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Scaling hydraulic properties of a macroporous soil

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Abstract. Macroporous soils exhibit significant differences in their hydraulic properties for different pore domains. Multimodal hydraulic functions may be used to describe the characteristics of multiporosity media. I investigated the usefulness of scaling to describe the spatial variability of hydraulic conductivity ($K(-h)$) functions of a macroporous soil in Las Nutrias, New Mexico. Piecewise-continuous hydraulic conductivity functions suitable for macroporous soils in conjunction with a hybrid similar media-functional normalization scaling approach were used. Results showed that gravity-dominated flow and the related hydraulic conductivity ($K(-h)$) functions of the macropore region are more readily scalable than capillary-dominated flow properties of the mesopore and micropore regions. A possible reason for this behavior is that gravity-dominated flow in the larger pores is mostly influenced by the pore diameter which remains more uniform as compared to tortuous mesopores and micropores with variable neck and body sizes along the pore length.

1. Introduction

Near-saturated hydraulic properties of soil are important for predicting water flow in macroporous soil. *Mohanty et al.* [1997] showed the presence of bimodal hydraulic conductivity functions using disc infiltrometers and multistep outflow methods with detached soil cores at a field site in Las Nutrias, New Mexico. New multimodal piecewise-continuous functions were used to describe the $K(-h)$ relationship for the macroporous soil:

$$K(h) = \sum_p k_p K_p(h) = \sum_p k_p \frac{\{1 - (\alpha_p|h|)^{n_p-1}[1 + (\alpha_p|h|)^{n_p}]^{-m_p}\}^2}{[1 + (\alpha_p|h|)^{n_p}]^{m_p/2}}$$

$$(m_p = 1 - 1/n_p) \qquad h \le h_K^* \qquad (1)$$

$$K_{np}(h) = K^* + K^*[e^{(h-h^*)\delta} - 1] \qquad h_K^* < h \le 0 \qquad (2)$$

$$K_{np}(h) = K^* + K^*[e^{(-h^*)\delta} - 1] \qquad h > 0 \qquad (3)$$

where K_p is the hydraulic conductivity of the pth capillary-dominated flow domain [L T^{-1}], K_{np} is the hydraulic conductivity for noncapillary-dominated flow domain [L T^{-1}], h is the equilibrium soil water pressure head of the bulk soil across all flow domains [L], $h_K^* \approx h_\theta^* = h^*$ is the critical or breakpoint soil water pressure head where flow changes from capillary-dominated to noncapillary-dominated flow or vice versa [L], K^* is the hydraulic conductivity corresponding to h^* [L T^{-1}], δ is a fitting parameter representing effective macroporosity or other structural features contributing to noncapillary dominated flow [L^{-1}], α_p [L^{-1}] and n_p [dimensionless] are fitting parameters [*van Genuchten*, 1980] for pth capillary-dominated flow domain, and k_p is the saturated hydraulic conductivity for capillary-dominated flow domain p subjected to $\Sigma k_p = K^*$[L T^{-1}].

Shouse and Mohanty [1998] showed the success of a hybrid scaling approach combining similar media scaling and functional normalization to coalesce a near-saturated $K(-h)$ data set measured using disc infiltrometers in an agricultural field near Ames, Iowa. In that study the saturated hydraulic conductivity (K_{sat}) at each measurement location (x) was found to be an ideal (physical) scale factor for K, while another empirical scale factor was estimated for h; all $K(-h)$ data could then be coalesced to one unimodal Gardner-type function. In this paper we will investigate the suitability of scaling schemes for bimodal-type piecewise-continuous hydraulic conductivity functions of a macroporous soil in Las Nutrias, New Mexico.

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2. Experimental Methods

The data used in this paper were taken from *Mohanty et al.* [1997]. The experimental field is a 6-ha commercial agricultural farm, situated on an alluvial flood plain of the Rio Grande, near Las Nutrias, New Mexico. The climate is arid to semiarid and has a wide range of temperatures and rainfall. The field site consists primarily of silty clay loam sediments with moderate to poor drainage properties, underlain by fine sands and with no impeding strata to a depth of about 7 m [*Natural Resource Conservation Service*, 1992]. Surface horizons, however, contained visible root channels, worm holes, and cracks thereby composing a complex network of preferential flow paths. Surface infiltration measurements at 0-, 3-, 5-, 7-, 10-, 13-, 15-, 17-, 20-, 25-, 30-, 60-, and 150-mm soil water tensions were made at nine sites in the summer of 1994, using ponded and tension infiltrometer procedures as described by *Mohanty et al.* [1994, 1997]. To minimize any site disturbance, the water-supply tower of the tension infiltrometer was refilled during the experiment using the procedure of *Ankeny* [1992]. The sites were selected such that they represent different soil series and textures across the field. *Ankeny* [1992] cautioned that infiltration for tensions below 10 mm should be interpreted carefully because of possible inaccuracies in the applied tension due to bubbling. As such, we interpreted the infiltration rate in this narrow range of tensions only in a qualitative manner so as to reveal relative trends near saturation. The data were used to develop piecewise-continuous hydraulic conductivity functions whose fitting parameters could be further adjusted during model calibration [*Mohanty et al.*, 1997]. For the

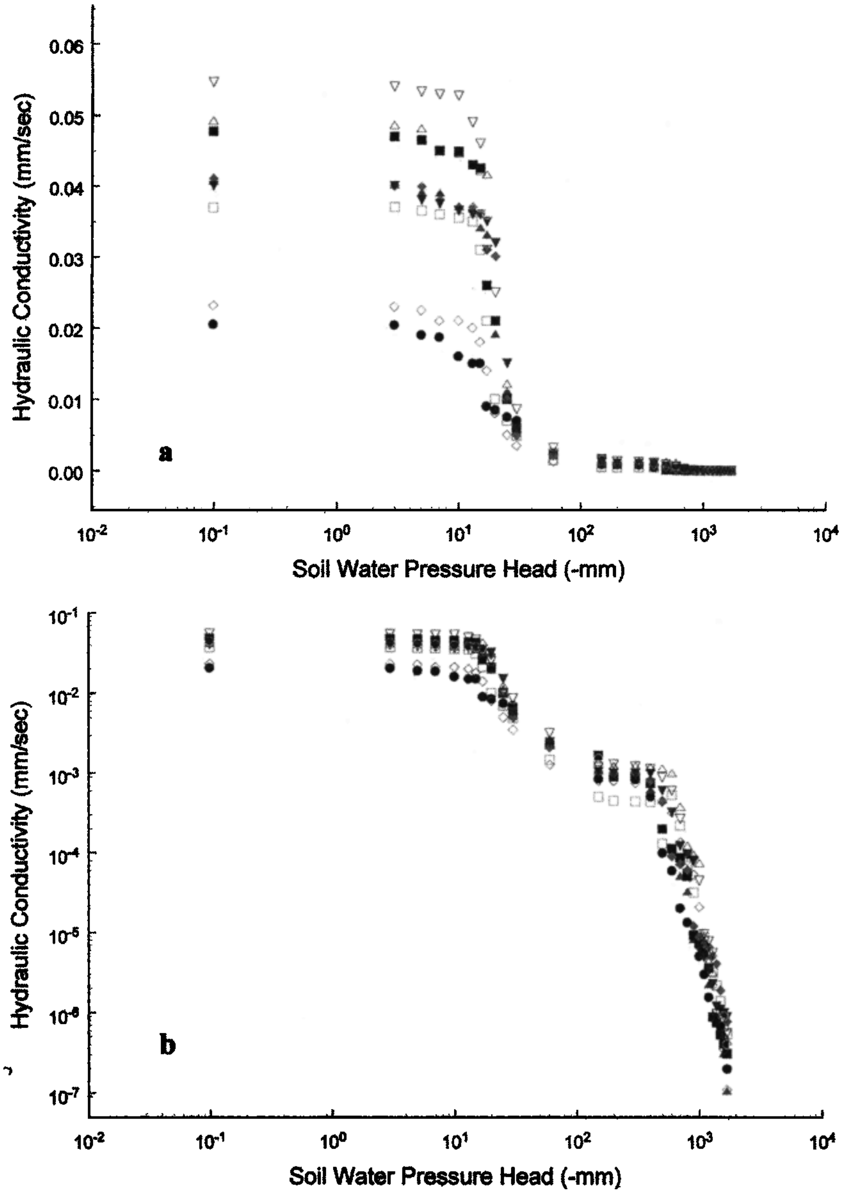

Figure 1. Raw hydraulic conductivity $K(h)$ of the Las Nutrias field site measured using a disc infiltrometer and the multistep outflow method; (a) semilog scale and (b) log-log scale.

above intensive infiltrometer measurements we used thin layers (1~2 mm) of 40 to 60-mesh contact sand (minimum K = 48.81 m d^{-1}) to minimize any interfacial effect of the sand on the infiltration rate close to saturation as reported by *Everts and Kanwar* [1993]. The actual tensions at the soil surface as reported here were estimated by assuming unit-gradient water flow through the sand layer and by reducing the tension at the base of the infiltrometer with the average sand depth.

Following the surface infiltration measurements, 50-mm-diameter and 50-mm-long undisturbed soil cores were collected directly below the disc from the same nine sites representing different soil series and textures across the field. The soil cores were subsequently used for measuring water retention and hydraulic conductivity data over a wider range (0 to 1700-mm at 100-mm increments) of soil water tensions. The laboratory method of *Richards* [1965] and the multistep outflow approach of *Gardner* [1956] were used for measuring soil water retention and hydraulic conductivity functions, respectively. Similar to our approach in the field, we were mostly interested in relative trends rather than absolute values of the retention data near saturation. While the actual numbers may be imprecise, they could be adjusted by subsequent calibration of selected parameters in the piecewise-continuous retention function so as to better represent the flow scenario at the field site [*Mohanty et al.*, 1997]. For this purpose the in situ and laboratory experiments were carried out as carefully as possible using very small increments in the tension near saturation. The $K(h)$ data from all the three methods were subsequently superimposed for the scaling analysis.

3. Scaling Analysis

Scaling methods in soil science include dimensional analysis, inspectional analysis, and functional normalization [*Tillotson and Nielsen*, 1984]. *Miller and Miller* [1956] introduced the concept of geometric similitude and similar media scaling

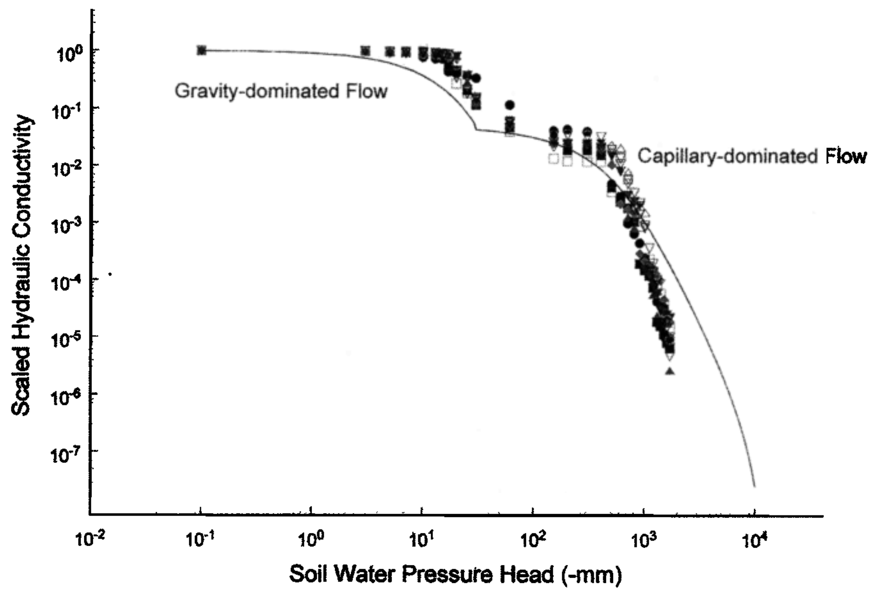

Figure 2. Scaled $K(h)$ after using K scale factors ($K_{\mathrm{sat},x}$).

based on a characteristic length scale that is related to particle size and pore dimension in a particular geometric arrangement. *Hopmans* [1987] and *Vogel et al.* [1991] compared different methods to scale soil hydraulic properties. *Jarvis and Messing* [1995] used a two-line exponential model to describe and scale the $K(-h)$ relationship of a macropore-mesopore system. More recently, *Shouse and Mohanty* [1998] scaled the near-saturated $K(h)$ using a hybrid scheme of similar media scaling and functional normalization. K_{sat} was used as a (physical) scale factor for K, while an empirical scale factor (α) was used for h. The saturated hydraulic conductivity was used to reduce gravity-dominated macropore flow variability, if present, near saturation because of soil structural features that have little or no influence on water flow under unsaturated conditions [*Mohanty et al.*, 1996, 1997]. For the near-saturated K the similar media concept is not necessarily limited to geometrical similarity as given by *Miller and Miller* [1956] but may be more of a combination of dynamic, kinematic, and geometric similarity represented by K_{sat} [*Shouse and Mohanty*, 1998]. Dynamic similarity refers to the force (e.g., gravitational, capillary, adsorptive, and viscous) relationships between different soil pore systems. Kinematic similarity represents relationships between motions (e.g., progress of a wetting front with time or depth) in different soil systems. Geometrical similarity accounts for size (e.g., particle dimensions, pore dimensions, and tortuosity) relationships between the soil systems.

At the field site, soil hydraulic conductivities measured with a disc infiltrometer between 0 and −30-mm soil water pressure head were found to be, at all measurement locations, several orders of magnitude higher than the hydraulic conductivity at −30-mm soil water pressure head thus indicating an approximate matching point between the gravity-dominated macropore flow region and the capillarity-dominated matrix flow region (Figures 1a and 1b). An arbitrarily selected (K, h) data set was used to define the reference model for the bimodal piecewise-continuous formulation, that is,

$$K(h) = K^* \frac{(1 - (\alpha|h|)^{n-1}[1 + (\alpha|h|)^n]^{-m})^2}{[1 + (\alpha|h|)^n]^{m/2}} \qquad (4)$$

$$h \le -30 \text{ mm}$$

$$K(h) = K^* + K^*[e^{(h+30)\delta} - 1] \qquad -30 \text{ mm} < h \le 0 \qquad (5)$$

$$K(h) = K^* + K^*[e^{(30)\delta} - 1] \qquad h > 0 \qquad (6)$$

where $K^* = 0.00555$ mm s^{-1}, $\alpha = 0.0015$ mm^{-1}; $n = 1.80$; $m = 0.444$; and $\delta = 0.098$ mm^{-1}. As given by *Shouse and Mohanty* [1998], $K_{\mathrm{sat},x}$ at any position x in the field was used to scale K_x for the entire tension range (0–1700 mm). This was repeated for all measurement locations. Before conducting any scaling for h, we examined the scaled K for inherent characteristics associated with the bimodal-type hydraulic conductivity functions. The translation of the raw K data (Figure 1) to relative or scaled K data (Figure 2) showed two important features: (1) K between 0 and −30-mm soil water pressure head (gravity-dominated flow region) showed better coalescing behavior as compared to K below −30-mm soil water pressure head (capillary-dominated flow region), and (2) $K_{\mathrm{sat},x}$ was a suitable and effective physical scale factor for K near saturation, consistent with observations by *Shouse and Mohanty* [1998]. In the individual flow regions (gravity-dominated flow ($h > -30$ mm) and capillary-dominated flow ($h < -30$ mm)), paired $[K_{\mathrm{rel}}(h), h]_{i=1,\cdots,N,x=1,\cdots,X}$ data points were subsequently scaled to the reference curve using functional normalization of h [*Tillotson and Nielsen*, 1984]. Following *Shouse and Mohanty* [1998], the (empirical) scale factor β_x for a certain infiltration sequence at site x consisting of N data points (i.e., tension steps) was then found by minimizing the sum-of-squared differences (SS) between the reference relative hydraulic conductivity curves and the scaled data points for each flow region (p):

$$\mathrm{SS} = \sum_{i=1}^{i=N} [\beta_x h_i - h_i^{\mathrm{ref}}]^2 \qquad (7)$$

$$\beta_{i,x} = \frac{h_{i,x}^{\mathrm{ref}}}{h_{i,x}} \qquad i = 1, \cdots, N \qquad \forall x \qquad (8)$$

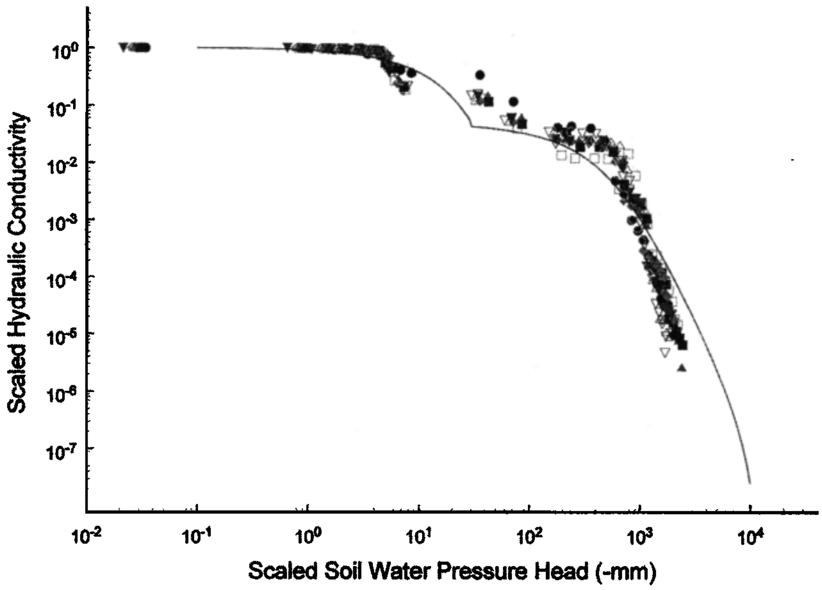

Figure 3. Scaled $K(h)$ after using both K scale factors ($K_{\text{sat},x}$) and h scale (β_x).

$$\beta_x = \frac{1}{N}\sum_{i=1}^{i=N} \beta_{i,x} \qquad \forall x \tag{9}$$

Once $\beta_{x,(p)}$ for a spatial location (x) was known, scaled $h^*_{i,x}$ ($=\beta_{x,(p)}h_{i,x}$) values were calculated for all tension steps ($i = 1, \cdots, N$) in the flow region (p). Figure 3 shows the scaled $K(-h)$ curves for the biporous soil system; the hybrid scaling approach appears quite adequate in representing the hydraulic conductivity variability of the gravity-dominated flow region ($h > -30$ mm) across the study area. However, as given by *Jury et al.* [1987], in the capillary-dominated flow region ($h < -30$ mm), further scaling of K is necessary (results not shown here).

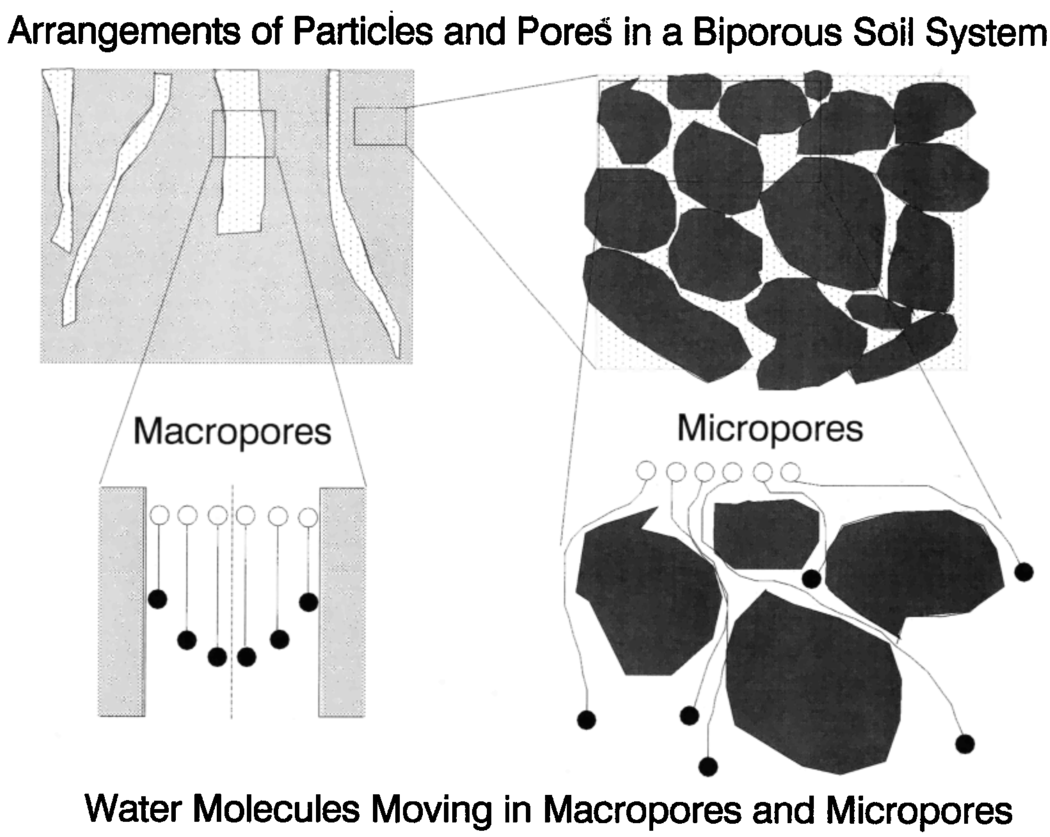

Figure 4. Typical soil particle and pore arrangements and flow paths taken by water molecules in a biporous media. Open and solid circles indicate water particles at time t_0 and t_1, respectively.

4. Discussion

Water flows through different pore groups (e.g., macropores, mesopores, and micropores) in soil. Different physical forces (e.g., gravity, capillarity, and absorption) dominate the flow process for different pore groups. For example, gravity may dominate the flow process in the larger pores (or macropores), whereas capillary forces dominate the flow process in mesopores and micropores. In this paper, pore region including mesopores and micropores is loosely defined as micropores. Because of differences in the governing forces, hydraulic conductivity increases/decreases several orders of magnitude across one or more threshold soil water pressure head(s) separating different flow regions. At the Las Nutrias field site the threshold soil water pressure head between the two different flow regions was approximately -30 mm. K in the fast flow region (0 mm $> h > -30$ mm) showed a consistent pattern at all measurement locations with a proportional effect across the field. The proportional effect could be scaled to a reference $K(-h)$ curve by using K_{sat} as the scale factor for K and an empirical h-scale factor (β). In addition to K_{sat} and β, $K(h)$ in the capillary flow region ($h < -30$ mm) requires further scaling of K to coalesce all data to the reference $K(-h)$ curve. Scaling reduced the root mean square error (as defined by *Loague and Green* [1991]) between the data (raw or scaled) and the reference model by more than 95% in the fast flow region (0 mm $> h > -30$ mm) as compared to 36% in the capillary flow region ($h \leq -30$ mm). This result indicates that gravity-dominated flow and related hydraulic functions for the macropore region are more readily scalable (i.e., no extra empirical K-scale factor of *Jury et al.* [1987] is needed) than capillary-dominated flow properties associated with the micropores. We speculate that one possible reason for this behavior is the fact that gravity-dominated flow in larger pores is mostly influenced by the pore diameter which remains more uniform compared to more tortuous micropores (Figure 4). The micropores have variable neck and body sizes that are influenced by the soil particle size and their arrangement along the pore length thus causing different water particles to move significantly different distances in time (Figure 4).

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