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The application of mathematical models incorporating water flow theory to practical hydrologic problems (e.g., runoff, groundwater recharge, irrigation) requires a characterization of soil hydraulic properties. Such properties may vary widely over land areas of practical interest so that many measurements may be required for adequate soil characterization; thus, the methods used must be relatively rapid and economical. Field methods which have been successfully used in Hawaii to measure the hydraulic conductivity and sorptivity of surface soil are described. The methods are relatively simple and thus are useful for characterizing land areas on the scale of plantation fields or small watersheds. Underlying principles and detailed procedures are given for each method.

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FIELD METHODS FOR UNSATURATED HYDRAULIC CONDUCTIVITY AND SORPTIVITY

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for
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Water Conducting and Storage Properties of Watershed Soils

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ABSTRACT

The application of mathematical models incorporating water flow theory to practical hydrologic problems (e.g., runoff, groundwater recharge, irrigation) requires a characterization of soil hydraulic properties. Such properties may vary widely over land areas of practical interest so that many measurements may be required for adequate soil characterization; thus, the methods used must be relatively rapid and economical. Field methods which have been successfully used in Hawai'i to measure the hydraulic conductivity and sorptivity of surface soil are described. The methods are relatively simple and thus are useful for characterizing land areas on the scale of plantation fields or small watersheds. Underlying principles and detailed procedures are given for each method.

CONTENTS

INTRODUCTION

Application of soil water flow theory to many practical problems on specific soils requires estimates or measurements of hydraulic conductivity or of soil water diffusivity for unsaturated soil over the water content range of interest. In some cases, laboratory measurements on cores taken from the field or calculation methods which utilize the water content-pressure head relationship may provide useful conductivity or diffusivity data. In situ field measurements are generally preferred, however, if resources are available to conduct field measurements and if the site is sufficiently accessible and physically amenable to field measurement (e.g., reasonably level topography, not too stony, and predominantly vertical flow during drainage). A larger area of measurement and preservation of field structure are inherent advantages of field methods over laboratory methods. Field studies have demonstrated that hydraulic conductivity values of surface soils vary widely over relatively small areas. Adequate representation of the hydraulic properties of the surface soil in an agricultural field or small watershed may require a large number of measurements (Bresler and Green 1982). Thus, a potentially useful field method must be relatively rapid and simple. The method proposed here for unsaturated hydraulic conductivity and diffusivity satisfies these requirements and is appropriate for well-drained soils.

Sorptivity, a hydraulic property of unsaturated soils which was defined by Philip (1957*b*) in his development of infiltration theory, is closely related to hydraulic conductivity and soil water diffusivity. Sorptivity measurements can be made in the field under two types of water-entry conditions: slight positive pressure (ponded), and slight negative pressure (water entry through a porous plate). The former (ponded) method is described here. Field methods of measuring sorptivity will likely undergo continued development and testing as interest grows in the use of sorptivity data in infiltration prediction and for characterizing the spatial variability of hydraulic properties of surface soils.

The purpose of this report is to provide detailed descriptions of methods which have been developed or tested in a previous research project supported by the University of Hawaii at Manoa Water Resources Research Center (Chong and Green 1979; Chong, Green, and Ahuja 1981). Hydraulic conductivity and sorptivity data for a number of Hawaii soils which were characterized by these methods have been published (Green et al. 1982), but many soils remain to be characterized and the methods are still relatively unknown. Hopefully, this publication will aid those who require such methods, both in Hawaii and elsewhere.

SIMPLIFIED UNSTEADY DRAINAGE-FLUX METHOD OF MEASURING HYDRAULIC CONDUCTIVITY

Principles

This simplified method calculates $K(\Theta)$ by using only the experimental results obtained from periodic measurements of $\Theta(z,t)$ during redistribution of water in the soil profile following steady ponded infiltration. In addition to the assumption of negligible horizontal flow in the soil layer for which $\partial\Theta/\partial t$ is being evaluated, the simplified method assumes a unit hydraulic gradient, i.e., $\partial H/\partial z = -1$, during drainage.

The assumption of unit gradient during redistribution of soil water without evaporation following infiltration in a uniform soil profile was introduced by Black, Gardner, and Thurtell (1969). This assumption was used in the determination of hydraulic conductivity of unsaturated soil by Nielsen, Bigger, and Erh (1973). Thus, with $\partial H(z,t)/\partial z = -1$ at z_1, the equation relating unsaturated hydraulic conductivity at $z = z_1$ to the transient water content profile is

$$K(\Theta)z_1 = -z_1 \ (d\Theta^*/dt). \tag{1}$$

Formally, $\Theta^* = 1/z_1 \int_0^{z_1} \Theta \ (z,t) \ dz$ but Θ^* may be defined in different ways, depending on the experimental and calculation procedure used.

While equation (1) may be directly used to calculate $K(\Theta)$ over discrete time intervals in which measured changes in soil water content can be used to provide an estimate of $d\Theta^*/dt$, further refinements have been developed. Chong, Green, and Ahuja (1981) used the assumption of Richards, Gardner, and Ogata (1956) and Gardner, Hillel, and Benyamini (1970) that water content in the soil profile during post-infiltration redistribution diminishes with time in a manner such that

$$\Theta^* = at^b \tag{2}$$

where a and b are constants. Substitution of the derivative of equation (2) into equation (1) gives K as a function of t, i.e.,

$$K = -z_1 bat^{b-1} \quad . \tag{3}$$

Solution of equation (2) for t explicitly and substitution in equation (3) yields the working equation developed by Chong, Green, and Ahuja (1981),

$$K(\Theta)_{z_1} = -z_1 ba^{1/b}[\Theta^*]^{(b-1)/b} \quad . \tag{4}$$

This equation gives the hydraulic conductivity at depth z_1 for an average soil-water content over the depth increment $z = 0$ to $z = z_1$. Calculation of $K(\Theta)$ for a second depth z_2 would require an assessment of the rate of change of average water content for the entire layer above z_2, i.e., $z = 0$ to $z = z_2$, rather than for the increment z_1 to z_2.

The above procedure yields $K(\Theta)$ as a power function, a form widely used to represent conductivity as a function of water content (Brooks and Corey 1964; Campbell 1974; Clapp and Hornberger 1978). Alternative simplified field procedures based on measured values of Θ versus t for a draining soil profile assume an exponential K-Θ relationship (Libardi et al. 1980); these alternatives are briefly discussed in the Comments section.

Use of a single tensiometer at the depth z_1 for which K is to be determined provides a means of obtaining K as a function of soil-water pressure head, h, when Θ^* versus t is measured as described above. Expressing the absolute value of pressure, $|h|$ (namely, suction), as a power function of time (Chong, Green, and Ahuja 1981), i.e.,

$$|h| = mt^n \tag{5}$$

and substitution of equation (5) in (3) gives

$$K(h)_{z_1} = -z_1 abm^{-(b-1)/n}[|h|]^{(b-1)/n} \quad . \tag{6}$$

SOIL WATER DIFFUSIVITY. The field data required to determine $K(h)$ by equation (6), i.e., Θ^* versus t and h versus t (the latter only at the depth of interest, z_1), are sufficient to calculate $D(\Theta)$ as proposed by Gardner (1970), i.e.,

$$D(\Theta)_{z_1} = -L\ (dh/dt) \quad . \tag{7}$$

Further substitution of the derivative of equation (5) into (7) and the use of equation (2) to change the dependent variable from t to Θ^* (as was done previously in going from equation [3] to [4]) result in the power function,

$$D(\Theta)_{z_1} = -z_1 mna^{-(n-1)/b}[\Theta^*]^{(n-1)/b} \quad . \tag{8}$$

Method

EQUIPMENT AND PROCEDURE. The field set-up for the simplified method of determining $K(\Theta)$ requires an infiltration ring at least 0.3 m in diameter with a buffer zone around the inner ring. A double-ring infiltrometer is preferred over the larger plot with wood borders for mobility and ease of installation. Hydraulic conductivity at field saturation is estimated by the measurement of steady infiltration. The unsaturated K determination requires only the measurement of Θ^* versus t for the depths of interest. This can be accomplished with a neutron moisture probe or by soil sampling at various times during drainage with subsequent gravimetric determination of water content. The soil sampling method has the disadvantage of requiring soil bulk density values for each horizon to allow conversion of gravimetric water contents to volumetric water contents. However, soil sampling will frequently be the most practical way to obtain profile water contents over time because (1) soil variability may require calibration of a neutron probe at each field site and (2) neutron probe results for soil near the soil surface (top 15 cm) are frequently inaccurate. Since the simplified method will likely be used mostly for measuring $K(\Theta)$ in the upper zone of the soil profile (e.g., 0-0.5 m), the requirement of bulk density data is not a serious problem, as undisturbed core samples can be taken easily near the surface after water content sampling is completed. Bulk density data (and an assumed or measured particle density) allow calculation of porosity, which in turn can be used to estimate the field-saturated value of Θ^* from porosity x 0.85 (Chong, Green, and Ahuja 1981).

Following infiltration and initiation of drainage, measure the water content of the soil profile to depth z_1 in increments of 20 to 15 cm at about 2, 4, 8, 16, 24, and 48 hr. If soil sampling is used, place each increment of soil in a separate soil moisture can—with proper labeling of depth and sequence number—for subsequent determination of water content. Samples should be taken from two places, one on each side of the inner ring of the infiltrometer. If the inner ring is no larger than 0.3 m in diameter, soil samples must be taken in the buffer zone near the inner ring. Only two sets of samples can likely be taken within the inner ring. If the depth z_1 (to which the average water content is to be determined) is sufficiently deep to require more than two increments of soil sampled from a given sampling hole, sampling at greater depths is facilitated by using a 0.05 m diameter bucket auger for the upper depths and a smaller diameter screw auger for greater depths. If several depth increments are required, the bucket auger can be used to enlarge the hole to the greatest depth yet sampled with the screw auger, to allow easy sampling of the subsequent depth interval with the smaller diameter auger.

If $K(h)$ or $D(\Theta)$ are to be determined, a soil moisture tensiometer must be installed at each depth z_1 for which calculations are to be accomplished. Tensiometers will normally be placed near the center of the inner ring. Manometer-type tensiometers, preferably with rapid-flow porous ceramic tips, are required. Tensiometer readings can be easily taken more frequently than soil-water-content measurements (unless a neutron soil-moisture probe is used). Soil-water pressure and water contents do not have to be measured at the same times.

To apply equation (6) at pressures near zero, the conductivity can be assumed to be constant between $h = 0$ and $h = h_a$, the air entry pressure (Chong, Green, and Ahuja 1981). The air entry pressure to use in equation (6) corresponding to the saturated water content, Θ^*_s, can be obtained by solving (5) for the value of t that gives Θ^*_s when inserted in equation (2).

ANALYSIS. When soil-water contents (volume basis) are determined for equal depth increments, either by neutron probe or soil sampling, the mean soil-water content to depth z_1 for n increments is simply $\Theta^* = 1/n \sum_{i=1}^{n} \Theta_i$. If depth increments are of unequal length, then the water content for each increment must be weighted for the increment length, i.e., $\Theta^* = [1/(\sum_{i=1}^{n} L_i)] \sum_{i=1}^{n} \Theta_i L_i$, in which L_i is the length of the ith depth segment.

Preferred units for measured quantities, if the final result is to be in SI units, are Θ^* in m^3/m^3; t in s; z_1 in m; and h in m. Other units of length and time can be used, but the units must be consistent throughout the calculations. This is illustrated in the calculation example at the end of this section, where length is expressed in centimeters, time in minutes, and water content in cubic centimeters/cubic centimeters. Thus, the units for Θ^* and t in equation (2) must be used also in equation (4) when the parameters a and b obtained from (2) are used in equation (4). The parameters a and b in equation (8) are determined by converting equation (2) to the linear form by logarithmic transformation, i.e., log Θ^* = log a + b log t and fitting the transformed equation to the experimental data by linear least-square analysis. The same procedure is followed in determining m and n in equation (5) when $K(h)$ or $D(\Theta)$ are to be determined. Values of h must be expressed as suction head (positive values representing the absolute value of soil-water-pressure head), both in the determination of m and n in equation (5) and in calculating $K(h)$ with equation (6).

SAMPLE CALCULATION. The calculation of $K(\Theta)$ is illustrated with data obtained in a previous field study. The experimental site (M4-1, also labeled HSPA-A) was located on the Hawaiian Sugar Planters' Experiment Station, Field L, in Kunia, O'ahu. The approximate location is shown on a map given in Figure 1 of Green et al. (1982). The soil is the Molokai series, a member of the subgroup, Typic Torrox. The volumetric water contents obtained at each sampling time after cessation of ponded water infiltration are presented in Table 1.

TABLE 1. VOLUMETRIC WATER CONTENTS (Θ^*) OF MOLOKAI SOIL AT VARIOUS TIMES (t) DURING DRAINAGE AFTER FLOODING, KUNIA, O'AHU, HAWAI'I

t (min)	Θ^* (cm^3/cm^3)
58	0.471
446	0.430
1498	0.400
4277	0.365
8784	0.352
13090	0.347
20261	0.336

NOTE: For 0-20 cm depth increment.

A least-squares fit of the equation, log Θ^* = a + b log t to the data in Table 1 resulted in a = 0.6079 and b = -0.0595, with a correlation coefficient of -0.995. The log-log plot of the data in Table 1 is shown by Chong, Green, and Ahuja (1981, Fig. 1).

Using the values of a and b obtained above and the soil depth z_1 = 20 cm, we can now calculate the hydraulic conductivity—water content relationship with equation (4). The $K(\Theta)$ results for selected values of Θ^* are given in Table 2.

TABLE 2. HYDRAULIC CONDUCTIVITY (K) AT SELECTED SOIL-WATER CONTENTS (Θ) CALCULATED BY EQ. (4)

Selected Soil-Water Contents, Θ (cm^3/cm^3)	Hydraulic Conductivity K (cm/min)
0.50	2.23×10^{-2}
0.45	3.41×10^{-3}
0.40	4.18×10^{-4}
0.35	3.88×10^{-5}
0.30	2.49×10^{-6}

NOTE: Molokai (Typic Torrox) soil; HSPA Site A (M4-1), Kunia, O'ahu.

While $K(\Theta)$ may be calculated for any value of Θ, the determination of K is most accurate for water contents close to those measured in the field. Chong, Greenn, and Ahuja (1981) found that the upper limit of Θ, i.e., the field-saturated water content, could be estimated by 0.85 x total porosity, where total porosity was calculated from measured bulk density (ρ_b) and estimated or measured particle density (ρ_p) by the formula, porosity = $(1 - \rho_b/\rho_p)$.

COMMENTS

1. The simplified method will generally be used to determine hydraulic functions for the upper soil horizons in soils which are reasonably homogeneous with depth with respect to hydraulic conductivity. Determination of $K(h)$ and $D(\Theta)$ requires tensiometers at each depth for which these functions are required and thus makes the methodology more complex and perhaps less useful when numerous measurements are required and resources are limited. Thus, the method is most useful when only $K(\Theta)$ is required.
2. The assumption of unit hydraulic gradient is not strictly satisfied in most field soils because of natural horizon differences and tillage effects on hydraulic conductivity of soil at various depths in the profile. The relative hydraulic conductivities at various depths in the

soil profile (in relation to the conductivity at a given reference depth) will normally vary with time. For example, a tilled surface horizon may have the highest conductivity in the profile at the beginning of profile drainage (near saturation), but the highest relative conductivity may be associated with a deeper horizon a day or two later. This change in relative conductivity between various depths over time results in a change in hydraulic head gradient over time. Chong, Green, and Ahuja (1981) found that well-drained Oxisol soils achieved unit gradient at intermediate times (about 17 hr), with gradients less than one at earlier times and greater than one at later times.

3. Two alternative methods of calculating $K(\Theta)$ with only Θ versus t experimental field measurements were developed by Libardi et al. (1980). These methods, called the "Θ-method" and the "flux-method" utilize equation (1), assume that $\partial H/\partial z = -1$, and require that K is related to Θ by an exponential relationship, namely, $K = K_o \exp [\beta(\Theta - \Theta_o)]$, in which β is a constant and K_o and Θ_o are the values of K and Θ during steady infiltration. In a comparison of these two methods of calculating $K(\Theta)$ with the method of Chong, Green, and Ahuja (1981) described above (designated the "CGS-method" in their paper), Libardi et al. (1980) found no significant difference in the results obtained by the three methods. If one desires to express the $K(\Theta)$ relationship by an exponential function, then one of the Libardi et al. (1980) methods would be preferred over the CGA-method because the latter method gives $K(\Theta)$ in the power function form.

SORPTIVITY BY PONDED INFILTRATION

Principles

Cumulative horizontal infiltration, I, into a horizontal soil column having uniform properties and water content is proportional to the square root of time, t, i.e., $I = St^{\frac{1}{2}}$. The coefficient of proportionality, S, which varies with initial water content and differs between soils, was termed "sorptivity" by Philip (1957*b*). Downward one-dimensional cumulative infiltration for sufficiently small times can be described by a rapidly converging power series in $t^{\frac{1}{2}}$ (Philip 1957*a*), namely,

$$I = St^{\frac{1}{2}} + At + Bt^{\frac{3}{2}} \; .... \qquad (9)$$

For a brief time after initiation of infiltration, perhaps less than 120 s, the first term of the infiltration equation dominates flow. Thus, it is possible to estimate the value of the sorptivity, S, for a given antecedent water content by measuring the cumulative infiltration versus time immediately following the application of water.

Method

This field sorptivity procedure involves measurement of cumulative infiltration as the head of ponded water in a single-ring infiltrometer falls over time. The technique is essentially that developed by Talsma (1969). It is a simple and rapid method with no specialized equipment required. Vertical flow is assured by insertion of the infiltration ring 0.10 to 0.15 m into the soil—depths which usually exceed that of the wetting front during the brief period of infiltration. The effect of decreasing head on the measured sorptivity is minimized by starting with a small head of only 0.02 to 0.03 m of water. Talsma (1969) assessed the effect of head change on S for several soils differing in soil texture and concluded that the error introduced by the drop in head during measurement is usually negligible compared with natural soil variability. Alternative constant-head, ponded infiltration techniques were found to be more difficult and less precise.

The ponding method is appropriate only for soils in which there is negligible flow through large cracks or channels (see Comments). The method is usually satisfactory for finely tilled surface soils or subsoils.

EQUIPMENT

1. Infiltrometer ring, 0.30 m in diameter and 0.15 to 0.20 m high
2. Graduated (mm) metal scale (200 mm long) or graduated glass tube (about 5 mm inside diameter) for measurement of water surface level over time, including clamping device to hold scale at proper angle (thermometer clamp attached to 12.5 mm diameter rod pushed vertically into soil just outside the ring provides a good clamp for capillary tube; capillary action in glass tubing facilitates reading water level in inclined tube).
3. Can or plastic beaker, 0.002-m^3 (2-ℓ) volume, to apply known volume of water to ring
4. Porous material with low water holding capacity, such as a plastic scrubbing pad, to break impact of water poured into infiltration ring

5. Stopwatch with provision for stopping and reading a sequence of event times ("lap times") while total elapsed time continues
6. Tape recorder, if measurements are to be made by one person

PROCEDURE

1. Prepare a level, plant-free surface about 0.5 m by 0.5 m.
2. Install infiltration ring to a depth of 0.10 to 0.15 m—in tilled soils installation should require only minimal force; compact soil next to ring with a pencil-sized rod to prevent boundary flow.
3. Take a composite soil moisture sample to a depth of 50 to 100 mm, preferably outside the ring, using a small diameter core sampler.
4. Install the measurement scale or capillary tube to the ring or auxiliary support rod at a 5 to 30° angle relative to the horizontal soil surface. The change in water level is amplified by a factor of $(\sin \alpha)^{-1}$, when the calibrated scale is at an angle α from the horizontal plane; e.g., when $\alpha = 30°$, a two-fold amplification in the reading is achieved. Measure the height "**a**" from the ground surface inside the ring to a reference point near the top of the inclined scale. The bottom of the scale should rest on the ground surface. The height "a" and the length "h" from the bottom of the tube to the upper index provide the data needed to calculate the factor **a/h** needed to convert readings on the inclined capillary scale to vertical positions that the water surface has dropped at various times. If a glass tube is used to measure water level, place a 50 mm by 50 mm flexible plastic sheet on the soil surface so that the lower end of the glass tube rests near the center of the plastic sheet; this will reduce the movement of soil particles into the glass tube.
5. Calibrate the beaker which is to be used to apply a known volume of water into the ring. A volume of 1.6 ℓ will provide a layer of water 20 mm deep in a 0.30-m diameter ring.
6. Quickly apply the water on the porous pad at the side of the ring opposite the measuring scale and simultaneously start the clock. A tape recorder can be used by a single operator to record both time and water level during the brief measurement period.
7. Obtain as many water height versus time readings as possible after the water surface is reasonably "quiet", until the water surface is about 5 mm above the soil surface. Generally, the best practice is to read

the times corresponding to preselected indices on the inclined measurement scale. Five or more readings are required for calculation of a reliable value of *S*.

8. Calculate the equivalent vertical position of the water surface for each time using the proper amplification factor. Since it is difficult to establish the exact position of the water level at $t = 0$ and, also, the actual cumulative infiltration is not required for calculating the slope of the cumulative infiltration curve, we need only the successive positions of the water level to calculate the slope, which is equivalent to the sorptivity.
9. Plot equivalent vertical position Z versus $t^{\frac{1}{2}}$ and determine the linear portion of the curve. Delete the data for early times which may be in error due to water roughness and other factors; also delete data for later times which do not conform to the linear $I - t^{\frac{1}{2}}$ relationship.
10. Calculate the slope *S* by a least-squares fit of $Z = St^{\frac{1}{2}}$ + intercept to the remaining data points.
11. Calculate the antecedent soil water content associated with the measured *S* value.

SAMPLE CALCULATION. The soptivity calculation is illustrated with field-measured data obtained in a study described by Chong and Green (1979). Eight sorptivity measurements were made in 1.8 m x 1.8 m plots surrounding the plot in which the hydraulic conductivity measurement was made for HSPA, Site A (see previous section). Data for plot 1 are given in Table 3. The measured values of **a** and **h** were **a** = 2.6 cm and **h** = 15.9 cm, so that the factor needed to convert distance in the inclined capillary tube to vertical distance for the falling water head is **a/h** = 0.1635.

Thus, the reading y on the inclined centimeter scale is multiplied by 0.1635 to obtain Z, the vertical position of the water level corresponding to each time. These increasing distances—the starting position being arbitrary—can be plotted directly against $t^{\frac{1}{2}}$ to determine the linear portion of the Z versus $t^{\frac{1}{2}}$ curve. These results are given in Table 4 and plotted in Figure 1. The "eye-fitted" line drawn through the points in Figure 1 indicates that none of the data points in Table 4 deviate severely from the linear relationship for Z versus $t^{\frac{1}{2}}$, so that it is not necessary to discard any of the data in the least-squares determination of the slope, which corresponds to the sorptivity, *S*. The analysis of Table 4 data gives

TABLE 3. WATER POSITION (y) IN INCLINED TUBE FOR VARIOUS ELAPSED TIMES (t) IN FIELD MEASUREMENT OF SORPTIVITY

Elapsed Time, t (s)	Water Position, y (cm)
10.3	9.4
21.4	10.5
32.4	11.7
45.2	12.9
55.2	13.6
69	14.8
91	15.6

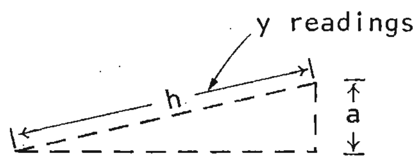

TABLE 4. VERTICAL POSITION (Z) OF FALLING WATER SURFACE VERSUS $t^{\frac{1}{2}}$

$t^{\frac{1}{2}}$ ($min^{\frac{1}{2}}$)	Z* (cm)
0.41	1.54
0.60	1.72
0.73	1.91
0.87	2.11
0.96	2.22
1.07	2.42
1.23	2.55

NOTE: Data calculated from Table 3.
*Z = y x 0.1635 (see text).

S = 1.30 cm/$t^{\frac{1}{2}}$ with a correlation coefficient of r = 0.996. The soil-water content corresponding to this sorptivity was 0.211 cm^3/cm^3 or 21.1% by volume. Sorptivity values diminish rapidly to zero as soil water contents approach saturation (Chong and Green 1979).

COMMENTS

1. Each sorptivity measurement, as described above, yields one point on an $S(\Theta_n)$ curve, where Θ_n is antecedent water content. A linear approximation of the $S(\Theta_n)$ relationship for a range of water contents from

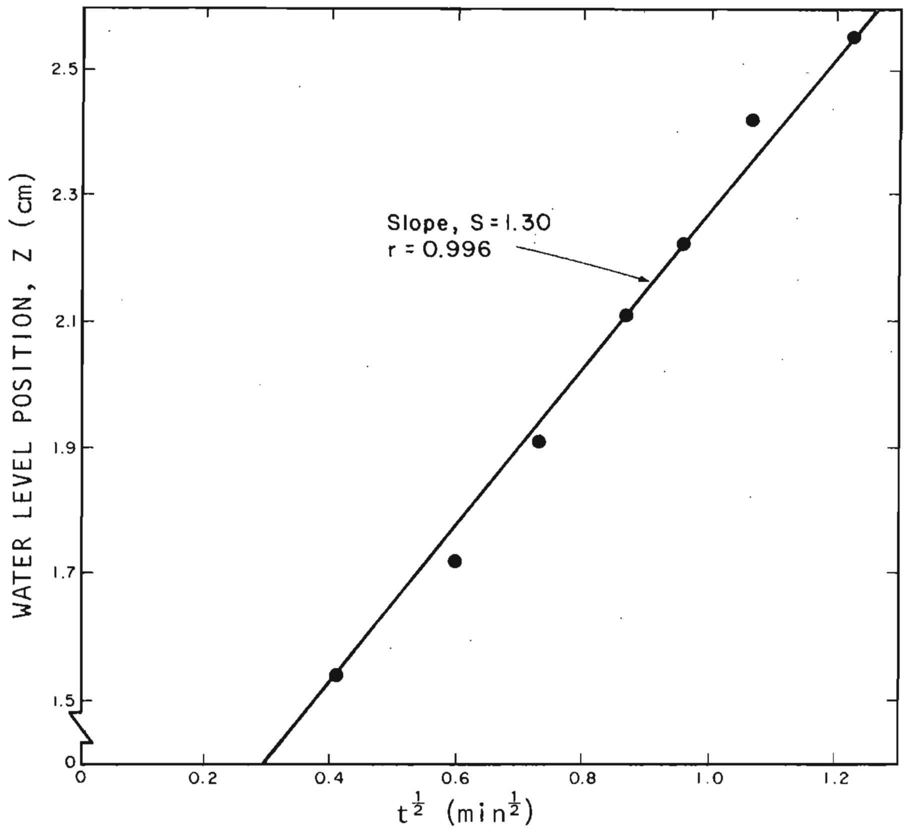

Figure 1. Plot of water level position, Z below a reference point versus $t^{\frac{1}{2}}$ to determine sorptivity.

saturation (Θ_n) to air dry can be obtained from the line plotted from $S = 0$ at $\Theta = \Theta_s$ through the measured value of S at $\Theta = \Theta_n$. Such an approximation was found to give reasonable estimates of infiltration at a number of field sites (Chong and Green 1979). Ponding of the soil surface may result in inaccurate estimates of sorptivity in soils which have worm holes, root channels, large cracks, or other large void spaces which are not representative of the soil matrix. In addition to the likely problem of inaccurate representation of larger areas with sorptivity measurements on such soils, sorptivities measured by this method are also subject to errors due to (1) the dependence of sorptivity on head and (2) the contribution of terms containing powers of t greater than $t^{\frac{1}{2}}$ in equation (9), so that the hydraulic conductivity at saturation contributes significantly to the measured cumulative infiltration. Talsma (1969) found that such soils had relatively large values of the ratio of hydraulic conductivity to sorptivity.

2. Sorptivity values measured over a field or watershed are likely to be log-normally distributed; this conclusion results from theoretical considerations (Brutsaert 1976) and from field measurement experience (Chong and Green 1979; Sharma, Gander, and Hunt 1980). Thus a representative sorptivity value for a large area from many measurements may best be obtained from the geometric mean rather than the arithmetic mean.

GLOSSARY

$K(\Theta)$	hydraulic conductivity (cm/min) as function of volumetric water content, Θ
$\Theta(z,t)$	volumetric water content (cm^3/cm^3) at depth, z (cm) and time, t (min)
Θ^*	weighted average water content from soil surface to depth, z_1 (cm^3/cm^3)
H	hydraulic head (cm)
h	soil water pressure head (cm)
$D(\Theta)$	soil water diffusivity (cm^2/min)
S	sorptivity ($cm/min^{\frac{1}{2}}$)

REFERENCES

Black, T.A.; Gardner, W.R.; and Thurtell, G.W. 1969. The prediction of evaporation, drainage and soil water storage for a bare soil. *Soil Sci. Soc. Am. Proc.* 33:655-60.

Bresler, E., and Green, R.E. 1982. *Soil parameters and sampling scheme for characterizing soil hydraulic properties of a watershed.* Tech. Rep. No. 148, Water Resources Research Center, University of Hawaii at Manoa, Honolulu. 42 p.

Brooks, P.H., and Corey, A.T. 1964. Hydraulic properties of porous media. Hydrol. Paper No. 3, Colorado State University, Fort Collins, Colorado.

Brutsaert, W. 1976. The concise formulation of diffusive sorption of water in a dry soil. *Water Resour. Res.* 12:1118-24.

Campbell, G.S. 1974. A simple method for determining unsaturated conductivity from moisture retention data. *Soil Sci.* 117:311-14.

Chong, S.K., and Green, R.E. 1979. Application of field-measured sorptivity for simplified infiltration prediction. In *Proc. Symposium on Hydrologic Transport Modeling*, New Orleans, Louisiana, 10-11 December 1979, ASAE Publ. 4-80, pp. 88-96.

______; ______; and Ahuja, L.R. 1981. Simple in situ determination of hydraulic conductivity by power function descriptions of drainage. *Water Resour. Res.* 17:1109-14.

Clapp, R.B., and Hornberger, G.M. 1978. Empirical equations for some soil hydraulic properties. *Water Resour. Res.* 14:601-604.

Gardner, W.R. 1970. Field measurement of soil water diffusivity. *Soil Sci. Soc. Am. Proc.* 34:832-33.

______; Hillel, D.; and Benyamini, Y. 1970. Post-irrigation movement of soil water. 1: Redistribution. *Water Resour. Res.* 6:851-61.

Green, R.E.; Ahuja, L.R.; Chong, S-K.; and Lau, L.S. 1982. *Water conduction in Hawaii oxic soils.* Tech. Rep. No. 143, Water Resources Research Center, University of Hawaii at Manoa, Honolulu. 122 p.

Libardi, P.L.; Reichardt, K.; Nielsen, D.R.; and Biggar, J.W. 1980. Simple methods for estimating soil hydraulic conductivity. *Soil Sci. Soc. Am. J.* 44:3-7.

Nielsen, D.R.; Biggar, J.W.; and Erh, K.T. 1973. Spatial variability of field-measured soil-water properties. *Hilgardia.* 41(7):215-60.

Philip, J.R. 1957*a*. The theory of infiltration. 1. The infiltration equation and its solution. *Soil Sci.* 83:345-57.

______. 1957*b*. The theory of infiltration. 4. Sorptivity and algebraic infiltration equations. *Soil Sci.* 84:257-64.

Richards, L.A.; Gardner, W.R.; and Ogata, G. 1956. Physical processes determining water loss from soils. *Soil Sci. Soc. Am. Proc.* 20:310-14.

Sharma, M.L.; Gander, G.A.; and Hunt, C.G. 1980. Spatial variability of infiltration in a watershed. *J. Hydrol.* 45:101-22.

Talsma, T. 1969. *In situ* measurement of sorptivity. *Aust. J. Soil Res.* 7:269-76.