Spatial Variability of Hydraulic Properties in a Multi-Layered Soils of Japanese Larch (*Larix leptolepis*) Stand*

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落葉松林分의 多層構造 土壤에 있어서 水理特性의 空間 變異*

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ABSTRACT

Soil structure and organic matter have been known to strongly affect water flow and solute transport, yet little information is available concerning soil hydraulic properties related to soil physical and chemical properties in the forest site. The purpose of this study was to quantify the spatial variability and spatial correlation of the measured parameter values from the plots established with the rainfall simulator on Japanese larch(*Larix leptolepis*) dominated site in Kwangju, Kyunggi-Do. Measurement of soil water flux and retention were made with the inherent soil texture, soil structure, and organic matter. The method was based on the observation that when water was applied at a constant rate to the soil surface on each plot. The method was simple to apply and consists of following steps: (i) Wet the soil from a rainfall simulator with several known discharge rates on a relatively leveled soil surface with and without organic matter. (ii) Once the borders of the ponded zone were steady, saturated hydraulic conductivity(K_s) and the matric flux function(F) was evaluated from a regression of flux vs. the reciprocal of the ponded area. A conductivity of the form $K_{i+1}(\psi_c)= K_i(\psi_c)$ $[1-d\psi/dz]$ where flux continuity implies. For this, continuity of matric potential at the interface at all times are as follows: $\psi_1(Z_c) = \psi_2(Z_c) = \psi_c$ for steady state intake from water ponded on the soil surface. Results of this investigation showed the importance of understanding spatial variability in wide differences of water retention and saturated hydraulic conductivity with respect to pore geometry and organic matter contents which influenced the water flux throughout the soil profile.

Key words : *Larix leptolepis*, spatial variability, hydraulic properties, multi-layered soil, water potential

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요 약

본 연구는 경기도 광주지역의 낙엽송 임분에 인공 강우기를 설치하여 수리특성과 관련한 공간 변이와 상관관계를 구명하기 위하여 수행되었다. 수분유속과 보유능은 원래 토성, 구조와 유기물을 근거로 지표면에서 인공강우기로 수분을 일정속도로 가하면서 측정하였다. 이 방법은 간단히 유기물을 제거하거나 기존의 상태에서 수평한 지표면에 3가지의 강우 강도에서 실시하였으며 한편 경계면은 안정하며 포화상태의 습윤 토양에서 투수계수와 매트릭 유속기능을 조사하였다. 투수계수는 $\psi_1(Z_c) = \psi_2(Z_c) = \psi_c$ 이라는 지속적 유속 조건하에서 $K_{i+1}(\psi_c) = K_i(\psi_c)\ [1-d\psi/dz]$의 공식을 적용하였다. 본 연구 결과 토양내에서 유속에 영향을 미치는 공극 특성과 유기물 함량이 광범위하게 수분 보유능과 포화투수계수의 공간 변이에 영향을 미치는 것으로 조사되었다.

INTRODUCTION

Preferential paths and spatial variability in transport properties, such as hydraulic conductivity, greatly influence the transport of solutes from the cycling of nutrients in terrestrial ecosystem, particularly in high rainfall season. In rangeland site, the development of soil horizons is greatly dependent on the erosive and illuvial patterns of soil particles and organic matter governed by climatic factors and vegetation cover, resulting in a distinctive soil textural and structural characteristics through soil profile. Understanding the indicator values of vegetation and soil characteristics which are correlated over broad areas of the forest landscape can be useful to hydrologists and environmentalists, although vegetation and soil characteristics are independent integrators of environmental factors.

Soil hydraulic properties are functions of soil texture and structure. Soil sorptivity(Phillips, 1957), macroscopic capillary length, mean pore size, and hydraulic conductivity have been proposed as the appropriate parameters for fundamental soil hydraulic properties(Bower, 1966; Clothier and White, 1981.). Especially, hydraulic conductivity is a measure of the gravity-driven contribution to infiltration flow. Macroporous forest soils are characterized by large increases in hydraulic conductivity as the soil water pressure head approaches zero and the soil becomes saturated in continuous macropores(Beven and Germann, 1982; Warrick *et al.*, 1977).

Soil hydraulic conductivity-water content and matric head water content functions are needed to predict the factors that influence the management of soil water in forest areas. These measurement are not easily made. Moreover, laboratory and field measurement often differ from each other due to the differences in soil conditions. Adequate field evaluation may also require many measurements to reflect spatial variability.

In this paper we observed the spatial variability of the soil hydraulic properties of Japanese larch(*Larix leptolepis*) dominated forest site. The objectives were to characterize the soil hydraulic properties of this forest area and to determine the changes in this properties and the implications of this changes to soil textural and structural changes caused by natural habitat and climatic conditions.

MATERIALS AND METHODS

Sites and Soils

The experiment was conducted on three 0.5 × 0.5m plots on *L. leptolepis* dominated forest site on the experimental station of Kyunghee University, located in Kwangju, Kyunggi-Do. Table I summarize relevant information about the investigation site that the rainfall simulator was established to observe the spatial variability of hydraulic properties.

Near a small watershed three rainfall simulator plots (50 × 50cm) represented a broad range of soil horizon were chosen for the rainfall simulator experiment, but they did not represent all possible variation within the site. Soil characteristics were described in Table II.

For each plot, PVC plot borders were driven to a depth of 20cm while soil was carefully excavated from outside the plots. Then the soils were filled and tamped back in place, resulting in disturbance of only about 1cm of soil around the inside of the each plot. Plastic pipe 2.5cm in diameter were connected to the PVC plot border for letting runoff while applying water from the rainfall simulator mounted on 50cm high above the soil surface. In the middle of the plot, three tensiometers connected to Hg monometer were installed to the depths of -10cm, -20cm, and -30cm in measuring the soil matric water potential. On the other hand, three rainfall-simulation intensities were controlled using the syringe needles(5, 25, 50*G*) which produced intensities of 2.5, 5, and 10cm h^{-1}, respectively. A prewet simulation of 1cm hr^{-1} was made to sufficiently moisten the initially unsaturated soil so that the two subsequent simulations would be made on soil having similar(near field capacity) moisture conditions. After the prewet simulation, two intensity simulation were made.

Gravimetric soil water samples were collected with an auger at 9 locations within the middle

Table 1. Characteristics of *L. leptolepis* forest site to study the spatial variability of hydraulic properties.

Items	Contents
Location	37° 28′ 30″ N, 127° 21′ 00″ ~ 127° 22′ 30″ E
Soil Horizon (depth)	O Horizon (0~5cm), A Horizon (5~35cm), B Horizon (35~55cm), C Horizon (below 55cm)
DBH (cm)	18.6
Slope gradient	2~50 °
Subspecies	*Stephanandra incisa*, *Morus bombycis*, *Aralia elata*, *Rosa multiflora*
Basal area (m^2 hr^{-1})	13.8

Table 2. Characteristics of bulk densities(BD), organic matter(OM) content, and soil texture(ST) above C horizon in the *L. leptolepis* dominated forest site.

Horizon	O			A			B		
	BD[1]	OM[2]	ST[3]	BD	OM	ST	BD	OM	ST
	1.18	3.4	Sandy Loam	1.26	2.0	Sandy Loam	1.26	1.4	Sandy Loam

[1] BD(Bulk Density) : g/cm^3

[2] OM(Organic Matter) Content: dry mass vs. dry mass of soil(%)

[3] ST(Soil Structure)

plot prior simulation and at 9 locations for other two plots immediately after simulation. Soil water tension was also measured with two tensiometers *Hg* monometer throughout both rainfall intensities of 2.5, and 5cm h^{-1}. The water content at pF 2.02 and 4.2 were measured from the soil samples of the experimental plots with the pressure chamber.

Theory of General Soil Hydraulic Characteristics

The general equation of water flow through porous media, often called Richard's equation, results in

$$\frac{\partial \theta_{\nu}}{\partial t} = [Kw \frac{\partial \varphi_{h}}{\partial S_{z}}] \qquad [1]$$

in which K_w and $\frac{\partial \varphi_{h}}{\partial S_{z}}$ are the hydraulic conductivity and hydraulic gradient of the soil which are strongly dependent on θ_{ν}.(Hanks, 1992 a, b) Also, the water content for a different soil layer can be determined the differences between water content with time by Eq. [2].

$$\frac{\partial \theta_{\nu}}{\partial t} = \frac{\theta_{\nu,i+1} - \theta_{\nu,i}}{\partial t} \qquad [2]$$

In the above numerical approximation "*i*" and "*j*" subscripts refer to the depth and time increments, respectively.

For vertical water flow into soil Eq. [1] becomes

$$\frac{\partial \theta_{\nu}}{\partial t} = [Kw \frac{\partial(\varphi_{h} + \varphi_{m})}{\partial S_{z}}] \qquad [3]$$

when θ_{ν} decreases less than 1 that the pressure potential(ψ_p) disappears as soil becomes unsaturated condition. Thus, the change in K_w with change in the water content(θ_{ν}) is caused by continuous downward water flow. And the water content with time can be estimated by the hydraulic potential differences between the two points (Eq. [4]).

$$\frac{\partial \theta_{\nu}}{\partial t} = [(\varphi_{m,i-1,j} + \varphi_{m,i,j}) + \Delta Z](\frac{K_{w,i-1/2,j}}{\Delta Z^{2}}) \qquad [4]$$

However, this approximation can be a good estimate only the size of the time and depth increments are small, as well as the soil layer is homogeneous, which is the hydraulic conductivity is same through the soil depth investigated. At the layer interface the hydraulic conductivities may be different due to the hydraulic gradient, resulting in a higher hydraulic gradient s limiting factor. At large time or large infiltration, the potential gradient in the lower hydraulic gradient will approach unity. Therefore, the hydraulic conductivity should be modified to describe the real spatial water distribution. First, we must have continuity of matric potential at the interface at all times.

$$\psi_{1}(z_{c}) = \psi_{2}(z_{c}) = \psi_{c} \qquad [5]$$

Also, flux continuity implies that, for steady state flow,

$$K_{1}(\psi)\left[1 - \frac{d\psi}{dz}\right] = K_{2}(\psi_{c}) \qquad [6]$$

In Eq. [6], c refers to conditions just at the interface. However, the hydraulic conductivity(K_{eff}) for water flow through a layered soil containing n layer of thickness L_j may not be adequately described by Eq. [6]. For a multi-layered soil having different hydraulic properties, the effective hydraulic conductivity(K_{eff}), Eq. [5] suggested by Jury(1994) can be a suitable for a real hydraulic conductivity through a layered soil containing n layer of thickness L_j.

$$K_{eff}=\frac{\sum_{j=1}^{N} L_j}{\sum_{j=1}^{N}\left(\frac{L_j}{K_j}\right)} \qquad [7]]$$

In eq. [7], K_j can be substituted by eq. [6] to satisfy the continuity for the flow regime in soils. However, this effective hydraulic conductivity can be worked between the adjacent soil layers. On the other hand, the matric potential-water content [8] and hydraulic conductivity-water content [9] functional forms may be described below.

$$\theta(h)=[1+\alpha(-h)^{N}]^{-M} \qquad [8]$$

$$K(\theta)= K_s\, \theta^{1/2}[1-(1-\theta^{1/M})^{M}]^{2} \qquad [9]$$

where $\theta = (\theta - \theta_r)/(\theta_{vsat} - \theta_r)$, and M = 1 -1/N. θ_{vsat} and θ_r denote saturated volumetric water content and the residual water content.

To solve the water flow through a multi-layered soil, the initial condition can be stated as

$\theta_v = \theta_{vi}$ for $t = 0$ and $x > 0$

At the instant flow begins, the water content at the water front can be brought instantaneously to saturation by placing it in contact with water resource. Thus, the boundary condition can be stated as

$\theta_v = \theta_{vsat}$ for $t > 0$ and $x = 0$

However, the initial and the boundary condition should be modified according to the real situation in soils to be investigated.

RESULTS

After flowing of water using the rainfall simulator, we examined the water content at the depth of 5, 20, and 30cm from the soil surface. The water retained by soils containing indigenous organic matter at *p*F 2.02 were much different, while those at *p*F 4.2 decreased with increasing soil depth, probably due to the decrease in organic matter content. On the other hand, the water retention of soils after removing organic matter at the same *p*F values was significantly decreased compared to the water retention characteristics of soil with organic matter.

Especially, the water retention was increased with increasing soil depth. To account for this water retention characteristics, water can be

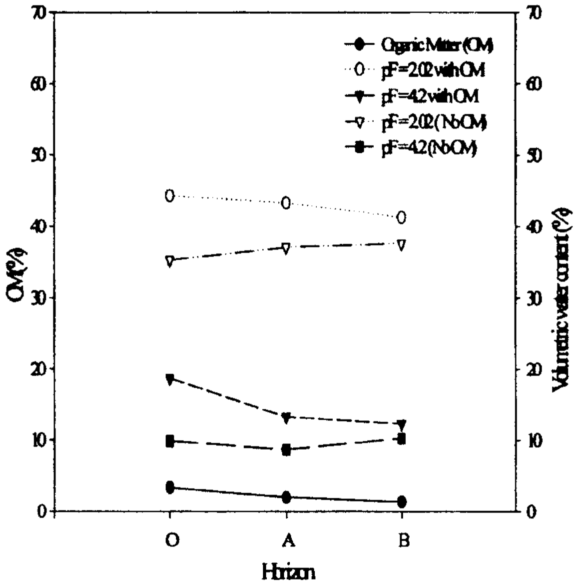

Fig. 1. Organic matter contents, volumetric water content with organic matter and without organic matter content at pF 2.02 and 4.2 on soils obtained from *L. leptolepis* site.

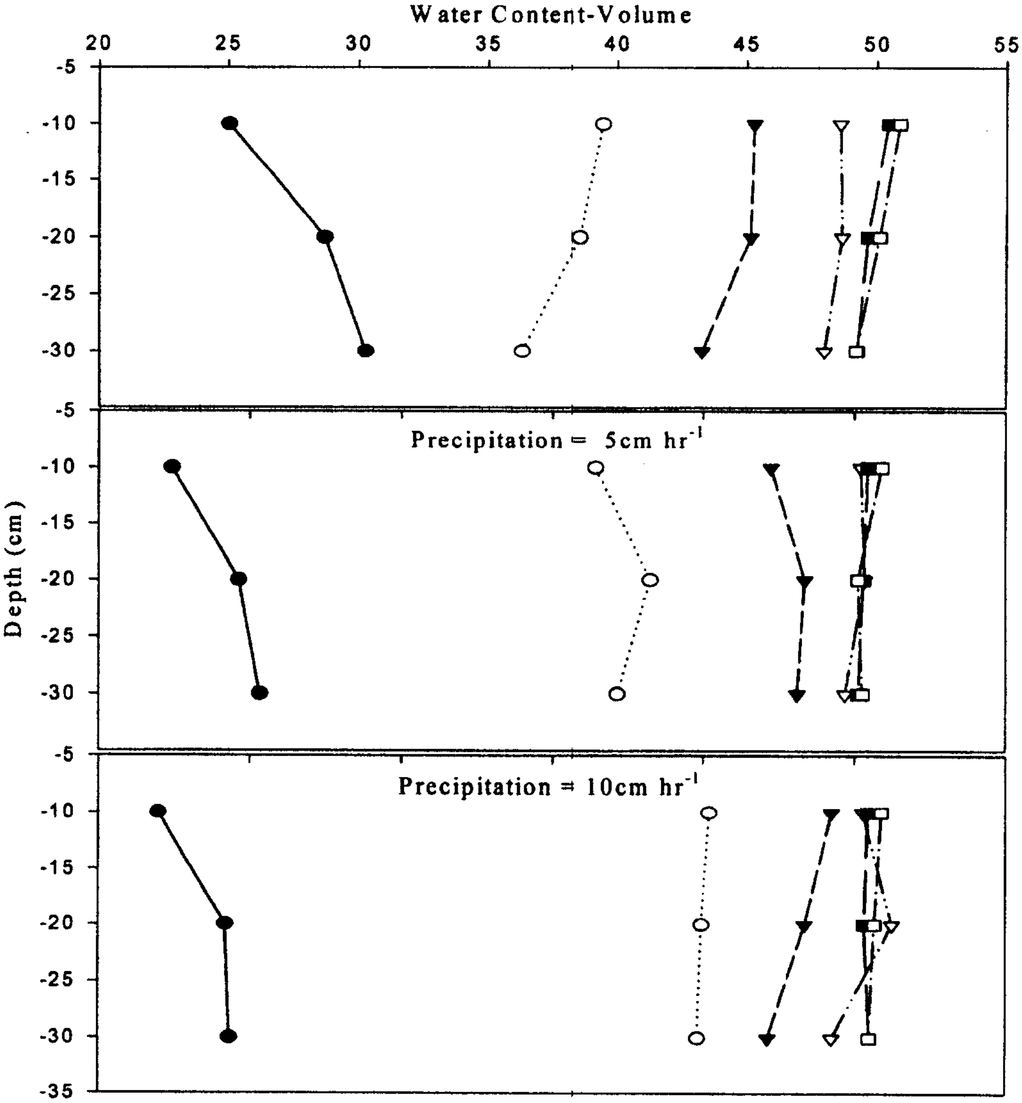

Fig. 2. Water content changes during the simulated rainfall of 2.5, 5, and 10cm hr^{-1} as a function of depth in the soil for six different times.

—●— T=0hr ···○··· T=1hr —▼— T=3hr —▽·-- T=6hr —■— T=9hr —□·-- T=12hr

retained in two ways: a constant amount by saturated portions of soil mineral silt/clay matrix and the rest by decomposed organic matter. However, the amount held by the mineral matrix would not be significantly changed as far as the portions of soil mineral matrix are not changed by elluviation ot illuviation in that soil profile(Wilson and Luxmore, 1988).

The water distribution patterns through soil profile while applying water at the rates of 2.5, 5, and 10cm hr^{-1} by rainfall simulator are shown in Fig. 2. Three soil plots were relatively unstable and therefore the soil properties varied with time. This problem was particularly pronounced at the soil surface where the impact of water drops from rainfall simulator caused aggregate breakdown, resulting in drastic changes in infiltration rate at the soil surface as

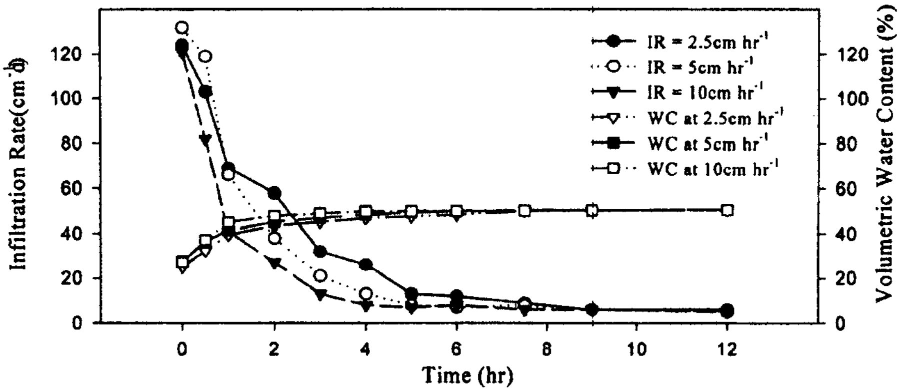

Fig. 3. Water content and infiltration rate at the soil surface with a water application rates of 2.5, 5, and 10cm hr^{-1} using the rainfall simulator.

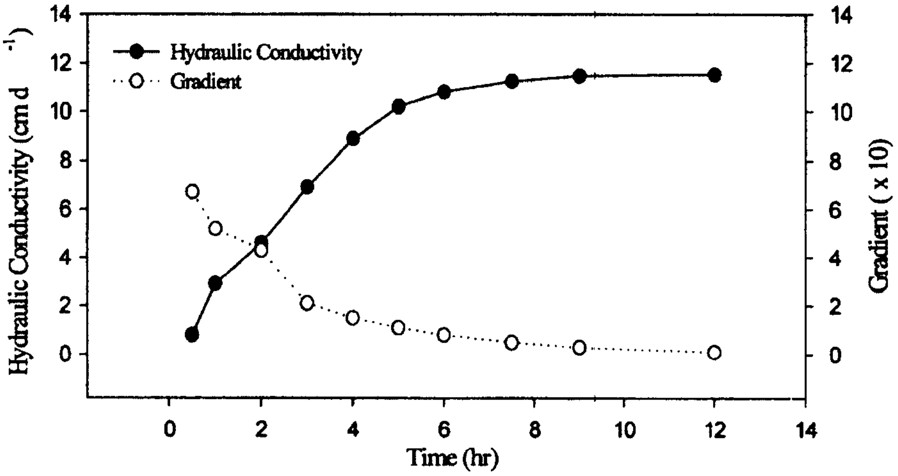

Fig. 4. Changes of saturated hydraulic conductivity and hydraulic gradient as a function of time for the infiltration situation with 5cm hr^{-1} water application.

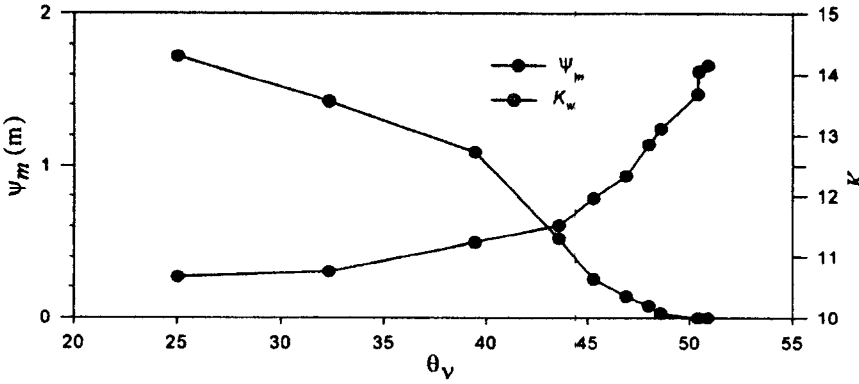

Fig. 5. Hydraulic conductivity, and matric potential as influenced by water content for top 10 cm soil depth.

the water application continued.

During redistribution of water in soil illustrated in Fig. 2, most part of the soil were initially unsaturated. The change of water content with application of water from the soil surface for three depths showed that the time required to reach the saturated condition took more time as the rainfall intensity decreased in the order of 10, 5, and 2.5cm hr^{-1}. The initially low water content rapidly built to the saturated value and remained constant thereafter. Consider what happened to the hydraulic gradient and K_w in the situation discussed above, shown in Fig. 3. In the first stage, θ_v at the soil surface increases followed by consequent increase of K_w. In the second stage the soil surface became saturated and θ_v reached to saturation but infiltration rate still decreased. This indicates that the water content in the subsurface did not reach to saturated, and different soil characteristics such as pore structure, clay content and organic matter which can influence water flow.

In Fig. 4, we observed the relationship between hydraulic conductivity and gradient as a function of time for the infiltration on condition with 5cm hr^{-1} water application. Gradient gradually decreased to 0.1 as K_w became stabilized to 11.44cm d^{-1} around 7 hours later. If infiltration remains constant and K_w increases then the hydraulic gradient ($\Delta\psi_h$ / Δz) must decrease accordingly. But unsaturated flow can occur throughout the soil profile if the rainfall rate is less than K_w, and the hydraulic gradient approaches 1 as K_w, near the soil surface, approaches the water application rate after some time has occurred. Thus θ_v and ψ_m in the initially wet soil may essentially uniform.

Comparisons of K_w, $\Delta\theta_v/\psi_m$ as related to θ_v were shown in Fig. 5. ψ_m rapidly decreased to approximately 0 as K_w reached maximum. This meant that ψ_m was influenced by increasing water content as water flew down. But this ψ_m did not account for water redistribution below the 10cm soil depth. To approximate the real ψ_m in the field condition, we need to consider the effective hydraulic conductivity throughout soil profile. One interesting feature of the infiltration rate and hydraulic conductivity is that soil texture and organic matter content is clearly expressed in the measured spatial hydraulic properties(Nielson *et al.*, 1973)

CONCLUSIONS

Measurements of the spatial hydraulic properties in a forest soils which formed by the functions of sedimentation and elluvial processes showed that the initial volumetric water content was increased with increasing soil depth as far as the portions of soil mineral matrix are not changed. The water redistribution pattern at the rainfall intensity of 2.5, 5, and 10cm day^{-1} represented that the time required to reach the saturated condition took more time as the rainfall intensity decreased. This suggests that a gradual increase in K_w close to saturation by variable rainfall intensity must not only depend on the soil physical properties including soil pore-size distribution in a wide range of soils, but also by affinity of water into clay particles and organic matter content. However, bulk density referred each soil horizon formed at each plot as an indicator of K_w was not a productive variable, due to ambiguity as an indicator of both grain size and compaction, while organic matter was significantly related to K_w. Finally, values obtained by using soil physical property variables represent that this approach may need to require a wide range of variables than those presented here for the best estimation of spatial variability of water redistribution.

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