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Measuring Hydraulic Conductivity for Use in Soil Survey

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Measuring Hydraulic Conductivity for Use in Soil Survey

Soil Survey Investigations Report No. 38

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Introduction

Quantitative prediction of soil moisture movement is increasingly important for soil survey interpretations. Modern soil physics theories and simulation procedures allow predictions of, for example, the infiltration rate into the soil, redistribution of water after infiltration, and capillary rise of water from the water table. The information needed for these predictions includes hydraulic conductivity, moisture retention curves, and definition of physical boundary conditions. Unfortunately, theories and procedures may not always apply to natural soils with cracks and root or worm channels. Also, theories and simulation procedures have developed faster than have procedures for measuring hydraulic conductivity in natural soils.

Application of current soil physics theory to the soil survey program, therefore, requires more than just simple adaptation of existing theory and procedures. It is also necessary to simplify physics theory for applied field work and to establish to which soils current theory can be applied and what can be done to overcome problems in the remaining soils. Water movement, therefore, is discussed here in terms of flow either through voids in the soil matrix (matrix flow) or through voids that are not part of the soil matrix because they are significantly larger than voids that result from the packing of individual soil particles (nonmatrix flow).

Many studies of water movement deal exclusively with the *volume* of water that can be conducted by the soil and do not include the *patterns* of water movement within soil horizons that may be important in studying purification of wastes, leaching of fertilizers, etc. For two reasons it seems advisable to integrate these two aspects in the soil survey program. First, both aspects are important for soil survey interpretations. Second, patterns of movement can best be predicted from soil morphology because processes of soil genesis are highly influenced by these patterns.

Finally, application of current soil physics theory to large geographical areas within the context of soil survey programs requires many measurements, particularly of hydraulic conductivity. Therefore, existing methods for measuring hydraulic conductivity are reviewed and evaluated.

Measuring Hydraulic Conductivity

Water in soil flows in response to differences in energy. The energy content of a quantity of soil water at a given temperature can be expressed as the sum of two types: gravitational and pressure (or matric). These are generally expressed as potentials, i.e., an amount of energy is potentially available to perform work on the system. Work has the same units as energy. If, for example, water moves from a high potential to a low potential (from a state of high energy to one of low energy) it performs work on the system. Thus, the water gives off energy or uses some of its potential. On the other hand, for water to move from a state of low energy to one of high energy, work must be performed on the water, i.e., energy must be added from the system to the water. An example of this would be water moving from a warm area to a cool area because of the energy added in the form of heat.

Gravitational potential is a result of position. The higher the object (quantity of water) is from a reference plane, the more potential energy it has. Pressure potential is a result of matric forces acting on water in the soil. Visualize the soil as a bundle of different size capillary tubes. Water moves higher in the smaller tubes because of adhesive and cohesive forces. Thus it takes more energy to remove a given quantity of water from small tubes (small pores) than from larger ones. This is why large pores in soils drain first.

The potential energy of a unit volume of water is expressed in units of joules per cubic meter (J/m^3). Since energy is a force with units of newtons (N) applied through a distance (i.e., 1 J is a N·m), J/m^3 can be written as $N{\cdot}m/m^3$ or N/m^2 or pascal (Pa), which is force per area or pressure. This explains the use of the term pressure potential; i.e., units of energy per unit volume are equivalent to pressure units. In other words, pressure is one form of energy.

Strictly speaking, gravitational potential (in units of J/m^3 or Pa) is a distance (z) above a reference plane, times the mass or density (ρ) of the object (water in soils) times the acceleration due to gravity (g). Likewise, pressure potential can be expressed as an equivalent height (h) of a column of water, times its density (ρ) times gravity (g). For convenience, these two potentials can be divided by the density of water and the acceleration due to gravity and then expressed as distances z and h, respectively. Thus, an expression in terms of energy per unit weight is obtained. The hydraulic potential (H) is the sum of z and h in units of meters.

Soil water moves proportionately to the energy gradient:

$$\frac{Q}{At} \propto \frac{\Delta E}{\Delta L} \qquad [1]$$

where ΔE is the change in energy (J/m³ or Pa) over the distance ΔL(m), Q is the quantity of water (m³) moving across a given cross-sectional area A (m²) during a time interval t in seconds (s). Thus Q/At has units of velocity (m/s) and is sometimes replaced by v.

We have seen that the energy E can be replaced by the hydraulic potential H. To make the left side of equation [1] equal to the right side, we introduce a factor $-K$ (the minus sign keeps K positive), which we will call the hydraulic conductivity. The units of K are meters per second if the potentials are expressed in meters (for explanation of units, see definition of hydraulic conductivity in the glossary). Replacing Q/At with v, inserting a proportionality factor $-K$, and replacing E with H, we have

$$v = -K \frac{\Delta H}{\Delta L} \qquad [2]$$

or, since $\Delta H = \Delta h + \Delta z$,

$$v = -K \frac{\Delta h + \Delta z}{\Delta L} \qquad [3]$$

For vertical flow, $\Delta L = \Delta z$ thus

$$v = -K \frac{\Delta h + \Delta z}{\Delta z} \qquad [4]$$

or

$$v = -K\left(\frac{\Delta h}{\Delta z} + 1\right) \qquad [5]$$

Equations [2] through [5] are common forms of Darcy's law, the basic equation describing both saturated and unsaturated water movement in soils. Saturated flow is a special case of unsaturated flow and occurs when the pressure potential is zero or positive. The total pore space does not have to be filled with water for saturated flow to occur. Commonly, air-filled porosity in soils at $h = 0$ is about 5 percent. If the soil remains saturated for long periods, air-filled porosity approaches zero.

Darcy's law is the basis for all hydraulic conductivity measurements. Since heat is a form of energy and can contribute to the total energy required to move water, all hydraulic conductivity measurements are made with the assumption that the temperature is constant. Efforts are generally made to validate this assumption.

Methods for measuring hydraulic conductivity can be divided into two groups: steady state and nonsteady state. In the steady-state measurements the flux Q/At and the gradient $\Delta H/\Delta L$ are constants. The gradient is usually made equal to 1 so that the hydraulic conductivity K is equal to the flux. In the nonsteady-state measurements both the flux and gradient vary with time, and must be monitored. Nonsteady-state methods are generally faster and considered to be less accurate.

Saturated Hydraulic Conductivity Measurements

Method: Double tube.
Measurement: Saturated hydraulic conductivity.
Range: $h = 0$ only.
Precision: Poor because of visual limitations while reading water levels in standpipes, use of calibrating graphs, and small sample size.
Accuracy: Fair, because of uncertainty in defining flow direction and the small soil volume tested.
Procedure: Two concentric 1-m-long tubes are placed in the soil to a given depth, and water flow is manipulated to move from the inner tube into the outer tube at a high and changing hydraulic head. Readings are transformed to saturated hydraulic conductivity values by using tables and graphs.
Time: At least one-half day per measurement.
Cost: About $650 for double-tube apparatus. As much as 1.2 m³ of water are needed per measurement.
Advantages: Relatively fast, *in situ* method.
Disadvantages: The calculation procedures may vary. The value measured, because of undefined boundary conditions, is a mixture of vertical and horizontal hydraulic conductivity. Results are unreliable if there are vertically continuous large voids. Requires considerable physical labor.
Reference: Bouwer, H. 1962.

Method: Auger hole.
Measurement: Saturated hydraulic conductivity below a water table.
Range: $h = 0$ only.
Precision: Fair. Good if piezometer method is used because the flow system is better defined.
Accuracy: Fair because hydraulic conductivity of all soil horizons below the water table is integrated. Good if piezometer method is used.
Procedure: A hole is made to extend below the water table. The sides of the hole must not be smeared. Water is removed from the hole, and the resulting rate of water flow into the hole is

determined from the rate of water rise in the hole. Hydraulic conductivity is calculated from the rate of water rise and the geometry of the system.
Time: Rapid, usually only a few minutes.
Cost: Inexpensive. Equipment includes an auger, measuring stick, and hand pump or bail to remove water. Labor requirement is low because of the short time required.
Advantages: Fast, inexpensive, *in situ* method using a minimum of equipment.
Disadvantages: Smearing the soil could result in erroneously low values in heavy soils. Restricted to horizons below a water table. The value measured, because of undefined boundary conditions, is a mixture of vertical and horizontal hydraulic conductivity.
Comments: Hole can be lined (piezometer method) to restrict measurement to a single horizon. There are many variations of this basic method, some of which use multiple holes in monitoring the drop and rise in the water table as water is pumped out of and into different holes.
References: Van Bavel and Kirkham, 1948; Luthin and Kirkham, 1949.

Method: Column.
Measurement: Saturated hydraulic conductivity.
Range: $h = 0$ only.
Precision: Good because of large sample size and use of buret with mariotte device.
Accuracy: Good because of large sample size, minimum of disturbance, and simplicity.
Procedure: A large column of soil 30 cm high and 30 cm in diameter is carved out *in situ* at the required depth. A ring infiltrometer is placed on top and the column is encased in gypsum and removed from the pit. The steady infiltration rate is measured using a buret with a mariotte device that maintains a very shallow head on the freshly exposed upper surface of the column. The encased column is, in fact, a very large soil core. Measurement should be made only in naturally wet soil.
Time: Two to three measurements per day are possible.
Cost: Inexpensive. Equipment consists of a simple infiltrometer (less than $100), gypsum, and a buret (constant head device). A source of water is required. Low labor requirement.
Advantages: Direct test that immediately yields saturated hydraulic conductivity values. Relatively inexpensive field method. Carving out the column causes little disturbance. Method integrates soil structural properties because of large sample size.
Disadvantages: Does not work in noncoherent soil. Initially the soil must be wet or very moist.
Reference: Bouma, 1977, p. 50.

Method: Percolation test.
Measurement: Percolation rate.
Range: $h = 0$ only.
Precision: Low because of unknown gradients and flow path.
Accuracy: Poor in structured soil. Better in sandy soil.
Procedure: A hole is dug or augered into the soil to the depth required. A few centimeters of gravel is placed in the hole and water is ponded to a height of 30 cm for at least 4 hours to saturate the soil. In heavy soils measurements are made the next day. In sandy soils measurements are made immediately after the 4-hour saturation period. The vertical velocity of the water is determined and reported in minutes per inch or correctly, but rarely, in seconds/meter. The value measured after 3 hours is considered representative.
Time: Rapid, a few hours per test. Usually three holes are needed per test. If presoaking is required, the test takes 2 days.
Cost: Very inexpensive. Equipment includes soil auger, measuring tape, and gravel. A source of water is required.
Advantages: Results relate in a correlative manner to a large body of information on performance of subsurface septic seepage systems.
Disadvantages: Infiltration is measured in a small hole from which three-dimensional infiltration occurs. Physically undefined test because of unknown lateral and vertical gradients.
Comments: The percolation rate from this test should not be used for sizing filter fields. Correlation with system performance at numerous sites has been used empirically to define critical values. A method has been published that can be used to estimate saturated hydraulic conductivity from percolation test data (Kessler and Oosterbaan, 1974). The method may present problems in structured soils and is not discussed here because of the lack of practical field results.
Reference: U. S. Dept. HEW, 1969.

Method: Laboratory core.
Measurement: Saturated hydraulic conductivity.
Range: $h = 0$ only.
Precision: Poor to fair because of small sample.
Accuracy: Fair. Pores or channels that are not continuous in the field may be continuous in this small sample, thus inflating hydraulic conductivity values.

Procedures: A constant head of water is maintained over a saturated soil core to establish a steady-state rate of flow through the soil. The gradient is equal to the length of the soil core plus the height of water above it, divided by the length of the core. Saturated hydraulic conductivity is calculated from the flux and gradient.
Time: Relatively short. A few hours to a few days, depending on the soil.
Cost: Very simple and inexpensive equipment consisting of sample rings and container that can maintain a constant head of water. Labor cost is very small. Only a few minutes are needed to start the procedure and to collect readings.
Advantages: Controlled laboratory conditions. Many samples can be run simultaneously. Inexpensive, rapid, simple.
Disadvantages: Hydraulic properties of soil may change because of handling or of running water through the soil to establish steady-state conditions. It may be difficult to collect samples in soils containing coarse fragments. Many samples are required because the small sample displays variability. The average may not be representative of the hydraulic conductivity in certain soils regardless of the number of samples.
Comments: In one variation of this method, a falling head is used rather than a constant head. This then becomes a nonsteady-state method.
References: Uhland, 1949; Klute, 1965.

Unsaturated Hydraulic Conductivity Measurements

Method: Crust test.
Measurement: Unsaturated hydraulic conductivity as a function of pressure potential.
Range: Approximately −0.5 kPa to −4 kPa depending on the type of soil.
Precision: Good because of direct measurement of flux and pressure potential using pressure transducers and mariotte devices.
Accuracy: Good because samples are large.
Procedure: A large column of soil 30 cm in diameter and 30 cm high is carved out *in situ* at the required depth. A ring infiltrometer is placed on top of the soil and the column is wrapped with aluminum foil. If a measurement of saturated hydraulic conductivity is to be made later (column method), the column is encased in gypsum. The infiltration rate is measured through gypsum-sand crusts that are applied successively to the top of the column, starting with the crust of highest resistance. The flux is equal to hydraulic conductivity at the measured subcrust pressure potential because the gradient is generally 1.
Time: One series of tests per day for fluxes greater than 60 nm/s.
Cost: Inexpensive. Equipment includes an infiltrometer (custom made for less than $100), buret, and gypsum. A source of water is required.
Advantages: Direct test immediately yielding unsaturated hydraulic conductivity values. Inexpensive field method. Carving out the column minimizes disturbance.
Disadvantages: Time consuming, especially at low flux corresponding to pressure potential lower than about −4 kPa.
Comments: Saturated hydraulic conductivity can be measured if no crust is applied.
Reference: Bouma, 1977, p. 47.

Method: Instantaneous profile.
Measurement: Unsaturated hydraulic conductivity as a function of water content or pressure potential. Moisture retention curves between saturation and "field capacity."
Range: Saturation to "field capacity" (0 to between −5 and −33 kPa).
Precision: Fair to good. Better on coarse-textured soils where water moves faster.
Accuracy: Probably the most accurate method because samples are large.
Procedure: A relatively large area is saturated, then covered with plastic to prevent evaporation. Drainage is monitored with tensiometers and a neutron soil moisture probe. Other methods such as gravimetric sampling can also be used. If only one method is used to monitor soil water, bulk density and a moisture retention curve must be provided for each soil horizon so that both pressure potential and volumetric water content can be found. The plastic establishes a plane of zero water flux at the soil surface. Water is assumed to move only vertically. The change in volumetric water content above any depth during a time interval is the flux across that depth. Tensiometers measure the hydraulic gradient at each depth. The hydraulic conductivity is calculated using Darcy's law to give curves of hydraulic conductivity versus water content or pressure potential for all soil depths.
Time: Frequency of reading depends on the rate of change in water content. Readings are made more often at the beginning of the experiment than toward the end. Most of the data are obtained within a week. Time required to measure hydraulic conductivity increases as hydraulic potential decreases or soil depth increases.

C st: Equipment consists of a neutron probe (about $2,500), two or three access tubes (about $20), two or three sets of tensiometers (between $100 and $500, depending on the number and type), and a large sheet of plastic (about $20). A source of water (2 to 6 m^3) is needed. Labor requirements are about 1 day to set up, and 30 minutes per set of readings.
Advantages: Minimum soil disturbance and large plots help make this the most accurate method. It reflects natural soil conditions better than any other method. All horizons can be measured simultaneously.
Disadvantages: Sites must be fairly level. It may be difficult to install tensiometers and neutron probe access tubes in soils containing coarse fragments. Hydraulic conductivity can be determined only between saturation and "field capacity." The method may require excessive time in soils with low hydraulic conductivity or restricting layers. Longer time is required for hydraulic conductivity measurement at greater depths. Small hydraulic gradients can increase the error associated with this method.
Comments: It is desirable to use both a neutron probe and tensiometers since moisture retention curves based on laboratory results do not always agree with field data. A faster but somewhat less accurate variation of this method allows evaporation by omitting the plastic. Calculations then must be made to determine the location of the plane of zero flux that changes position as the soil loses water. If one suspects the flow may not be vertical, a trench can be constructed around the plot to insure vertical movement.
References: Ogata and Richards, 1957; Rose, Stern, and Drummond, 1965; Nielsen, et al., 1964.

Method: Hot air.
Measurement: Diffusivity (*D*) as a function of soil water content. Hydraulic conductivity can be calculated from diffusivity and moisture-retention data.
Range: Approximately -3 kPa to $-5{,}000$ kPa.
Precision: Fair because of dependence on slope of graph, determination of gravimetric moisture content, and small samples.
Accuracy: Good for strictly matrix flow. Fair overall because of small samples.
Procedure: A small undisturbed core in a cylinder 10 cm high and 5 cm in diameter is saturated with water. A hot air stream (250°C) is applied to the upper surface of the soil for approximately 10 minutes. For the mathematical assumptions used in the calculations to be valid, the loss of weight must be proportional to the square root of time. The lower part of the soil core should remain at the original water content. The soil is removed from the cylinder with a piston, and water content is measured gravimetrically in 5-mm-thick slices. A graph of water content versus depth is used in calculating diffusivity.
Time: Thirty minutes for measurement, including moisture determination
Cost: Inexpensive. Equipment includes a hot air gun (about $200), cylinders, and a balance.
Advantages: Rapid, inexpensive, simple.
Disadvantages: Requires moisture retention curve to calculate unsaturated hydraulic conductivity. Not suitable at pressure potential near saturation.
Reference: Arya, Farrell, and Blake, 1975.

Method: One-step outflow.
Measurement: Diffusivity (*D*) as a function of soil water content. Hydraulic conductivity can be calculated from diffusivity and moisture-retention data.
Range: About -10 kPA to -300 kPa. Range can be extended in low end by using higher pressures.
Precision: Fair because of small sample.
Accuracy: Fair because of assumptions made in theory and used in calculations and small sample.
Procedure: A soil core is placed in a pressure chamber on a porous ceramic plate. A predetermined air pressure is applied and the rate of water outflow is determined as a function of time. Diffusivity can be calculated from the instantaneous rate of outflow, water content, and sample geometry.
Time: Two to six weeks, depending on the thickness and hydraulic conductivity of the sample.
Cost: Inexpensive. Equipment consists of a small pressure chamber for the sample, an air pressure source, and a buret or some means of measuring water outflow. If the procedure is not automated, much time is required the first day or two to make the readings.
Advantages: Controlled laboratory conditions. Many samples can be run simultaneously. Hydraulic conductivity can be determined over a relatively wide range of moisture content.
Disadvantages: May not reflect actual soil conditions adequately. Handling the sample may alter its hydraulic characteristics. Sample geometry is difficult to define in soils containing coarse fragments. Hydraulic conductivity cannot be determined in the wet range (less than -10 kPa). Moisture-retention curves are needed to

calculate hydraulic conductivity from diffusivity.
References: Gardner, 1956; Doering, 1965.

Method: Core.
Measurement: Unsaturated hydraulic conductivity as a function of pressure potential or water content.
Range: Approximately 0 to -20 kPa.
Precision: Poor to fair because samples are small.
Accuracy: Fair because samples are small.
Procedure: A soil core is placed in a specially designed pressure chamber between two porous plates or membranes. A constant pressure head of water is applied to each end of the core. Lower pressure at the lower end creates a hydraulic gradient across the sample. The water content of the soil is regulated by using an external gas pressure system. Hydraulic conductivity is calculated from the steady-state rate of water flow through the core and the hydraulic gradient as measured with tensiometers placed in the side of the soil core.
Time: A few hours to several days depending on thickness, hydraulic conductivity, and water content of the sample.
Cost: Inexpensive. Equipment consists of pressure chamber, pressure source, and burets.
Advantages: Controlled laboratory conditions. Many samples can be run simultaneously.
Disadvantages: May not adequately reflect actual soil conditions because of small sample and possible disturbance of soil when collecting sample. Steady-state flow may alter soil hydraulic properties. Sample collection can be a problem in soils containing coarse fragments. Restricted range of measurement. Highly variable results require many samples from each horizon.
Reference: Klute, 1965.

Limitations of the Existing Flow Theory for Describing Hydrology of Natural Soils

Flow equations are available to describe water movement through isotropic, homogeneous soil materials. Results relate to infiltration of water, redistribution after infiltration, capillary rise from a water table, etc. Necessary data are hydraulic conductivity and moisture-retention curves. The flow equations, which are based on Darcy's law, yield results that define flux of water in the soil. The flux is equivalent to the velocity of a water surface if there were no soild phase. In other words, the *real* velocity of water in the soil pores is always higher than the flux given by the Darcy equation. In soils, the average velocity v_a in the soil pores is equal to v/θ where v is the Darcy velocity and θ is the volumetric soil water content (m^3/m^3). The pore flow velocity can be used to calculate travel time t of water m ving through the soil, as follows: $t = L/v_a$ where L is length of soil to be traversed. Travel times are important because they govern the effectiveness of adsorption, filtration, and precipitation processes and are essential when considering use of the soil for waste disposal or leaching chemical fertilizers and pesticides.

Most natural soils, however, are *not* isotropic and homogeneous. Interconnected large and fine pores induce irregular flow patterns because water moves preferentially along the large pores, bypassing finer pores inside peds. Flux can be measured well under these conditions but only if large, representative samples are used. Optimal sample size should be determined with the aid of morphological analysis. However, travel times are difficult to predict because of preferential flow patterns. Two methods can b used to characterize these patterns.

Construction of Physical Breakthrough Curves Using Chemical Tracers

A more detailed discussion of this topic was presented by Bouma (1977, p. 62). However, a flow diagram of saturated flow illustrates the basic processes. The percolating water is displaced at time zero by water with a tracer (for example, calcium chloride). The tracer concentration in the effluent is measured as a function of time. Piston-flow theory predicts that the water in soil is displaced as a sharp front, and the tracer concentration in the column effluent becomes equal to the influent concentration rather abruptly (curve 1 in fig. 1).

However, in natural soils, the tracer moves very rapidly through large continuous voids and the effluent is only slightly diluted by untraced water that is simultaneously displaced from the finer pores (curve 2 in fig. 1). The shape of the breakthrough curve, therefore, is an indicator of pore-continuity patterns in soil. Of course, physical analysis does not allow determination of real pore sizes nor can different types of pores be distinguished. Breakthrough data can predict travel time (as a function of different flow regimes) in soils by means of the theory of hydrodynamic dispersion. Unfortunately, natural soils often display nonideal behavior. Then theory fails, and direct measurement is the only alternative. Descriptions of macrostructure may be helpful to extrapolate results of such measurements empirically. (Anderson and Bouma, 1977b; Bouma and Anderson, 1977.)

Determination of Flow Patterns Using Dyes as Tracers in Micromorphometric Analysis.

In contrast to the soil physical analysis, morphological research allows conclusions on which *types* of pores conduct water. Use of dye as a tracer (for example, methylene blue) is essential to characterize the water-conducting pores functionally. Recent research in the Netherlands has shown that flow patterns thus obtained can be used to calculate saturated hydraulic conductivity in certain soils (Bouma, et al., 1977; Bouma, Jongerius, and Schoonderbeek, 1979). More important, it was established that the large water-conducting pores occupied only a small volume of soil (often less than 1 percent by volume). Such a small volume can be characterized only by morphometric techniques.

So far, the soil conditions discussed can generally be characterized according to existing theories of water movement and hydrodynamic dispersion, provided that samples of pedal soils are of adequate size. However, conditions in dry or in many slightly moist clay soils with cracks or large continuous pores cannot be described by existing theory. But such conditions have significant effects on the infiltration rate of many natural soils in the growing season. In such soils water may move down rapidly along the large pores (nonmatrix flow) bypassing dry, fine, porous peds. This phenomenon has been called short circuiting (Bouma and Dekker, 1978). The diagrams in figure 2 (p. 8) further illustrate this. Water infiltrates into large, vertically continuous pores only if the capacity of the soil between the large pores (often prisms) is inadequate to conduct the water vertically. This capacity is relatively high in a dry prism and decreases with time, as shown in the infiltration curve in figure 2 in which a single prism was

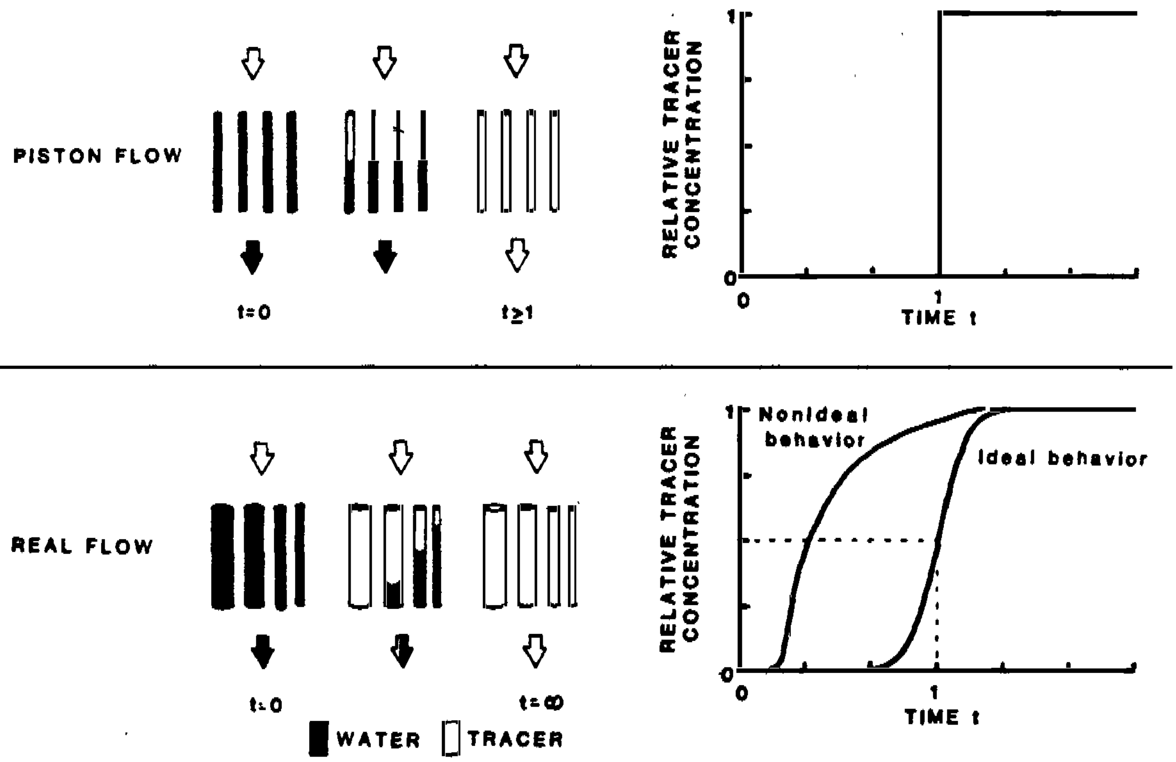


Figure 1. Hypothetical breakthrough curves illustrating piston flow in idealized soil and real flow in natural soil. Ideal behavior can be described with an apparent dispersion coefficient, whereas nonideal behavior cannot. Time = 1 represents the time required for one pore volume of water to be displaced.

measured with a small infiltrometer. In other words, the infiltration rate at the end of the prisms (matrix flow) determines whether water flows rapidly down through the large pores. Some relevant flow systems are represented schematically in figure 2:

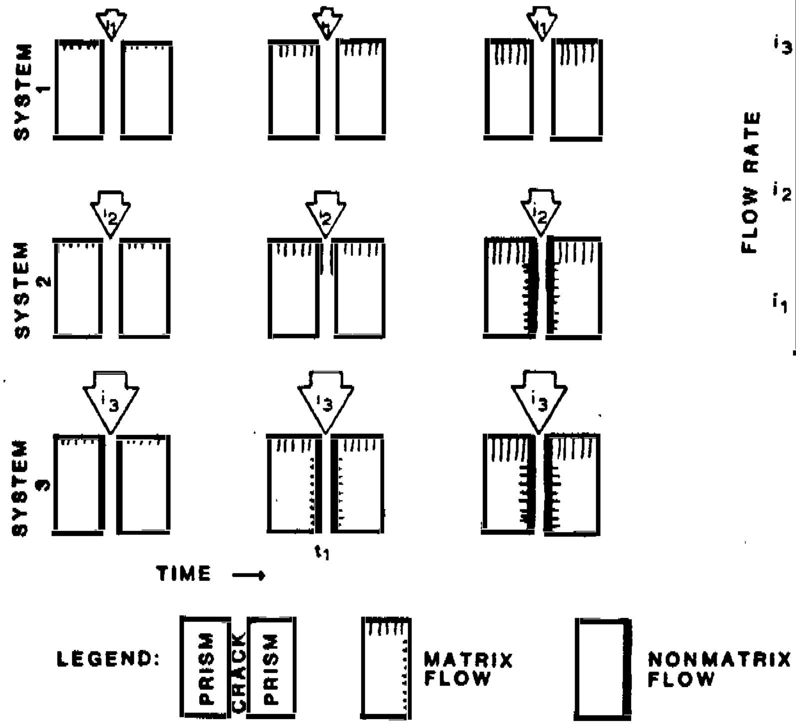
System 1 has an applicate rate (i_1) lower than the minimum infiltration rate into the prism. Only matrix flow occurs and there is no short circuiting.

System 2 has an application rate (i_2) that is initially lower than the infiltration rate into the prism but becomes higher after time t_1 because of decreasing infiltration rate into the prism. Matrix flow occurs until time t_1; after that there is short circuiting and nonmatrix flow.

System 3 has an application rate (i_3) that is higher at all times than the maximum infiltration rate into the prism. There is immediate short circuiting and both matrix and nonmatrix flow. Note that there is more short circuiting if the prisms are wet initially (assuming the large pores are still continuous).

When downward flow occurs, any existing lateral hydraulic gradient induces lateral infiltration into the prism. The ultimate distance of lateral infiltration will depend on the magnitude of this gradient and on the resistance of any clay skins present. In any case, at the same depth there is free-flowing water adjacent to unsaturated soil. Standard flow theory does not recognize this. In fact, we should consider the soil to consist of two separate subsystems of flow, one composed of the continuous pores and a second composed of the peds. The short-circuiting phenomenon is probably very common in natural soils. Close cooperation between soil morphologists and hydrologists is required to develop practical procedures to predict such phenomena.

Figure 3 summarizes the discussion of this section.



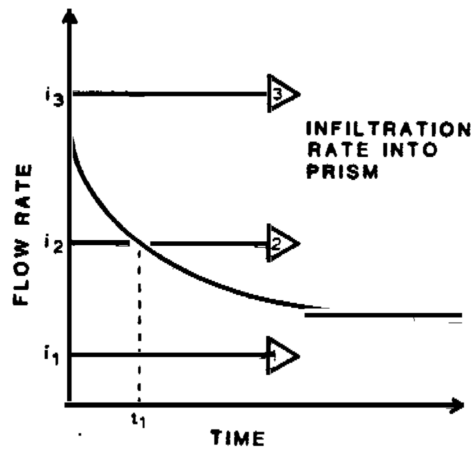


Figure 2. Schematic diagram of infiltration into cracked soil. System 1 has a flow rate small enough that only matrix flow occurs. System 2 has a flow rate that initially produces strictly matrix flow, but as the soil wets and the hydraulic gradient decreases, the infiltration rate is exceeded and nonmatrix flow occurs. Flow in system 3 is large enough that nonmatrix flow predominates.

Practical Field Use of Flow Theory

Simplified Flow Theory

A barrier to the use of physics theory for applied field work has been its highly mathematical character. However, water flow can be described nonmathematically. An example of this approach, relating to hydrodynamic dispersion, was presented in the preceding section. More examples appear in Bouma (1977, p. 32-39, 56-62). Another example concerns unsaturated steady-state upward and downward flow from a water table. Figure 4 (p. 10) shows two steady-state profiles for upward and downward flow. Water movement is governed by the gradients of the gravitational potential (z) and the pressure potential (*h*). No water moves when the total potential (*H*) has identical values at all levels in the soil. The total potential is zero at the ground-water level if we define $z = 0$ there. Since *z* increases linearly with height above the water table, pressure potential should decrease correspondingly if the total potential is to remain constant. Consequently, equilibrium is characterized by 45° lines of *z* and *h* (fig. 4).

Curves 1 and 2 in figure 4 were calculated without complex mathematics by graphic integration based on the Darcy equation (for details see Bouma, 1977, p. 57-60). Curve 1 describes a condition of downward flow. For example, an *h* value of -25 cm (h_1) is reached at 28 cm above the ground water. So $H_1 = +3$ cm (z_1 is 28 cm) and water moves downward to the ground-water l vel ($H = 0$). Curve 2 describes a condition of upward flow. For example, an *h* value of -25 cm (h_2) is reached at 22 cm above the ground water. So H_2 is -3 cm ($z_2 = 22$ cm), and movement is upward.

Pressure potentials higher than those at equilibrium (area above equilibrium *H* line) result in downward movement. Those that are lower (area below *H* line) result in an upward movement. Such diagrams can explain, for example, why water can move downward even though the soil becomes wetter at greater depths. It is not change in moisture content but the potential gradient that determines the direction of water movement.

Some Applications of Existing Concepts

A few examples are cited in which existing concepts can be refined by using the soil physical data discussed.

Terms of the water balance: Drainage rates at specified water contents in different soils can be estimated by using hydraulic conductivity and moisture retention curves. The "field capacity" concept can be refined by defining it as the water content at a very low flux, which in effect represents a semistatic condition (e.g., 1 nm/s). The "available water capacity" is then the difference between the moisture content at this

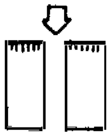
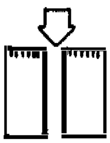
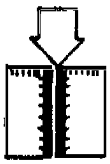
	APPLICABILITY OF EXISTING FLOW THEORY		ALTERNATIVES
	FLUX (DARCY THEORY)	BREAKTHROUGH (HYDRODYNAMIC DISPERSION)	
 MATRIX FLOW	YES	YES	--
 MATRIX FLOW WITH NON-MATRIX FLOW	YES (LARGE SAMPLES; USE MORPHOLOGY)	YES (IDEAL BEHAVIOR) NO (NONIDEAL BEHAVIOR)	-- EMPIRICAL MEASUREMENT
 NONMATRIX FLOW WITH MATRIX FLOW	NO	NO	DEFINE SUBSYSTEMS USING MORPHOLOGY. EMPIRICAL MEASUREMENT

Figure 3. Applicability of existing flow theory to predicting hydrology of field soils.

low flux and the moisture content at −1500 kPa (wilting point).

Infiltration capacity: This term describes the long-term infiltration rate, which is equal to the saturated hydraulic conductivity of the horizon within a meter of the soil surface that has the lowest saturated hydraulic conductivity value.

Potential frost action: Frost damage originates from water movement toward ice lenses. As with capillary flow from the water table, these flow processes can be described by the flow theory to explain why coarse sands and clays are less susceptible than silts and sandy loams. Hydraulic conductivity curves permit quantitative predictions for soils.

Septic tank filter fields: Percolation rates cannot be used to size filter fields because they describe saturated flow at unknown gradients. Measurements *in situ* below operating filter fields indicate that the soil is unsaturated because of clogging. Laboratory experiments show, moreover, that unsaturated conditions are needed in unclogged soils to obtain adequate purification of the liquid waste. Purification is correlated with long travel times and lack of short circuiting. Measured hydraulic conductivity can be used to derive optimal loading rates for different soils. Breakthrough curves for different flow regimes can be measured on large undisturbed cores to determine optimal loading rates in structured soil.

Permeability Estimates

Permeability is the sum of matrix and nonmatrix flux. In massive or weakly structured horizons that lack large channels caused by biological activity (worm holes, etc.), the pore size distribution of the matrix generally determines permeability. The fragmental class has the highest permeability, followed by sandy and sandy-skeletal materials. The sandy materials can be further subdivided according to the grading of the soil material (well-graded materials have lower permeability).

Within coherent materials, bulk density of the fine earth and particle-size distribution can be used to estimate the relative permeability. The nonmatrix flux contribution determines permeability in very moist or wet soils with moderate or strong grades of structure (other than platy) or with common or abundant channels of biological origin. Only a very few continuous nonmatrix voids result in high permeability. The pedologist must judge from the morphology the relative continuity of the

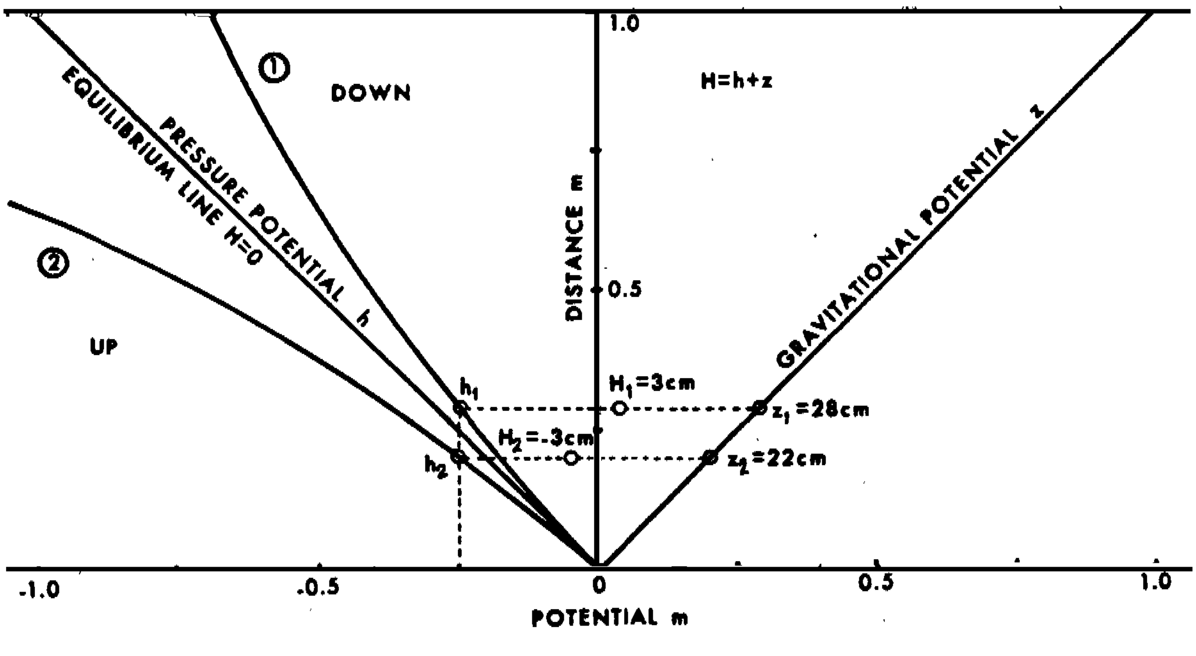


Figure 4. Examples of hydraulic potential profiles in soil for downward flow (curve 1) to a water table (water table occurs at distance = 0) and capillary rise (curve 2) from a water table. Total hydraulic potential H = pressure potential *h* + gravitational potential *z*. If the slope (tangent to the curve) of the pressure potential curve is greater than −1 (e.g., −½) then the direction of water movement is down. If the slope is less than −1 then water is moving upward

nonmatrix voids. There are a few clues for doing this. As the number of nonmatrix voids per unit area increases, the chances increase that a few voids, either alone or through interconnection with other voids, will provide continuous passageways. Probability for continuity also rises as the size of the void increases. Hence, other things being equal, large nonmatrix voids indicate greater permeability than small ones. The nature of the surfaces that surround the nonmatrix voids indicate continuity. Presence of clay skins or skeletans indicate continuity; the more prominent and complete these features are, the stronger is the evidence for continuity. Pressure surfaces, slickensides, and flattened roots indicate that the planar voids close when the horizon is wet and that continuity is probably low.

Using Unsaturated Hydraulic C nductivity Curves

Unsaturated hydraulic conductivity curves can be characterized by mathematical equations. For example (fig. 5):

$$K = ah^{-b} \qquad [6]$$

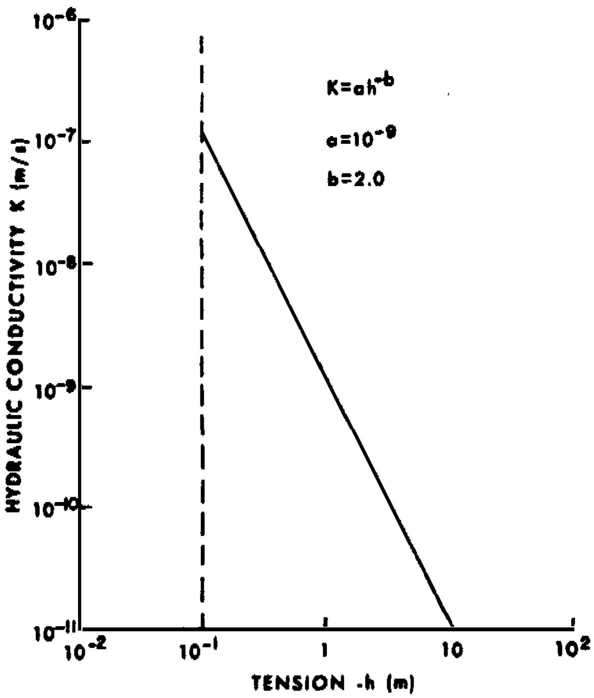


Figure 5. Hypothetical hydraulic conductivity curve as a function of soil water tension (negative of pressure potential).

where *a* and *b* are constants characteristic of specific soil horizons. Preliminary data from work in the Netherlands indicate that certain groups of unsaturated hydraulic conductivity curves can be defined in terms of characteristic populations of *a* and *b* values. The groups, as such, must be defined in such a way that they can be predicted from field pedological observations. It is expected that criteria for the groups will include texture, structure, clay mineralogy, and possibly others. The task now is to measure enough unsaturated hydraulic conductivity curves for major soil series to permit calculations of standard deviations.

Recent work on scaling of measured hydraulic conductivity curves may be important to express statistically the variability of, for example, measurements made within a given mapping unit (Warrick, et al., 1977). The scaling procedure is relevant only for soil conditions where measurement of hydraulic conductivity curves is physically meaningful and does not solve the problems of predicting water movement into and through dry soil with cracks.

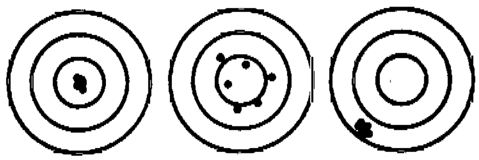
Summary and Conclusions

1. Flow theory is needed to refine soil hydraulic interpretations in soil surveys. However, its application to applied field work requires translation of the mathematical theory into a format that can be readily applied to practical problems. Examples are provided.
2. A distinction is proposed between systems having matrix flow and those having nonmatrix flow. Theory for the latter is inadequate now, and cooperation is urgently needed between hydrologists and morphologists to develop practical theories. So far, empirical procedures have been helpful. One example is discussed.
3. Soil survey application of flow theory should not be restricted to measuring flux but should also include observations of flow patterns that are important for many soil survey interpretations. Physical and morphological techniques are discussed.
4. Physical research has focused primarily on theory and simulation. Less emphasis has been placed on methods of measuring hydraulic conductivity. Currently used field and laboratory methods are discussed.

Because a soil can vary widely spatially, flow theory must be applied at many locations in the survey. This requires relatively inexpensive and fast measurements. The column, hot air, and outflow methods may be suitable.

Glossary

Accuracy and Precision: Accuracy is a measure of closeness to the true value. Precision is a measure of scatter between values. A set of measured values may have a lot of scatter (low precision), but if the average is close to the true value then it is accurate. See diagram.



Diffusivity: Diffusivity is the factor relating water flux to soil moisture content gradient and is defined as

$$D = K \frac{\Delta h}{\Delta \theta}$$

where D is the diffusivity (m^2/s), K is the hydraulic conductivity (m/s), h is the pressure potential (m) and θ is the volumetric soil moisture content (m^3/m^3). For horizontal flow, Darcy's law can be written in terms of diffusivity as:

$$v = -D \frac{\Delta \theta}{\Delta x}$$

The use of diffusivity requires the assumption that K, h, and $\Delta h/\Delta \theta$ are unique functions of θ.

Flux: The flux (v) of water (or any material) is the quantity (Q) moving through an area (A) during a time (t).

$$v = \frac{Q}{At}$$

This is dimensionally equal to velocity, and in reference to water movement v is called the Darcy velocity.

Gradient: The difference in quantities at two points divided by the distance between the points. For example, hydraulic gradient is the difference in hydraulic potential at two points in a soil divided by the distance between the points.

Homogeneous: A homogeneous medium implies that the spatial variation in the medium occurs on a scale smaller than that used for the determination (either visually or mechanically).

Hydraulic conductivity: The factor in Darcy's law relating water flux to hydraulic gradient. If potentials are expressed as energy per unit volume with units of J/m^3 or Pa, Darcy's law is written

$$v = -K' \frac{\Delta(h\rho g + z\rho g)}{\Delta L}$$

and the hydraulic conductivity (K') has units of m^3 s/kg. Most data reported are actually $K'\rho g$ (m/s) which results from expressing potentials in meters (i.e., the potentials are divided by ρ and g, the density of water and acceleration due to gravity, so that they are expressed as energy per unit weight. The hydraulic conductivity (K'), therefore, must be multiplied by ρ and g). Darcy's law is then written

$$v = -K \frac{\Delta(h + z)}{\Delta L}$$

where $K = K'\rho g$. Either K' or K can be used for hydraulic conductivity as long as the correct units are used, since both are factors relating a flux to a gradient.

Hydrodynamic dispersion: The phenomenon that results in a velocity distribution of water moving through soil. This is caused by size distribution of pores (water moves faster in large pores than in small ones) and tortuosity (actual path length relative to the mean direction of flow).

Isotropic: Soil properties that are isotropic are independent of direction. For example, if a soil is isotropic with respect to hydraulic conductivity, vertical hydraulic conductivity is equal to horizontal hydraulic conductivity.

Matrix flow: Flow entirely within pores or voids created solely by the natural packing arrangement of the primary soil particles (silt, clay, sand, and coarse fragments).

Nonmatrix flow: Flow in pores (cracks or channels) that are larger than voids resulting from natural packing of the primary soil particles.

Pascal (Pa): Unit of pressure (N/m^2) in the SI system. A pascal is equal to 10^{-5} bar or about 0.1 cm H_2O.

Permeability: This term is used in soil surveys for vertical saturated hydraulic conductivity. Strictly speaking, permeability (k) is saturated hydraulic conductivity (K in m/s) corrected for the density (ρ) and viscosity (μ) of water and the gravitational constant (g):

$$k = \frac{K\mu}{\rho g}$$

If the saturated hydraulic conductivity K' has units of m^3s/kg (see hydraulic conductivity definition) then permeability k is given by

$$k = K'\mu$$

The correct units of permeability are square meters (m^2).

Pressure potential: The pressure or energy required to remove a quantity of water held by matric (suction) forces in soil. It can be thought of as the pressure the water exerts on the soil. It is positive when free water is standing in soil and negative when the soil is unsaturated and holds the water with matric forces. Tension is pressure potential multiplied by -1, thus low pressure potential is equivalent to high tension and vice versa.

Saturation: The water content of a soil at a pressure potential of zero. In most soils air-filled porosity is about 5 percent at this point. If a soil is saturated for long periods (i.e., below a water table) the air-filled porosity may approach zero.

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