Soil Physics and Hydrology: Isotropy and Anisotropy

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Introduction

The directional behavior (isotropy or anisotropy) of soil hydraulic conductivity is of primary importance in analyzing rates and directions of water flow and solute transport in the vadose zone. While the hydraulic conductivity of a soil may vary from one location to another (heterogeneity), it can also vary from one direction to another at the same location (anisotropy). The directional behavior in the hydraulic conductivity of a soil governs many hydrological processes, such as infiltration, recharge, evaporation, and runoff as well as solute migration. Field soils with a large hydraulic conductivity anisotropy may result in extensive lateral spreading of fluid, solutes, and contaminants in the vadose zone. Such spreading can potentially impact groundwater recharge, irrigation scheduling, as well as plant root development and microbial activity.

In this section, we will discuss the hydraulic conductivity isotropy and anisotropy of soils under fully saturated and unsaturated conditions. Starting with Darcy's law, the control volume, and relative elementary control volume concepts, we then proceed in our discussion to causes of the isotropy and anisotropy and their relation to observation scales. Afterward, pressure head- or moisture-dependent anisotropy in unsaturated hydraulic conductivity of soils is introduced. Subsequently, the influence of anisotropy on the movement of moisture plumes in the vadose zone is discussed.

Darcy's law

Darcy's law is the fundamental principle for studying water flow in porous media. It states that the specific discharge of water in a porous medium is linearly proportional to the hydraulic head gradient and is in the direction of decreasing hydraulic head. The constant of proportionality is the hydraulic conductivity. A general form of Darcy's law for three-dimensional flow through unsaturated media can be written as

$$q_x = -K(h)\frac{\partial \phi}{\partial x}, \quad q_y = -K(h)\frac{\partial \phi}{\partial y}, \quad q_z = -K(h)\frac{\partial \phi}{\partial z} \qquad (1)$$

In the equation above, q is the specific discharge (LT^{-1}) and subscripts x, y, and z denote directions. The hydraulic head, $\phi=h+z$, has a unit of length (L). The pressure head is h (L), which is positive for saturated media and is negative for unsaturated media, and z is the elevation head (L), which is positive upward. The hydraulic conductivity (LT^{-1}) is denoted by K(h), which is the saturated hydraulic conductivity when the medium is fully saturated and is the unsaturated hydraulic conductivity when the media is partially saturated. The unsaturated hydraulic conductivity is a function of the pressure head, decreasing from the saturated hydraulic conductivity with increasing negative pressure head, or as the medium becomes less saturated.

At the pore scale, water flow in porous media takes place through a complex network of interconnected pores or openings. To describe such an intricate network in any exact mathematical manner is practically impossible. Consequently, Darcy's law considers only the average flow behavior over a certain volume of porous media, which must be greater than several pores. The volume over which the flow is averaged is defined as a *control volume* (CV). Using this CV approach, Darcy's law essentially bypasses both the microscopic level, at which we consider what happens to each fluid particle, and the pore-scale level, at which we consider the flow pattern within each pore and between pores. Our analysis of flow in porous media, based on Darcy's law, thus moves to the macroscopic level at which only average phenomena over the control volume are considered. The property defined at a point in our mathematical models therefore represents an average property over a CV and thus the property at every point in space varies smoothly such that our differential calculus applies. The medium and flow are subsequently being considered as the *Darcian continua*. This continuum concept is parallel to the continuum hypothesis in fluid mechanics and other branches of sciences.

Control Volume, Representative Control Volume, Homogeneity and heterogeneity

Using the CV concept, porous media can be classified as homogeneous or heterogeneous in terms of the hydraulic conductivity. If the hydraulic conductivity defined over a CV is constant throughout the entire porous medium in spite of the location of the CV, the medium is said to be homogeneous. Mathematically, homogeneity means that the hydraulic conductivity of the medium does not depend on the location. Under this condition, the CV is a representative elementary control volume (REV), implying the hydraulic conductivity measured over the CV in any part of the medium is representative of the entire medium. This definition of REV mandates the size of REV to be much smaller than the entire medium such that the hydraulic conductivity defined over the REV is constant regardless the location of the REV.

On the other hand, if the hydraulic conductivity defined over a CV varies from location to location, the medium is then characterized as being heterogeneous. Then, the REV does not exist for this medium. Mathematically, heterogeneity implies that the hydraulic conductivity of a medium is a function of the location.

Isotropy and Anisotropy

The hydraulic conductivity over a CV at any point in a medium can be either isotropic or anisotropic. The hydraulic conductivity isotropy implies that hydraulic conductivity values of a medium are independent of the direction. Mathematically, an isotropic hydraulic conductivity is a scalar, a quantity with magnitude but no direction. Conversely, if the hydraulic conductivity over a CV differs for different directions, the hydraulic conductivity of the CV is said to be anisotropic. For example, the hydraulic conductivity of the CV measured in the horizontal direction may be greater, or smaller, than in the vertical. In general, a porous medium may be homogeneous and nevertheless anisotropic, or it may be heterogeneous and yet isotropic at each location. Under most field conditions, soils exhibit both heterogeneity and anisotropy.
<Figure 1 near here>

To explain the causes of anisotropy in the hydraulic conductivity, consider steady state saturated flow through a series of inclined layers of homogeneous porous media (Figure 1a). Each layer has a different hydraulic conductivity value but the hydraulic conductivity of each layer is isotropic. Suppose that the upper and the lower boundary of the media are extended to infinity. The hydraulic head on the left-hand side of the medium is greater than the hydraulic head on the right-hand side, and they are maintained constant. Analysis of flow through each individual layer based on Darcy's law under the above specified conditions reveals that the specific discharge vector follows a zigzag path (solid vectors in Figure 1a) from the left to the right boundary.

Now, one visualizes the series of layers of homogeneous porous media as an equivalent homogeneous medium at a macroscopic level (Figure 1b)—homogenization-- and the average (macroscopic) hydraulic gradient is uniform in the x direction (long dashed line). The average (macroscopic) specific discharge vector (short dashed line) is no longer in the same direction as the average gradient. To produce such conditions, the hydraulic conductivity in Darcy's law for the three-dimensional homogeneous media thus becomes a hydraulic conductivity tensor with nine components. If one chooses a coordinate system in which the x axis is perpendicular to the layers and the y and z axis are parallel to the layers, the number of components of the tensor can be reduced to three. Darcy's law in three dimensions for the equivalent homogenous medium is then written as

$$q_x = -K_{xx}(h)\frac{\partial \phi}{\partial x}, \quad q_y = -K_{yy}(h)\frac{\partial \phi}{\partial y}, \quad q_{zz} = -K_{zz}(h)\frac{\partial \phi}{\partial z} \qquad (2)$$

In (2), the hydraulic conductivity is the macroscopic hydraulic conductivity--no longer the same as the local-scale hydraulic conductivity of each layer. In addition, the macroscopic hydraulic conductivity values, $K_{xx}(h)$, $K_{yy}(h)$, and $K_{zz}(h)$, in x, y, and z directions, respectively, become different, depending on the direction. In general, the macroscopic hydraulic conductivity in the direction parallel to layers is greater than the hydraulic conductivity in the direction perpendicular to the layers. That is to say, the macroscopic hydraulic conductivity for the equivalent homogeneous medium for Figure 1a is not isotropic but anisotropic. Based on the preceding discussion, the macroscopic hydraulic conductivity anisotropy of the equivalent homogeneous medium is a result of several factors. They are namely variation in K's of layers (local-scale heterogeneity), layers (structures), and most importantly, the homogenization (averaging) of the layered medium as an equivalent homogeneous medium.

Scales and Anisotropy

Anisotropy in the hydraulic conductivity varies with observation scale (i.e., the volume over which homogenization takes effect) as well as the scale of heterogeneity within the observation scale. In the following paragraphs, we will examine the hydraulic conductivity anisotropy at two observation scales, namely, a pore-scale anisotropy and a field-scale anisotropy.

<Figure 2 near here>

Pore-scale hydraulic conductivity anisotropy arises from the fact that we determine the macroscopic hydraulic conductivity of a soil over a certain volume of the soil (e.g., a soil core). Within the soil volume, one likely will find that depositional processes cause flat particles (minerals) to orient themselves with the longest dimension parallel to the plane on which they settle. This produces flow channels parallel to the bedding plane, which allows fluid flow with little resistance. Fluid flow in the direction perpendicular to the flat surface of particles however must detour and take more tortuous and longer paths than for flow parallel to the bedding plane. Therefore, under the same hydraulic gradient, more flow can occur through the soil core if the gradient is parallel to the bedding plane than for perpendicular to the bedding. The hydraulic conductivity in the direction parallel to the bedding (K_H in Figure 2) is thus greater than in the direction perpendicular to the bedding (K_V in Figure 2). The soil core thus possesses a pore-scale anisotropy in hydraulic conductivity.

<Figure 3 near here>

Field-scale hydraulic conductivity anisotropy arises from the fact that when we determine the hydraulic conductivity in a field situation, we often employ Darcy's law that assumes homogeneity of the medium over a large CV. In essence, we homogenize the soil in a large CV that likely includes numerous large-scale structural heterogeneities (such as stratification, cross-bedding, clay lenses, etc.) For example, consider a stratified medium comprised of layers of clay, silt and sand (Figure 3). Although within each layer the local-scale hydraulic conductivity may be considered homogeneous and isotropic, the macroscopic hydraulic conductivity for the equivalent homogeneous medium, an average over the conductivity values of the three layers, becomes anisotropic. In saturated media, the macroscopic hydraulic conductivity of the equivalent homogeneous medium in the direction parallel to bedding is given by a weighted arithmetic mean of hydraulic conductivity values of the layers:

$$K_H = \sum_{i=1}^{n} b_i K_i / \sum_{i=1}^{n} b_i \tag{3}$$

In (3), K_H is the macroscopic hydraulic conductivity in the direction parallel to bedding for the equivalent homogeneous medium. b_i and K_i are the thickness and the local-scale hydraulic conductivity of the i^{th} layer, respectively; the number of layers is denoted by n. On the other hand, the macroscopic hydraulic conductivity in the direction perpendicular to bedding, K_V for the equivalent homogeneous medium is given by a weighted harmonic mean of hydraulic conductivity values of the layers:

$$K_V = \sum_{i=1}^{n} b_i / \sum_{i=1}^{n} (b_i / K_i) \tag{4}$$

For soil and geological formations, where heterogeneity is not perfectly stratified, and the local-scale hydraulic conductivity exhibits a complex variation, stochastic methods have been used to derive the macroscopic hydraulic conductivity anisotropy of the equivalent homogeneous medium. Based on the stochastic analysis, the macroscopic hydraulic conductivity anisotropy depends on several physical properties of the formations. These include the variance of local-scale hydraulic conductivity (the variation in the hydraulic conductivity due to heterogeneity), and the correlation scale in different directions of the local-scale hydraulic conductivity (the average dimensions of the heterogeneity). The hydraulic conductivity anisotropy of a fully saturated medium is therefore considered as an intrinsic property of the medium. The field-scale hydraulic conductivity anisotropy is generally more significant than the pore-scale anisotropy for most geological media.

Pressure Head- or Moisture-Dependent Anisotropy

For flow through unsaturated media, the unsaturated hydraulic conductivity function also can be either isotropic or anisotropic. In contrast to the isotropy and anisotropy in saturated hydraulic conductivity, the isotropy in unsaturated hydraulic conductivity means that the unsaturated hydraulic conductivity function is the same in all directions. The unsaturated hydraulic conductivity is anisotropic, otherwise.

Few studies in the past have investigated the unsaturated hydraulic conductivity anisotropy, especially at the pore scale. The anisotropy in unsaturated hydraulic conductivity nevertheless has often been considered as being an intrinsic property of porous media, the same as the anisotropy in saturated hydraulic conductivity. Therefore, the unsaturated hydraulic conductivity anisotropy has been treated by scaling the unsaturated hydraulic conductivity in different directions with the anisotropy of the saturated hydraulic conductivity. The anisotropy of the unsaturated hydraulic conductivity thus remains constant over the full range of saturation or pressure head—a constant anisotropy.

Only recently, the anisotropy in the unsaturated hydraulic conductivity at field scales has been explored. A direct averaging approach has been used as a means to approximate the anisotropy. Following the approach for saturated flow, all local-scale unsaturated hydraulic conductivity values at a given pressure head are arithmetically averaged. This average value is then used to represent the macroscopic unsaturated hydraulic conductivity for flow parallel to bedding at that pressure head:

$$K_H(h) = \sum_{i=1}^{n} b_i K_i(h) / \sum_{i=1}^{n} b_i \quad (5)$$

In (5), $K_H(h)$ is the macroscopic unsaturated hydraulic conductivity in the direction parallel to bedding at a given pressure head, h. The parameters, b_i and K_i are the thickness and the local-scale hydraulic conductivity of the i^{th} layer, respectively. The number of layers is denoted by n. On the other hand, the macroscopic unsaturated hydraulic conductivity in the direction perpendicular to bedding, K_V, for the equivalent homogeneous medium at a given pressure head is given by

$$K_V(h) = \sum_{i=1}^{n} b_i / \sum_{i=1}^{n} [b_i / K_i(h)] \quad (6)$$

That is, all local-scale unsaturated hydraulic conductivity values at a given pressure head are harmonically weighted. This average value is then used to represent the macroscopic unsaturated hydraulic conductivity for flow perpendicular to bedding at that pressure head. The approximations by (5) and (6), however, follow the saturated flow case and do not consider the nonlinear pressure head distribution during flow in an unsaturated medium. Consequently, anisotropy based on (5) and (6) can be unrealistically large.

The macroscopic anisotropy for unsaturated hydraulic conductivity at field scales also has been derived by using a stochastic approach, which considers the nonlinear pressure head distribution in flow processes. The stochastic formulas relate the macroscopic unsaturated hydraulic conductivity to spatial statistics of parameters of the local-scale unsaturated hydraulic conductivity. For example, if the local-scale unsaturated hydraulic conductivity follows an exponential function:

$$K(h) = K_s \exp(\beta h) \quad (7)$$

where K_s is the saturated hydraulic conductivity and β is the pore-size distribution factor (L^{-1}), the macroscopic unsaturated hydraulic conductivity for flow parallel to bedding, K_H, is given as

$$K_H(h) = \exp\left[F + \frac{\sigma_f^2}{2(1+B\lambda_z)} - \left(B - \frac{(2\lambda_z - h)}{2(1+B\lambda_z)} \sigma_\beta^2 \right) h \right] \quad (8)$$

The macroscopic unsaturated hydraulic conductivity for flow perpendicular to bedding, K_v, is then expressed as

$$K_V(h) = \exp\left[F - \frac{\sigma_f^2}{2(1+B\lambda_z)} - \left(B - \frac{(2\lambda_z + h)}{2(1+B\lambda_z)} \sigma_\beta^2 \right) h \right] \quad (9)$$

In (8) and (9), F is the mean value of $\ln K_s$ (the natural logarithm of K_s), B is the mean value of β, σ_f^2 is the variance of $\ln K_s$, σ_β^2 is the variance of β, and h is the mean pressure head. The formulation in (8) and (9) assumes that the flow domain is perfectly stratified (λ_z = the average vertical thickness of the stratification (L), or the vertical correlation scale) and $\ln K_s$ and β are uncorrelated. A ratio of (8) to (9) gives the anisotropy of the macroscopic level unsaturated hydraulic conductivity:

$$\frac{K_H(h)}{K_V(h)} = \exp\left[\frac{\sigma_f^2 + \sigma_\beta^2 h^2}{(1 + B\lambda_z)}\right] \tag{10}$$

Equation (10) suggests a pressure head- or moisture-dependent anisotropy for unsaturated media. That is, the anisotropy for unsaturated K may increase with decreasing moisture content or with increasing pressure head. It may also decrease from the ratio at saturation to a ratio of one (isotropy) and then increase at lower degrees of saturation if K_s and β are correlated. Also shown in Equation (10) is that several factors control the macroscopic unsaturated hydraulic conductivity anisotropy. The factors are the mean, variance, and correlation scale of the saturated hydraulic conductivity and the pore-size distribution factor of the unsaturated hydraulic conductivity at local scales. More importantly, Equation (10) depicts the dependence of the anisotropy on the average pressure head or moisture content over the unsaturated media. Similar to hysteresis in moisture retention curves, the moisture dependent anisotropy may be subject to hysteresis effects.

<Figure 4 near here>

While the direct averaging approach is strictly applicable to saturated media, it can be used to explain the cause of the moisture-dependent anisotropy. Consider the following three cases (Figure 4). For Case I, a formation consists of many layers of porous media, each of which the local unsaturated hydraulic conductivity has different K_s but the same β. In Case II, each layer has a different K_s and β value, but the value of K_s and β in each layer are positively correlated. Case III considers each layer has a different K_s and β value, but the K_s and β values are independent. Now, take an arithmetic average of all local unsaturated hydraulic conductivity values at a given pressure head to yield the macroscopic unsaturated hydraulic conductivity for flow parallel to bedding at that given pressure head. In contrast, a harmonic average is used to yield the macroscopic unsaturated hydraulic conductivity for for flow perpendicular to bedding at that the given pressure head.

<Figure 5 near here>

Recognizing that a harmonic average weights heavily on the smallest K(h), one can plot the anisotropy ratio as a function of the pressure head for the three cases (Figure 5). Again, the anisotropy ratio is defined as the ratio of the macroscopic unsaturated hydraulic conductivity for flow parallel to bedding to the conductivity for flow perpendicular to bedding. Because the variation in K(h) remains the same over the full range of the pressure head in Case I, the anisotropy remains constant. In Case II, the variation in K(h) decreases toward a cross-over point and then increases with the pressure head. The corresponding anisotropy ratio first decreases, becomes unity (isotropic) at the cross-over point, and then increases again. In Case III, on the other hand, the variation in K(h) always increases with the pressure and in turn, the anisotropy grows with the pressure head.

While the moisture-dependent anisotropy has been derived for field-scale soils, similar moisture-dependent anisotropy at the pore scale also has been observed in the relative permeability of oil and gas measurements on sandstone cores. The sandstone, when dry or fully saturated, appeared to be homogeneous and isotropic. When the material was partially desaturated, however, thin and regular spaced strata became apparent. The air pemeability of the dry core was about twice as great parallel to the bedding planes as perpendicular to the bedding planes. Evidently, the material was quite uniform, but it was not isotropic. The effect of the anisotropy was to increase greatly the critical gas saturation and make the oil relative permeability curve steeper when flow was across the bedding planes.

<Figure 6 near here>

Influences of Anisotropy on Movement of Moisture Plumes in Soils

To illustrate impacts of the unsaturated hydraulic conductivity anisotropy on water movement in unsaturated soils, the moisture content distributions after infiltration from a surface source in three different soils were simulated. Figure 6a shows the simulated moisture content distribution in a homogeneous soil with isotropic unsaturated hydraulic conductivity. In this homogeneous and isotropic soil, water moves predominantly in the direction of the gravity. The moisture content distribution in Figure 6b corresponds to the simulated result for the same surface infiltration event in a homogeneous soil profile with a constant anisotropy in unsaturated hydraulic conductivity. Illustrated in Figure 6c is the simulated moisture content distribution in a homogeneous soil with a moisture-dependent anisotropy in unsaturated hydraulic conductivity. The moisture content distributions in both Figures 6b and c show significant lateral movement of water. The anisotropy in unsaturated hydraulic conductivity in the two anisotropic soils evidently prohibits vertical movement of water but enhances its lateral spreading. However, greater lateral spreading of water in the soil profile with the moisture-dependent anisotropy is evident than in the soil with the constant anisotropy.

<Figure 7 near here>

Effects of the anisotropy in unsaturated hydraulic conductivity also have been observed in numerous field experiments. For example, Figure 7 is a snapshot of moisture and tracer plumes in an experiment conducted in Rio Bravo alluvial deposits in Albuquerque, New Mexico. The media, according to the photo, is highly stratified, consisting of thin layers of materials of different textures. Moisture and tracer plumes spread great lateral distances and do not follow the prediction relying on a homogeneous isotropic concept—a manifestation of macroscopic anisotropy in unsaturated hydraulic conductivity. Similar phenomena were reported in field observations of pollutant migration from a septic tank drain field in glacial outwash deposits in the Spokane Valley, Washington. At the Hanford site near southeastern Washington, extensive lateral water movement but limited vertical movement was also observed in high-level radioactive waste leakage in the unsaturated zone. Field observations in the loess-loam area north of Beer-Sheba, Israel showed that rain infiltrated the soil to a limited depth with no net recharge of the groundwater—an effect of lateral flow in the unsaturated soil.

More recently, detailed soil-water tracer experiments in the shallow soils of a sandy hillslope at an arid site near Socorro, New Mexico indicated that significant horizontal downslope flow components developed despite the presence of a near vertical downward hydraulic gradient. This suggested that, at in situ pressure heads, the dune sands behaved as a highly anisotropic medium although field and laboratory permeability analysis of the sands showed that the sands were nearly isotropic at complete saturation. The field and laboratory measurements have provided strong supporting evidence for the concept that, for a layered heterogeneous porous media, the macroscopic media anisotropy varies as the state (pressure, moisture, and saturation) of the media varies.

Finally, while the anisotropy is important to vadose zone processes, its effects on surface hydrology processes also have been reported. The anisotropy of unsaturated soils and the slope of the land surface can produce strong lateral flow components, which, in turn, cause moisture accumulation in concave parts of the landscape leading to the point of saturation. The concentration of water in concave areas explained some rainfall-runoff and erosion phenomena that were previously unexplained by classical concepts of infiltration.

Further Reading

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Figure Captions

Figure 1. Schematic illustrations of flow through a layered soil and an equivalent homogeneous and anisotropic soil.
Figure 2. An illustration of the cause of the pore-scale hydraulic conductivity anisotropy.
Figure 3. An illustration of the field-scale hydraulic conductivity anisotropy, caused by strata of different materials.
Figure 4. Local-scale unsaturated hydraulic conductivity functions in layered soils for the three cases.
Figure 5. The corresponding anisotropy in unsaturated hydraulic conductivity for the three cases shown in Figure 4.
Figure 6. Simulated moisture content contours in soil profiles after infiltration from a surface source: a) a homogeneous and isotropic soil, b) a homogeneous soil with a constant anisotropy, and c) a homogeneous soil with a moisture-dependent anisotropy.
Figure 7. A snapshot of the moisture (dark area) and tracer (red and blue) plumes in Rio Bravo deposit in Albuquerque, New Mexico, U.S.A.

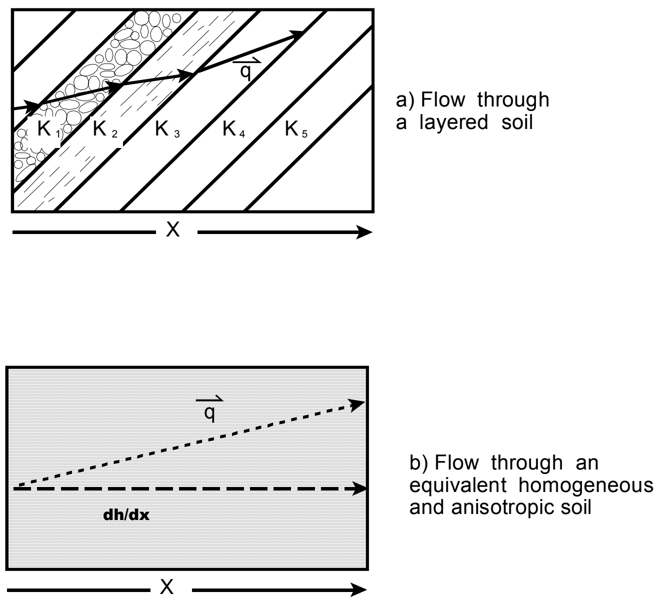

q
K 1
K 2
K 3
K 4
K 5
X
a) Flow through
a layered soil
q
dh/dx
X
b) Flow through an
equivalent homogeneous
and anisotropic soil

Figure 1.

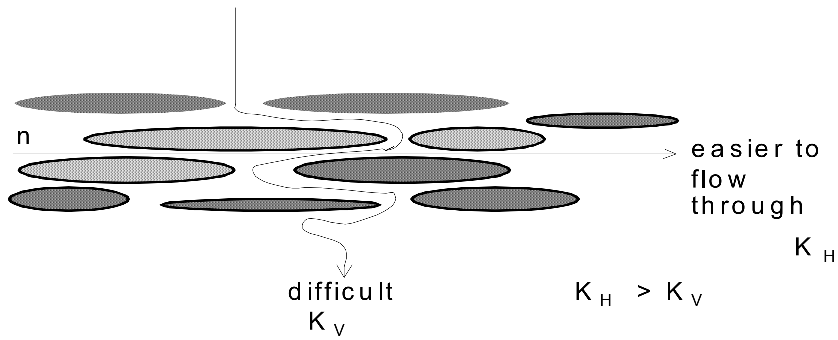


Figure 2.

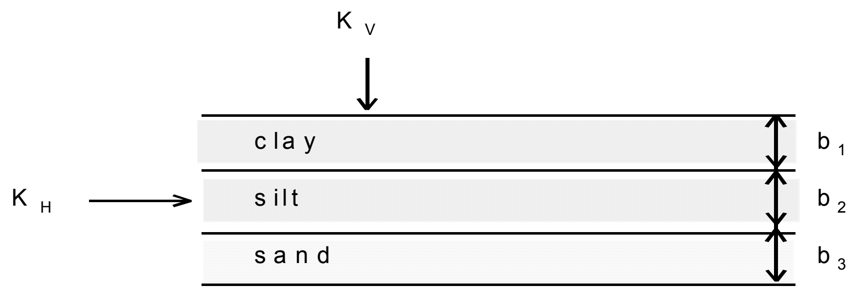


Figure 3.

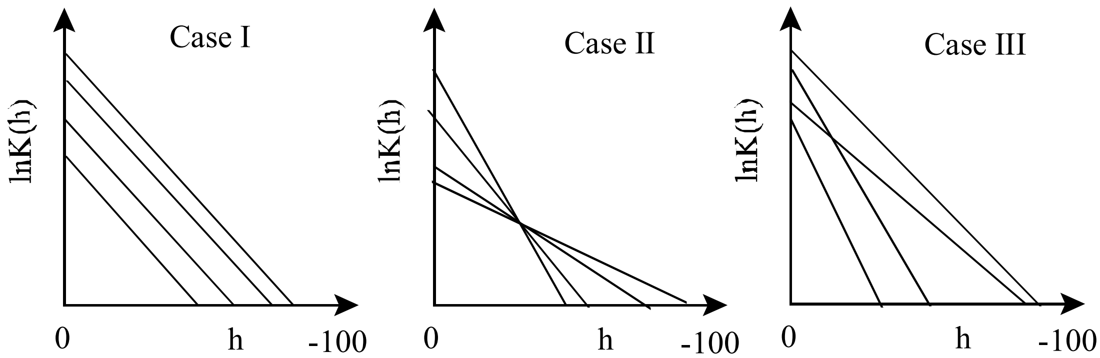
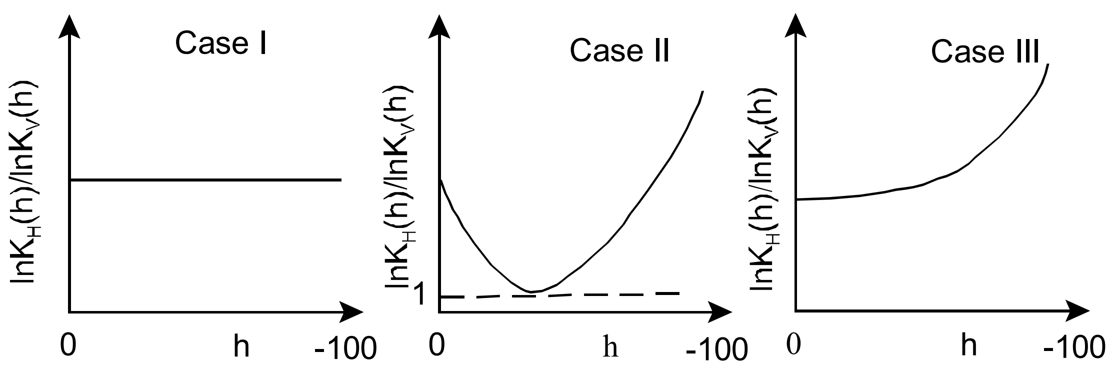
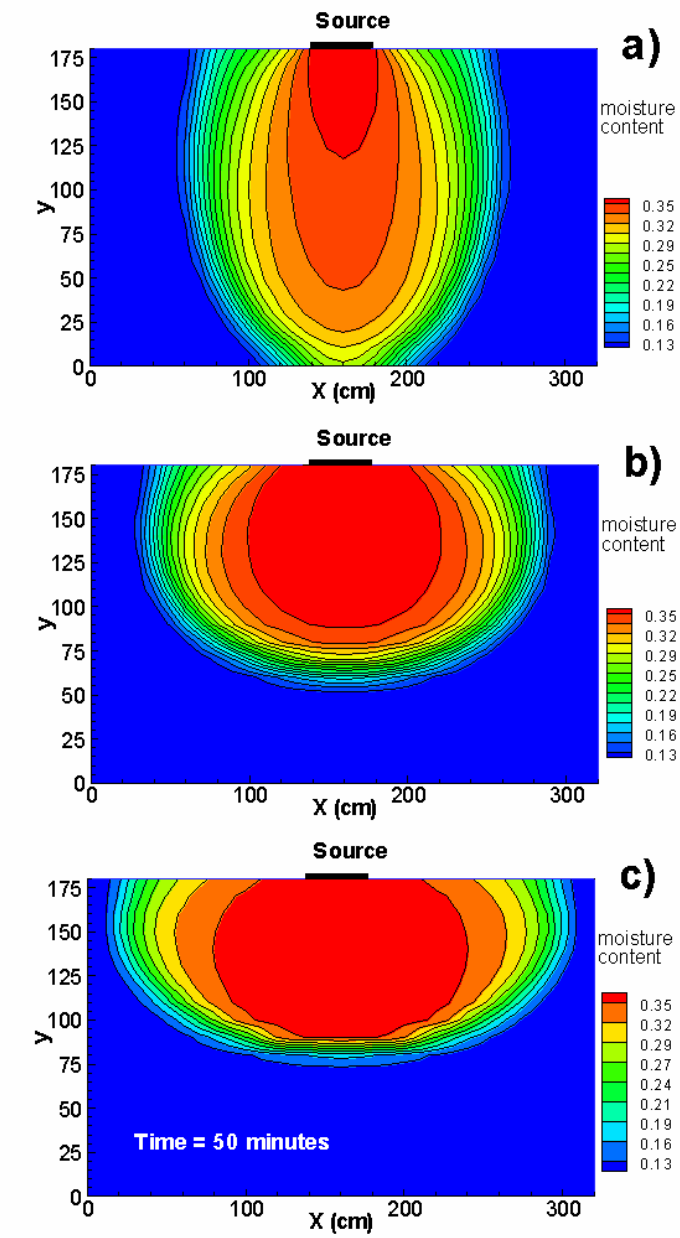


Figure 4.



Case I
$lnK_H(h)/lnK_V(h)$
0
h
-100
Case II
$lnK_H(h)/lnK_V(h)$
1
0
h
-100
Case III
$lnK_H(h)/lnK_V(h)$
0
h
-100

Figure 5.



Source
a)
moisture content
0.35
0.32
0.29
0.25
0.22
0.19
0.16
0.13
y
X (cm)
Source
b)
moisture content
0.35
0.32
0.29
0.25
0.22
0.19
0.16
0.13
y
X (cm)
Source
c)
moisture content
0.35
0.32
0.29
0.27
0.24
0.21
0.19
0.16
0.13
Time = 50 minutes
y
X (cm)

Figure 6.

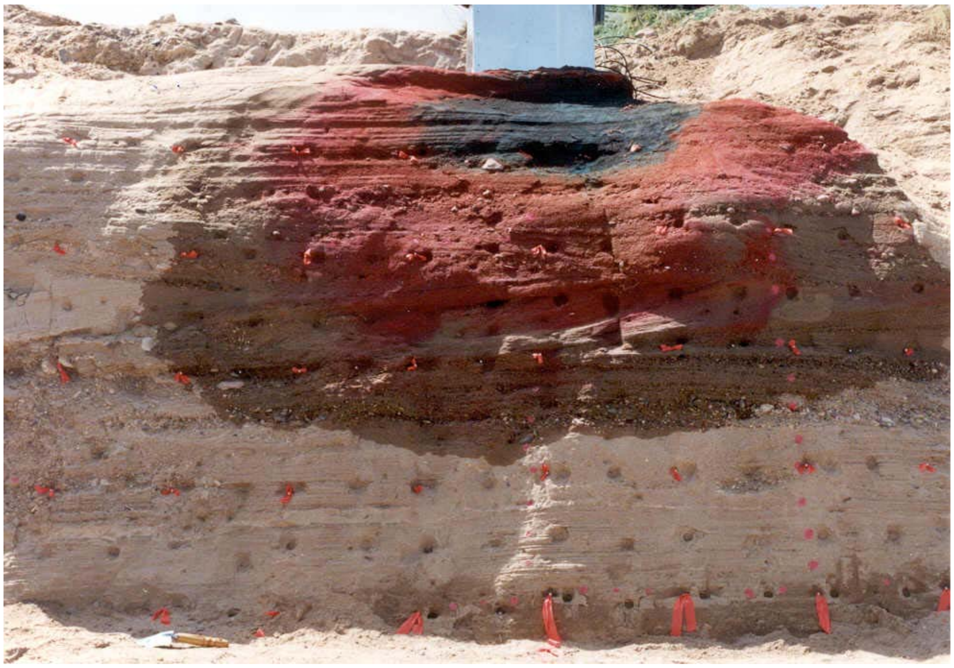


Figure 7.

Keywords.

Hydraulic conductivity, Unsaturated hydraulic conductivity, Isotropy, Anisotropy, Moisture-dependent anisotropy, Pressure-head dependent anisotropy, stochastic methods, and lateral flow.

Nomenclature.

K(h) the hydraulic conductivity (LT^{-1}).
q_x the specific discharge in the x direction (LT^{-1}).
q_y the specific discharge in the y direction (LT^{-1}).
q_z the specific discharge in the z direction (LT^{-1}).
ϕ the hydraulic head (L).
h the pressure head (L).
z the elevation head (L).
K_{xx} the hydraulic conductivity tensor component in the x direction (LT^{-1}).
K_{yy} the hydraulic conductivity tensor component in the y direction (LT^{-1}).
K_{zz} the hydraulic conductivity tensor component in the z direction (LT^{-1}).
K_H the hydraulic conductivity for flow parallel to bedding (LT^{-1}).
K_v the hydraulic conductivity for flow perpendicular to bedding (LT^{-1}).
b_I the thickness of the i^{th} layer (LT^{-1}).
K_i the hydraulic conductivity of i^{th} layer (LT^{-1}).
n the number of layers.
K_s the saturated hydraulic conductivity (LT^{-1}).
β the pore-size distribution parameter (L^{-1}).
F the mean value of $\ln K_s$.
B the mean value of β.
$\ln K_s$ the natural logarithm of K_s.
σ_f^2 the variance of $\ln K_s$.
σ_β^2 the variance of β (L^{-2}).
λ_z the vertical correlation scale, average vertical thickness of the stratification (L).