Water retention and flow characteristics of six calcarious soils of Montana
by Michael Frank Browne

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Soils
Montana State University

Abstract:
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A new hydraulic conductivity model has been proposed and warrants further study as a result of the hydraulic conductivity determinations from the flooded-plot experiment used in this field research. This new hydraulic conductivity model, should it prove to be valid with more thorough testing, determines maximum field hydraulic conductivity if the soil bulk density and 1/3 bar volumetric water content are known. Determination of the maximum field hydraulic conductivity is important because this value gives a good estimate of the rate at which soluble chemicals (fertilizers, herbicides, pesticides, etc.) can be transported by water through a soil system.

The effect of high amounts of CaCO3 on hydraulic conductivity was also investigated. Results from this research indicate no significant correlation between hydraulic conductivity and the total CaCO3 content of the soil.

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Symbol	Meaning
K	Hydraulic conductivity (cm3/cm3)
ψ_m	Matric soil water potential (cm)
Θ	Soil water content (cm3/cm3)
V	Soil water flux, or flow velocity (cm/sec)
ψ_n	Hydraulic soil water potential (cm)
z	Verticle distence, or a given depth (cm)
$K(\Theta)$	Unsaturated hydraulic conductivity
$K(\Theta/\psi_h)$	Unsaturated hydraulic conductivity - dependent on soil water content and potential conditions (cm/min)
$K(\Theta')$	Unsaturated hydraulic conductivity - dependent on the average soil water content (cm/min)
ψ_g	Gravitational soil water potential (cm)
$d\psi_h/dz$	Hydraulic gradient - saturated soil conditions (dimensionless)
$d(\psi_m+\psi g/dz)$	Hydraulic gradient - unsaturated soil conditions (dimensionless)
q	Quanity of flow (cm3)
t	Time
q/t	Flow rate
ρ	Fluid density (g/cm3)
g	Acceleration of gravity (cm/sec^2)
k	Intrinsic permeability (cm^2)
f	Fluidity of a liquid or gas (1/(cm sec)

LIST OF SYMBOLS-Continued

Symbol	Meaning
η	Fluid viscosity (g/cm min)
Q	Cumulative infiltration (Eq. [8]) (cm3/t)
a	The value of Q at t=1 (Eq. [8])
b	Slope of the infiltration line (Eq. [8])
ψ_t	Total soil water potential (cm)
ψ_p	Pressure potential (cm)
ψ_π	Osmotic potential (cm)
WC	Gravimetric water content for a given water content (Eq. [12])
WCR	Residual water content (Eq. [12])
WCS	Satiated water content (Eq. [12])
h	Matric soil water potential (Eq. [12])
a,m,&n	Dependent parameters used in Baumer's (Baumer and Rice, 1986) water retention model (Eq. [12])
FC	Field Capacity
NSSL	National Soil Survey Laboratory, Lincoln, Nebraska
MSU	Montana State University
K_m	Maximum hydraulic conductivity (cm/min)
Θ_m	Maximum soil water content (cm3/cm3)
β	m/m-1 where m is the slope of the regression using the Watson Equation (Eq. [16])
m	Slope of the regression using the Watson Equation (Eq. [16])
I	Intercept of the regression using the Watson Equation (Eq. [16])

LIST OF SYMBOLS-Continued

Symbol	Meaning
W	Total water stored above a depth z
θ'	Average soil water content above depth z (Eq. [17])
J	Slope of the regression using the CGA equation (Eq. [17])
A	Intercept of the regression using the CGA equation (Eq. [17])
ELD	Expected low density (Tables 3-8)
EMD	Expected medium density (Tables 3-8)
EHD	Expected high density (Tables 3-8)
VLP	Volume of large pores
$\theta_{-1/3\ bar}$	Soil water content at -1/3 bar soil water potential

ABSTRACT

A soil water retention model developed by Dr. Otto Baumer (Baumer and Rice, 1986) from the Soil Conservation Services' National Soil Survey Laboratory was tested on six different Montana soils. Results obtained using Baumer's model were similar to the results obtained in the laboratory. Baumer's model can compute the soil water retention gravimetrically or volumetrically at any pressure. This model can save time and money by eliminating expensive laboratory and field studies.

A new hydraulic conductivity model has been proposed and warrants further study as a result of the hydraulic conductivity determinations from the flooded-plot experiment used in this field research. This new hydraulic conductivity model, should it prove to be valid with more thorough testing, determines maximum field hydraulic conductivity if the soil bulk density and 1/3 bar volumetric water content are known. Determination of the maximum field hydraulic conductivity is important because this value gives a good estimate of the rate at which soluble chemicals (fertilizers, herbicides, pesticides, etc.) can be transported by water through a soil system.

The effect of high amounts of $CaCO_3$ on hydraulic conductivity was also investigated. Results from this research indicate no significant correlation between hydraulic conductivity and the total $CaCO_3$ content of the soil.

INTRODUCTION

Importance

Water movement through and drainage from soil profiles is an important factor that must be considered when addressing many environmental or agricultural problems. A quantitative analysis of two fundamental soil hydrologic properties must be considered when assessing and information concerned with water problems that result from drainage or irrigation of soils.

The first hydrologic property to be considered is the relationship of hydraulic conductivity to either the soil water suction or the soil water content. Hydraulic conductivity, as Koorevaar et al. (1983) define it, is a measure of the ability of the soil to conduct a flow of water and is very important when considering problems associated with irrigation or soil drainage. Subsurface drainage water can contribute to groundwater pollution as well as pollution of rivers and streams. Drainage from soil profiles is also responsible for the destruction of cropland as the result of saline seep.

A second important hydrologic property that should be considered is the relationship of soil water potential to soil water content. This is commonly referred to as the soil water retention curve (also referred to as the soil-water characteristic curve or the soil-water release curve). Soil water retention curves are used to predict soil moisture changes as well as hydraulic potential gradients in soil-water flow systems. Soil water retention curves are also used as input for some hydraulic conductivity models (Arya and Paris, 1981).

Soil water content and soil water potential can be measured independently but are most often described by the soil-water characteristic curve because they are functionally related. The soil-water characteristic curve is important because it expresses the influence of structure, porosity, pore-size distribution, and adsorption on the energy status of soil water. This energy status and how it varies in the soil profile determines the direction and influences the rate of soil-water movement and rate of water uptake by plants (Hillel, 1971).

The increasing need to more fully understand soil characteristics relating to movement of water in soil is complicated by the fact that, usually, the evaluation of soil properties for an entire field or drainage system is based on limited data from only a few locations. Extrapolation of results from just a few locations to larger land areas is experimentally convenient and necessary because of the expense and amount of time involved with laboratory measurements or in situ field determinations of either the hydraulic conductivity or the soil water retention curve. Measurements of the hydraulic conductivity and estimations of the soil water retention curve, for these reasons, are important. The six soils examined in this experiment are extensive in Montana and have properties that are similar to other soils in Montana. Therefore, the hydraulic conductivity and moisture retention data from these six soils can enhance the usefulness of soil data from similar soils in other parts of the state.

Extrapolation of this information becomes more important as governmental agencies such as the United State Department of Agriculture (USDA), Soil Conservation Service, begin using "allowable plant stress" as a method to help agricultural producers schedule irrigations.

California voters, in November 1986, passed Proposition 65 (Gransbery, 1987). This law prohibits any business from knowingly releasing into any source of drinking water, any chemical agent known to cause cancer or reproductive toxicity. It also makes individual farmers liable for drinking water contamination. It is not unreasonable to suggest that other states as well as Montana could pass similar laws in the future. If so, the need to more fully understand soil characteristics as they relate to water movement phenomenon becomes more important.

Objectives

The objectives of this study were as follows:

1. Verify Baumer's (Baumer and Rice, 1986) soil water retention model for six different Montana soils.
2. Calculate the unsaturated hydraulic conductivity for these six different Montana soils.
3. Attempt to characterize, quantify, and evaluate the effects of high amounts of soil carbonates on water movement and infiltration in these six Montana soils.

LITERATURE REVIEW

The relationship between hydraulic conductivity (K), soil-water potential (ψ_m), and soil-water content (Θ) is of much interest to many scientists concerned with the soil-water problems that exist in agriculture as well as the environment. A graphical representation of the relationship between soil water content and soil water potential (also referred to as matric potential) is often called the soil water retention curve. It is necessary to understand these soil-water concepts and the relationship that exists between them to interpret the results from models used to compute soil water retention curves or hydraulic conductivity. An understanding of these concepts may also help to interpret the effects of soil carbonates on water movement within the soil profile.

Saturated and Unsaturated Flow

The basic concept of water movement in soil is described by Gardner (1962):

> "When a soil is near saturation, the larger the pores the greater the rate of flow per unit of applied force. However, when a soil is not saturated, large pores contribute little to flow. Water moves on particle surfaces and through the finer pores under these conditions. Two forces cause liquid water to move through soil pores: gravity and adhesion. Gravity causes a downward pressure on water. This force is most important in saturated soil. The second force, adhesion, is due to the attraction of soil particle surfaces for water and becomes important in unsaturated soil. Adhesion - together with cohesion, which causes water molecules to hang together - is responsible for water rise in capillary tubes and the absorptive properties of blotting paper and other porous materials. When soil is very wet the gravitational force predominates. But in soil in which most crops grow, the major force causing water to move is due to adhesion. Water moves until these forces are balanced in the soil. Water films on soil particles will be uniform throughout any homogeneous soil, except for some verticle differences that

exist because of gravitation. Any prosity change occurring in soil affects the rate of water flow."

Liquid water flows in response to a hydraulic potential gradient and not necessarily in response to a water content gradient (Hanks and Ashcroft, 1980). The general equation, (Darcy's Law) used to describe liquid flow through a saturated soil is

$$V = K * d\psi_h/dz \qquad [1]$$

where V is the soil water flux (cm/s), K is the hydraulic conductivity (cm/s) and $d\psi_h$ is the difference in hydraulic potential (cm) between two points separated by a distance z (cm).

The moving force in a saturated soil is the gradient of a positive pressure potential whereas the moving force in an unsaturated soil is a negative pressure potential gradient (Hillel, 1982). Water flows from where the matric potential is higher to where it is lower (e.g., from -0.3 bar to -3.0 bar).

One of the most important differences between saturated and unsaturated flow is the hydraulic conductivity. In a saturated soil all of the pores are filled with water. Therefore, continuity of water filled pores and conductivity are at a maximum and flow occurs according to Darcy's Law (Eq. [1]). The saturated hydraulic conductivity (K) will be constant for any given soil. As soil water progresses from a saturated state to an unsaturated state, some of the pores become air filled and the conductive portion of the soil's cross sectional area decreases. The first pores to empty will be the largest and most conductive ones. This transition from saturation to unsaturation generally results in a steep drop in hydraulic conductivity with small changes in water content or potential. Therefore, the unsaturated hydraulic conductivity [$K(\Theta/\psi_h)$] is not a constant for a given soil but varies with the matric potential and the soil water and must be determined experimentally at different matric potentials. The unsaturated form of Darcy's Law is

$$V=K(\Theta/\psi_h)*d(\psi_m+\psi_g)/dz \quad [2]$$

The hydraulic conductivity (K) is now dependent on the soil water content and potential conditions (Θ/ψ_h) while the hydraulic gradient $d(\psi_m + \psi_g)/dz$ is determined by the matric (ψ_m) and gravitational potentials (ψ_g).

Hydraulic Conductivity

Hillel (1980 and 1982) defines hydraulic conductivity as:

(1) the ratio of the soil water flux (v) to the hydraulic gradient ($d\psi_h/dz$) or,

(2) the slope of the flux versus gradient curve.

One of the most important measurements that can be determined from the calculation of hydraulic conductivity as a function of water content [$K(\Theta)$] is the maximum field hydraulic conductivity (K_m). Determination of K_m is important because it describes water movement during the early stages of drainage and, thus, gives a good estimate of the rate at which soluble chemicals (fertilizers, herbicides, pesticides, etc.) can be transported by water through a soil system. It is also a valuable parameter in the design of irrigation systems. The true maximum field hydraulic conductivity will occur when the soil system is saturated. However, saturated soil systems, under field conditions, are very difficult to achieve because there will always be entrapped air in soil pores. Therefore, the K_m is somewhat less than K_s (saturated hydraulic conductivity). The maximum hydraulic conductivity is usually when the soil water content (θ) or average soil water content (θ') is at a maximum, θ_m, or θ'_m.

Soil Water Flux

The dimensions of soil water flux are quantity per area times time which reduces to length per unit time. Flux has the same dimensions as velocity and is often referred to as flow velocity. The flow velocity in soils is highly variable since the soil pores vary in shape, width, and direction. For example, wider soil pores will conduct water more rapidly than narrow pores (Hillel, 1982). The velocity of flow is also affected by the length of the path the water must travel. This length is actually greater than would be expected in a verticle soil column because the water particle must travel through pore passages that are not straight. This effect is referred to as tortuosity and is defined as the average ratio of the actual roundabout path to the apparent, or straight, flow path (Hillel, 1982). The tortuosity factor for water vapor movement is greater in wet soils than dry soils (Koorevaar et al., 1973) and the tortuosity factor for liquid water movement is less in wet soils than dry soils. Most water flow processes taking place in soils is laminar flow because of the relatively low flow velocities and narrow tube-like pores. Flow rates within individual pores, given laminar flow, are governed by Poiseuilles Law. Equations governing flow through cylindrical and planar voids, respectively, are

$$q/t = \rho g \pi \ (r^{**}4)/8n * d \psi_h/dz \qquad [3]$$

and

$$q/t = \rho g \pi (w^{**}3)/12\eta * d\psi_h/dz \qquad [4]$$

where q/t is flow rate, ρ is the fluid density, g is acceleration due to gravity, η is viscosity, r is the radius of a cylinder and w is the width of a planar pore (Hillel, 1971). It is evident from equation [3] and [4] that small changes in pore size can result in large changes in flow rate. The hydraulic conductivity as illustrated above is dependent on the pore geometry of the soil. This takes into account the total porosity, the distribution of pore sizes, and tortuosity but does not account for the fluid properties of the liquid.

It is possible in theory, and sometimes in practice, to separate K into two factors: intrinsic permeability of the soil (k), which is a description of the cross sectional area of pores available for flow, and fluidity of the liquid or gas (f) where

$$K=kf \qquad [5]$$

Fluidity is inversely proportional to viscosity:

$$f=\rho g/n \qquad [6]$$

therefore,

$$k=Kn/ g \qquad [7]$$

where n is the viscosity in poise units (dyne sec/cm**2), ρ is the fluid density (gm/cm^3), and g is acceleration due to gravity (cm/sec**2) (Hillel, 1982). The density of most liquids varies somewhat with temperature and solute concentration but for the most part remains nearly constant. Changes in fluidity result primarily from changes in viscosity. Although fluidity does vary with the composition of the fluid and temperature, it is evident that permeability is primarily affected by the pore geometry of the soil.

Infiltration into soils is a special case of soil water flux. The infiltration rate (IR) is defined by Hillel (Hillel, 1982) as the volume flux of water flowing into the profile per unit of soil surface area. The IR is measured in the same units as velocity (distance/time) and decreases with time. The initial IR is generally high, particularly in dry soil, because the matrix potential gradient causing water movement is large. As the soil wets the matrix potential gradient decreases with time and infiltration slows asymptotically, approaching a steady state. At this point the dominant force causing movement is the gravitational gradient. In relatively uniform soils, short-time (from minutes to hours, depending on the soil) infiltration can be described by the equation,

$$Q = at^b \qquad [8]$$

where Q is the cumulative infiltration, t is the time, a is the value of Q at time = 1 and b is the slope of the infiltration line when graphed on log - log paper. A and b are parameters that vary with the soil and the water quality. The magnitude of b is an indication of the rate of infiltration; the larger the b value, the greater the infiltration rate. By taking the logarithim of Q and t from Eq. 8 or plotting on log-log paper it can be transformed into a linear form.

Soil Water Content

The amount of water held in a soil system is referred to as the soil water content (Θ). The soil water content can be expressed on a volume basis (cm^3/cm^3) or a mass basis (g/g) [often referred to as the gravimetric water content (Θg)]. As can be seen in equation [2], water content is an important factor influencing unsaturated hydraulic conductivity ,$K(\Theta)$. The soil water content is clearly important for the growth of plants and plays an important role in such mechanical properties as consistency, plasticity, strength, compactibility, penetrability, stickiness, and trafficability. Soil water content is also important because it determines the air content and rate of gas exchange within the soil.

Hydraulic Gradient

The hydraulic gradient is the driving force causing water movement in the soil. The hydraulic gradient for a saturated soil, is ψ $d\psi_h/dz$) and for an unsaturated soil is $d(\psi_m + \psi_g)/dz$ (from Eq. [1] and [2]). The potential energy of water may differ from place to place and from time to time within the soil. Most water movement under normal conditions in the field occurs while the soil is in an unsaturated condition and the total soil-water potential is negative.

Hanks and Ashcroft (1980) define the total potential (ψ_T) of soil water as the sum of gravitational potential (ψ_g), matric potential (ψ_m), pressure potential (ψ_p), and osmotic potential ($\psi\eta$) as expressed in Eq. [9].

$$\psi_t = \psi_g + \psi_m + \psi_p + \psi_\pi \qquad [9]$$

Differences in total potential determine the direction and magnitude of water flow in a soil system. A measurement of total soil water potential is useful because it indicates the potential plants must overcome to remove water from the soil.

The gravitational potential of soil water is determined by the elevation of a given point relative to an arbitrary reference level. It is always positive or zero if the point of reference is below the point to be determined. The gravitational potential is always negative if the soil surface is chosen as a reference point.

The matric potential of soil water results from the forces of capillarity, as well as those of adsorption (hydration envelopes over the particle surfaces) on to the soil. These forces lower the potential energy below that of pure water and result in negative matric potentials. In unsaturated soils, capillarity results from adhesive and cohesive forces as well as the surface tension forces in the water menisci of the soil pore spaces. The matric potential of a saturated soil is zero. As a slight suction or pressure is applied to a saturated soil no outflow may occur until a certain critical value is exceeded at which time the largest pore begins to empty. This critical value is called the air-entry point and is generally small in coarse-textured soils and larger in fine textured soils (Hillel, 1982). The progressively smaller pores will continue to empty as suction or pressure is gradually increased. Only the very small pores will contain water at high suction or pressure values. Therefore, the amount of water remaining in a soil is a function of the number of soil pores, size of soil pores, and the amount of matric potential.

The pressure potential in a soil system is the pressure that is associated with a static head of water and exists only below a water table. It can be measured with a piezometer. There is no pressure potential in an unsaturated soil system.

The osmotic potential results from the presence of solutes in soil water. These solutes affect the thermodynamic properties of the soil water and lower its potential energy. This phenomenon does not significantly affect the flow of water in the soil profile unless there is a semipermeable membrane such as a plant root system. The sum of the matric and osmotic potentials is often used to characterize the energy status of soil water with respect to plant water uptake. In order to extract water from the soil, the water potential of a plant must be lower than the matric and osmotic potential of the soil water. In moist soils when the matric potential approaches zero, the osmotic potential may be the predominant component of the total potential. As the soil drys out, the matric potential becomes progressively more predominant in determining the total soil water potential. Matric forces are capable of lowering water potentials more drastically than most naturally occurring osmotic forces (Meidner and Sheriff, 1976).

The hydraulic potential is frequently used when the only concern is with the direction and magnitude of liquid water flow in the soil system. Hydraulic potential can be defined as

$$\psi_h = \psi_g + \psi_m + \psi_p \qquad [10]$$

remembering that either ψ_m or ψ_p is zero at any particular time.

Determination of Hydraulic Conductivity

Determination of hydraulic conductivity (K) as it relates to various suction and wetness values is required before any of the mathematical theories of water flow can be applied in practice. Hydraulic conductivity values must be measured experimentally and can be obtained using laboratory techniques or in situ field methods that have been recently developed. Hydraulic conductivity as a function of water content [K(Θ)] is usually determined in preference to hydraulic conductivity as a function of soil water potential $[K(\psi_h)]$ because $[K(\Theta)]$ it is easier to measure and is not affected by hysteresis.

Hysteresis is a phenomenon that accounts for the fact that water contents are greater during desorption than sorption at a given potential. Topp (1969 and 1971) as well as Staple (1970) determined that there was little, if any, effect of hysteresis on unsaturated hydraulic conductivity as a function of soil water content $[K(\Theta)]$ but the effects of hysteresis on unsaturated hydraulic conductivity as a function of soil water potential $[K(\psi_h)]$ was substantial.

Hydraulic conductivity $[K(\Theta)]$ values can be obtained from either steady-state or transient-state flow systems. In steady-state flow systems the flow characteristics (water content, hydraulic potential or gradient, and water flux) do not change with time but remain constant (Hanks and Ashcroft, 1980). Water is stored (or in some situations it is coming from storage) in the soil with transient or nonsteady state flow. Therefore in transient state flow, the water flux entering a soil would not equal the water flux leaving a soil and the difference between that entering and that exiting is the storage.

Steady-state flow and the determination of hydraulic conductivity can be accomplished in several ways. One method used to evaluate $K(\Theta)$ is to establish a steady flow of water to a water table. A second method used to establish steady-state flow involves the evaporation of water from a soil that has a water table at a fixed depth. A third method commonly used to attain steady-state flow conditions is to thoroughly wet a soil profile by flooding. Steady-state flow conditions are then approached during drainage of the soil profile. It takes a considerable amount of time , in all cases, to establish a steady-state. The change with depth in matric potential (ψ_m) from Eq.[10] is constant and equal to zero, when a steady-state of flow is attained. Consequently, the $d\psi_h = d\psi_g$ and flow will be entirely a result of the gravitational potential difference. The hydraulic gradient $[d(\psi_m + \psi_g)/dz$ [Eq. [2]] now equals the gravitational gradient ($d\psi_g$ / dz) which will be equal to one. This prompted Hanks and Ashcroft (1980) to establish the following general rule

> "If steady-state verticle water flow into soil has been established, the value of the hydraulic conductivity in the zone of constant matric potential (also constant water content) is numerically equal to the flux density of water application."

Eq. [2] can now be written as

$$V = -K \qquad [11]$$

The assumption of a unit hydraulic gradient such as that found in Eq. [11] was first introduced by Black et al. (1969). With this assumption the rate of change of soil water content in a soil profile can be used to calculate the hydraulic conductivity [$K(\Theta)$], as shown by Nielsen et al. (1973).

In the past, many studies dealing with water movement and behavior in soils have been conducted in the laboratory using columns of artificially packed or disturbed soils. Cassel et al. (1974) questioned the validity of extrapolating experimentally convenient laboratory results to field situations. Hillel (1982) also feels that all too often theories and models have been validated, if at all, only in highly artificial sets of laboratory produced conditions. Hillel states:

> "it is unrealistic to try to measure the unsaturated hydraulic conductivity of field soil by making laboratory determinations on discrete and small samples removed from their natural continuum, particularly when such samples are fragmented or otherwise disturbed".

However, Tommer (1986) did find that good saturated hydraulic conductivity data could be obtained from measurements on undisturbed cores in the lab. Klute (1965a,b) reviewed laboratory methods for determining both saturated hydraulic conductivity and unsaturated hydraulic conductivity.

Richards et al. (1956) and Ogata and Richards (1957) were the first to use flood plot methods on a bare field to determine K(Θ) under field conditions. K(Θ) was calculated from Eq. [1] using tensiometers to determine matric potentials and gravimetric methods for determining flux. By measuring the changes in water content due to drainage as a function of time and dividing this by the hydraulic gradient determined with the tensiometers, they were able to calculate K(Θ). The hydraulic conductivity values calculated from field data were found to be in agreement with values they had previously obtained using both laboratory and field methods. Arya et al. (1975) used a similar approach to determine K(Θ) in the presence of a growing crop. Nielsen et al. (1964) did a flood plot study but used a neutron probe to obtain water data and undisturbed cores for determination of water retention curves. A water budget approach for determination of hydraulic conductivities was used by Rose and associates (Rose et al., 1965; Rose and Stern, 1965). Watson (1966) developed the instantaneous profile method (IPM) for determination of K(Θ). This method is designed to include possible dynamic effects of non-steady-state flow in calculation of K(Θ). Cassel

(1974) developed a computer program for calculation of average Θ, flow rate, hydraulic gradient and conductivity for all depths and times, given desorption and tensiometer data collected during drainage of a covered flood plot.

Bouma and Dekker (1981) have more recently developed a field method for determination of saturated hydraulic conductivity that measures both the verticle and horizontal conductivity of a cube of soil. This cube of soil, about 25 cm^3, is carved out in situ and then covered with gypsum. Bouma and Dekker suggest that their method of determining saturated hydraulic conductivity can be used to measure both vertical and horizontal flow.

In recent years several unit gradient based mathematical models; the Lax alogorithm (Sisson et al., 1980), the "CGA" (Chong et al., 1981), and the Theta and Flux models (Libardi et al., 1980); that allow calculation of K(Θ) using only water content measurements at known times during drainage have been proposed. Tommer (1986) fit all of these models to drainage data and determined that the magnitude of the K(Θ) estimates followed the order CGA > Lax Θ methods > Lax W methods. Tommer felt that the CGA and W methods had an advantage in that they are based on measurements of profile storage rather than water contents at a single depth. Analysis of field data with both the CGA and Lax W (Watson equation) models allows a range of values of K to be obtained from a single data handling procedure. Tommer's results, with regard to the magnitude of K(Θ) values using these field models, were substantiated by Jones and Wagenet (1984).