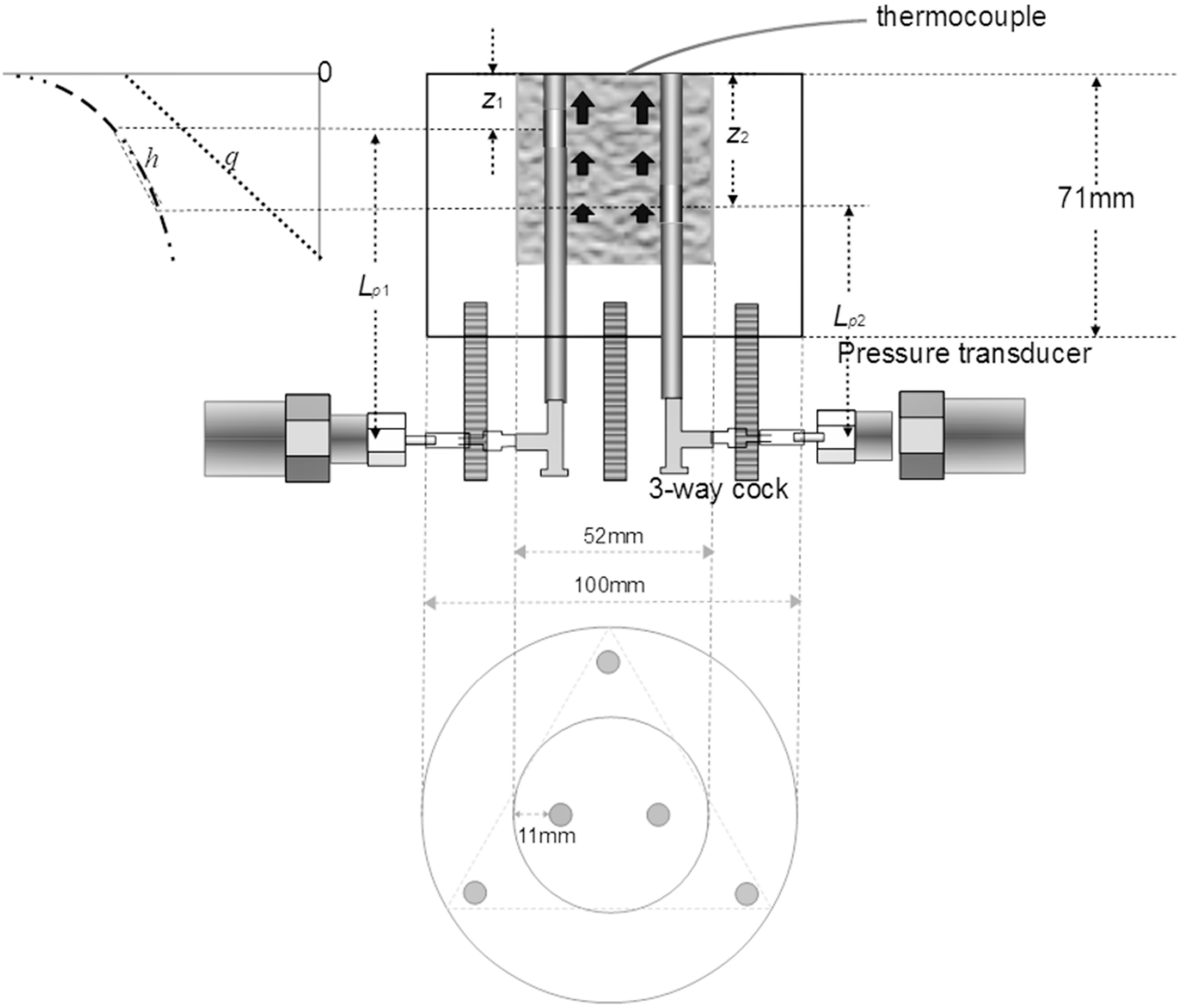

Fig. 2. Measured and fitted water content profile for the steady-state evaporation experiment

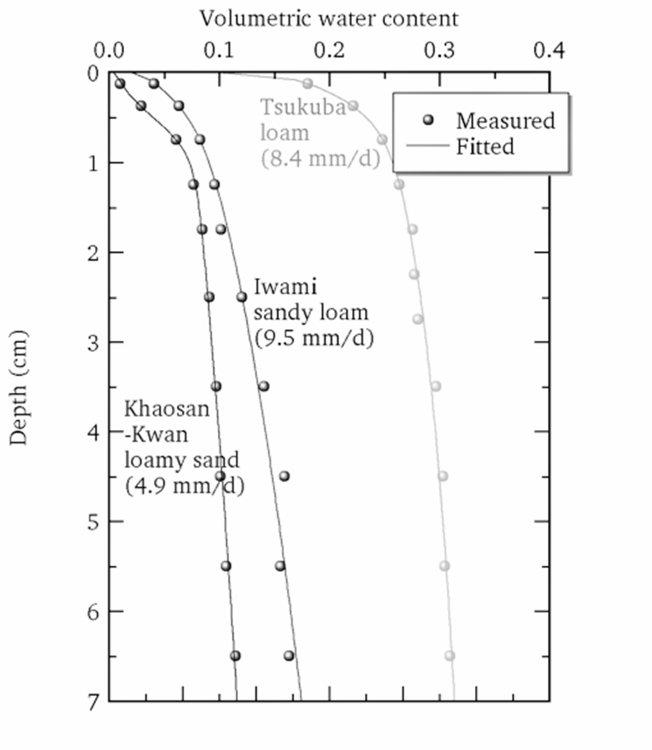

Fig. 3. Time evolution of tensiometer readings and cumulative evaporation

Figure caption: Fig. 3. Shows the time evolution of tensio meter readigs ad cumulative evaporation of Khaosan-Kwan lomy sand (a), iwami sandy loam (b) and Tsukuba loadm (c).

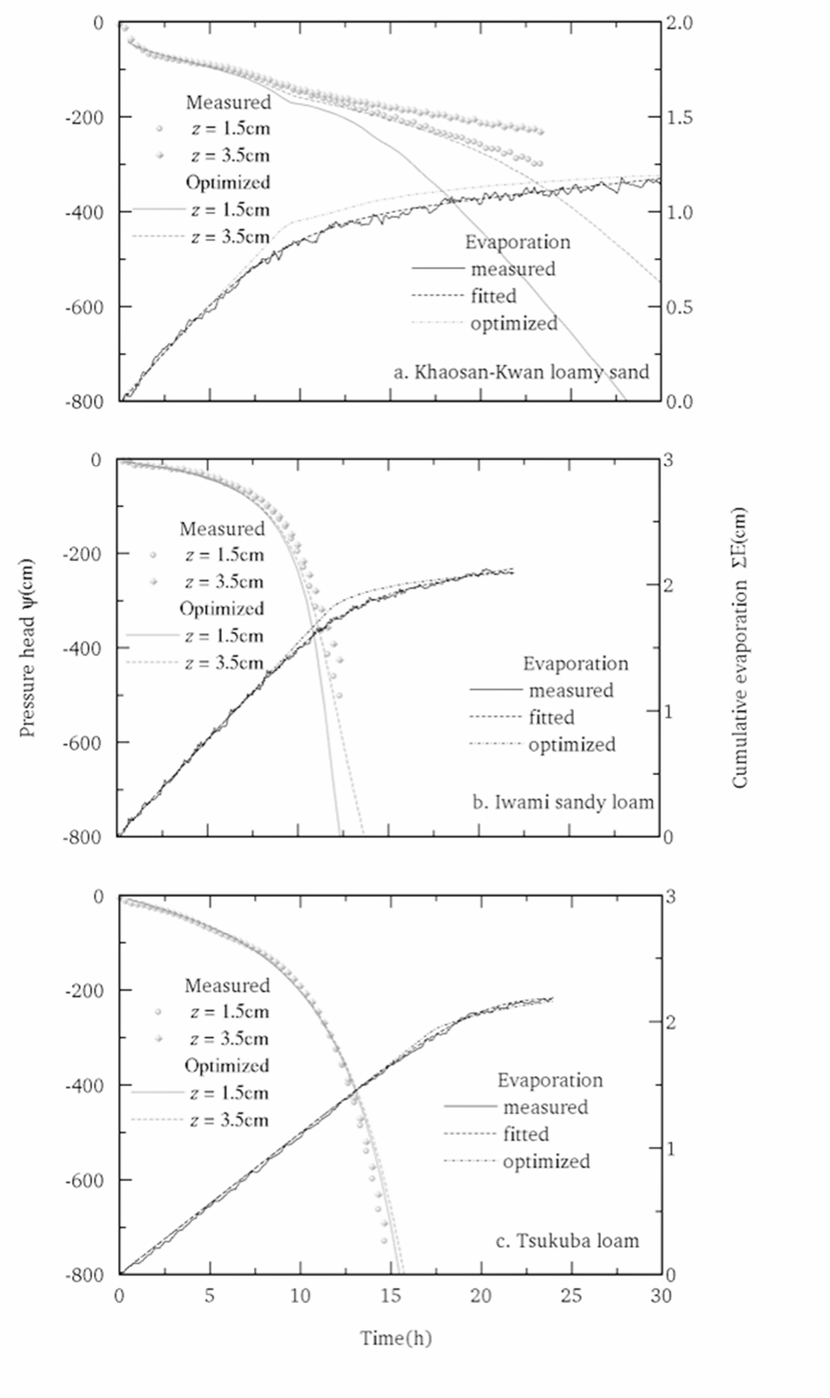

Fig. 4. Soil water retention curves for the soils.

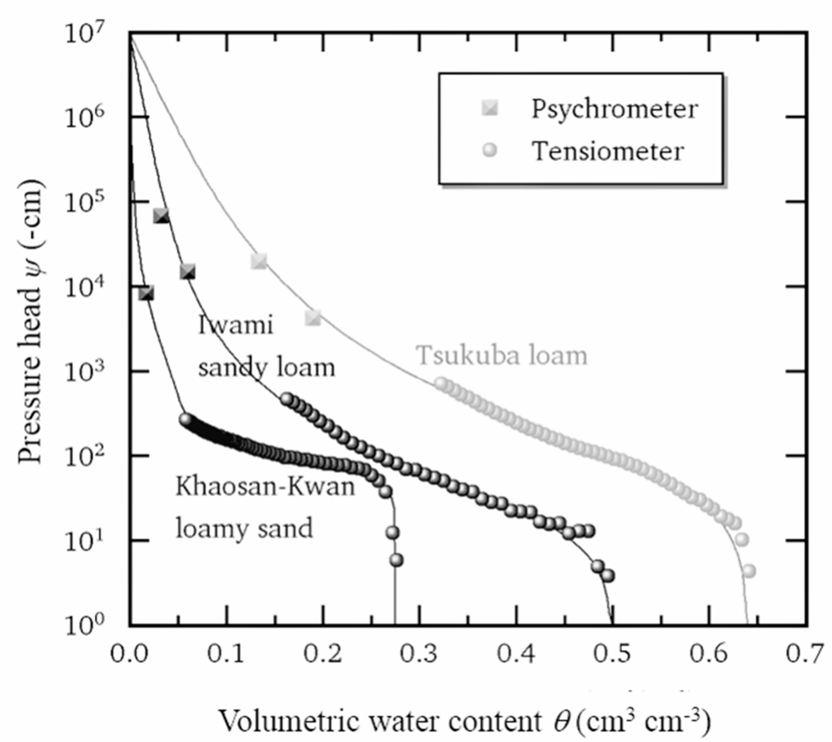

Fig. 5. Water content profiles at the end of the soils.

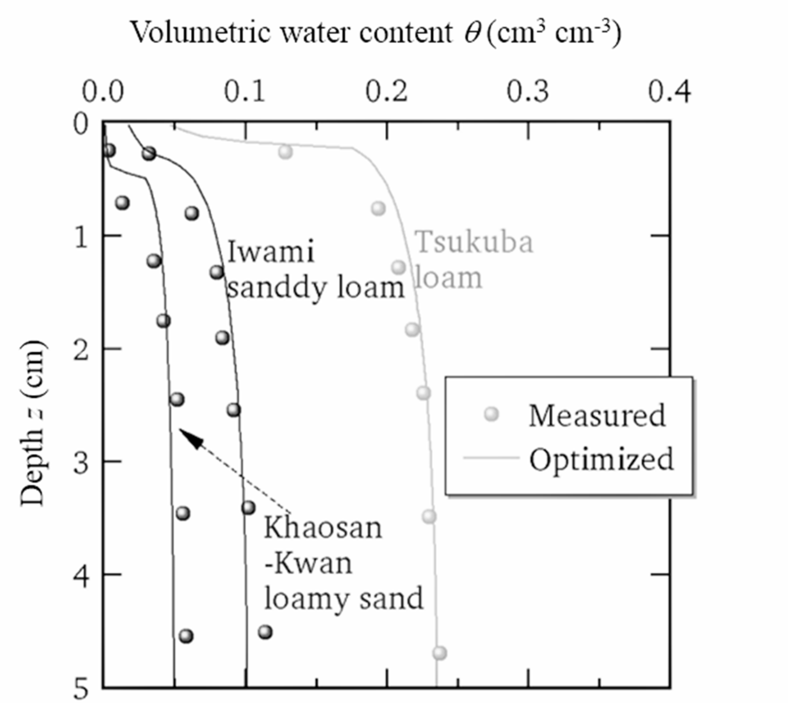

Fig. 6. Unsaturated hydraulic conductivity for the soils.