

Flux-Controlled Sorptivity Measurements to Determine Soil Hydraulic Property Functions¹

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ABSTRACT

Using a new method, the hydraulic diffusivity and conductivity functions $D[\theta]$ and $K[h]$ can be determined over most of their range in about 4 hours. This method uses a series of fast and simple, one-dimensional adsorption measurements for which the cumulative volumes of absorption are controlled proportional to the square root of time with syringes driven by cams. Each combination of motor speed, cam size, and syringe size represents an imposed sorptivity for which the water content and/or pressure head at the interface must be measured after they become steady. Steady values are normally reached after only a few minutes. Pressure heads can be recorded during the absorption runs, and at the end a soil sample for gravimetric water content determination can be obtained in about 15 sec. Hydraulic diffusivity and conductivity functions derived from measured sorptivity functions were obtained for a variety of undisturbed and hand-packed soil columns. They

all showed excellent agreement with independent data. The flux-controlled boundary resulted in uniform wetting almost every time, whereas previously used potential-controlled boundary conditions often caused nonuniform wetting, which prevented accurate evaluation of sorptivities. The speed, range, and accuracy of this improved version of the sorptivity method compare favorably with other available methods. The method is suitable for routine and large scale use.

Additional Index Words: hydraulic conductivity, diffusivity, infiltration, boundary conditions, gamma-ray attenuation.

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EXTENSIVE USE of unsaturated flow theory in field situations, such as watershed hydrology and irrigation management requires convenient methods of determining the soil hydraulic properties. Methods available for measuring hydraulic conductivity k and hydraulic diffusivity D as functions of either pressure

head h or volumetric water content θ are generally time-consuming and/or laborious. For relatively recent reviews on the subject see Klute (1972) and Bouwer and Jackson (1974).

A few years ago, I introduced a fast and simple method of determining these soil hydraulic property functions (Dirksen, 1975a). It involved determining sorptivity functions $S[\theta_1, \theta_0 = \text{constant}]$ or $S[h_1, h_0 = \text{constant}]$ from a series of one-dimensional absorption measurements, each initiated by a step-function increase of the water content from θ_0 to θ_1 or pressure head from h_0 to h_1 at the adsorption interface. Each measurement had to start from the same uniform initial values θ_0 or h_0 , and needed to be continued only long enough to establish the familiar linear relationship (Philip, 1969):

$$i = S \sqrt{t} \quad [1]$$

where i is the cumulative volume of absorption per unit area of absorption interface, and t is time. Theoretically, this relationship is valid only for one-dimensional absorption into a semi-infinite horizontal medium, and then for all time $t > 0$. Practically, S can often be identified after only a few min, whether the flow is horizontal or vertical.

Once the sorptivity function is obtained, the diffusivity function can be calculated according to:

$$D[\theta_1] = \frac{1}{1+\gamma} \times \frac{1}{(\theta_1-\theta_0)^\gamma} \times \frac{d}{d\theta_1} \frac{\pi S^2[\theta_1, \theta_0 = \text{constant}]}{4(\theta_1-\theta_0)^{1-\gamma}} \quad [2]$$

This equation was derived in the original paper (Dirksen, 1975a, Eq. [14]), and discussed in two later publications (Dirksen, 1975b, 1977). From a practical point of view, γ is a constant that can be varied between 0.50 and 0.67 without significantly affecting the results. It is easier to vary γ if it is eliminated from the differential as follows:

$$D[\theta_1] = \frac{\pi S^2}{4(\theta_1-\theta_0)^2} \left\{ \frac{(\theta_1-\theta_0)}{(1+\gamma) \log e} \times \frac{d}{d\theta_1} (\log S^2) - \frac{1-\gamma}{1+\gamma} \right\} \quad [3]$$

This equation also shows that $D[\theta_1]$ can be obtained by differentiating a polynomial regression of $\log S^2$ in terms of θ_1 . This makes best use of all the original data and can be performed easily with a programmable calculator.

Since the soil water retentivity function $\theta[h]$ is hysteretic, so is the diffusivity function $D[\theta]$. From the nature of the sorptivity measurements, it follows that the diffusivity function obtained is only valid for wetting. If the wetting retentivity function is known or also measured, the hydraulic conductivity function $k[\theta]$ can be calculated according to:

$$k[\theta] = D[\theta] \omega (d\theta/dh) \omega \quad [4]$$

The $k[\theta]$ function is often considered nonhysteretic, but there may be some secondary hysteretic effects (Dirksen, 1978).

Sorptivity measurements require only one controlled boundary. In the original paper, I used a simple Mariotte-type buret to supply water at constant negative pressure head through a porous ceramic plate to a smooth soil surface confined by a short, sharp-edged cylinder. No measurements needed to be made within the soil. Unfortunately, the Mariotte buret regulated the pressure head inside the ceramic plate, rather than at the soil interface. The soil provided the driving force for the water to overcome the regulated pressure in the buret and any hydraulic resistances. Unless the resistance of the ceramic plate is negligible, h_1 will increase as the flow decreases with time. Contact resistances between the plate and the soil were even more serious. They can be very large and may differ from site to site, be nonuniform across the plate, and vary with time. As a result, wetting fronts were often uneven and the rates of absorption did not follow the square root of time relationship, preventing proper identification of sorptivities.

The solution to these problems with plate and contact resistances is shifting control of the flow rate from the soil to the apparatus, that is, changing from a potential-controlled to a flux-controlled boundary condition. In this paper, this new experimental procedure is described and results with it are compared with independently determined data. The experimental problem is now changed from determining S for a known regulated value of θ_1 or h_1 to determining h_1 or θ_1 for an imposed value of S . Theoretically, these values should be determined at the interface itself. Practically, the wetting front initially advances fast enough and θ or h change little enough away from the interface that they can be determined in a shallow surface layer.

METHODS

A simple way of controlling the flow rate is to drive syringes with cams that displace the volume at a rate proportional to the square root of time. Plate and contact resistances are then automatically accommodated by the force the cam exerts on the syringe. If the cams are rotated at a constant speed ω rad sec⁻¹, they can be made from plots of the polar coordinates:

$$(\alpha, r) = [\alpha, R_0 + (C\alpha)^{1/2}] \quad [5]$$

The displacement of the syringes is then equal to the change in radial distance r , which starts with R_0 at $\alpha = \omega t = 0$ and increases according to $(C\omega t)^{1/2}$. If the diameters of syringe and soil core are d and D , respectively, then

$$i = (d/D)^2 (C\omega t)^{1/2} \quad [6]$$

Comparison with Eq. [1] shows

$$S = (d/D)^2 (C\omega)^{1/2} \quad [7]$$

Figure 1 shows a syringe pump and a number of cams with $R_0 = 1$ cm. The cam mounted on the pump has a value $C = 2$; those displayed on top of the scaler have C values of 1.0, 0.5, 0.2, and 0.1 from left to right. This is more cams than needed. Three cams for $C = 0.2, 0.63,$ and 2 , combined with standard syringes of 1, 5, and 50 (displayed on scaler), and a self-made syringe of about 200 cm³ (Fig. 2) provide 12 almost equal increments for the polynomial regression of $\log S^2$ in terms of θ_1 . With a 1/30-rpm ($\omega = 3.49 \times 10^{-3}$ rad sec⁻¹) motor, the corresponding values of S range from less than 10^{-6} to more than 10^{-4} m/sec^{1/2} which is adequate for many soils. For larger values of S , a 1/6-rpm ($\omega = 1.75 \times 10^{-2}$ rad sec⁻¹) motor was used. With this motor each measurement had to be completed within 6 min since $0 < \alpha \leq 2\pi$. This presented no problem since it was used only in the wet range. In the dry range, the limit of 30 min of the slower motor was also adequate. The drive shaft of

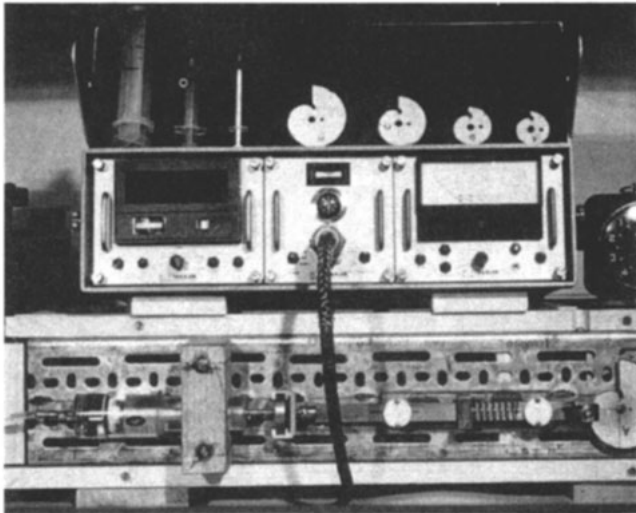


Fig. 1—Syringe pump with "square root of time" cams.

the pump was equipped with a thin rod that was supported by roller bearings and fit inside the slots in the cams near $\alpha = 0$. With longer slots the cams could be extended beyond $\alpha = 2\pi$. The rod rolled easily over the outside of the cam as it was being rotated, except at small values of α . There the cams tended to shear the rod rather than push the syringe, especially for larger values of C . This initial shearing was prevented by pushing either manually or by an adjustable compression spring such as shown in Fig. 1. The shearing problem can be eliminated by decreasing the radial distance of the cams according to

$$(\alpha, r) = [\alpha, R_0 - (C\omega)^{1/2}] \quad [8]$$

and modifying the drive shaft of the pump such that it stays in contact with the cam, e.g. with a spring. (This was suggested by P. Koorevaar, Laboratory for Soils and Fertilizers, Agricultural State University, Wageningen, The Netherlands.)

The controlled flux technique was tested in the laboratory on soil columns 10 cm in diam and 10- to 12-cm long. Except for a fine sand fraction that was hand-packed, the soil columns were obtained with a coring device and a hydraulic press, which left the cores virtually undisturbed from their field condition. A straight edge was used to smooth the surface of each core even with the edge of the brass sleeve in which the core was taken. To fill in any holes, a layer of the same dry soil was then sieved over that surface, compressed, and again made even with the brass sleeve. This gave uniform wetting fronts almost every time, even in very gravelly soil. The water content of the soil surface was at first measured with a 1-mm, collimated gamma beam from a 5-mCi ^{137}Cs source; later it was also determined gravimetrically. Moderately de-aired tap water with $EC \approx 0.06 \text{ S/m}$ was used.

The gravimetric sampling for measuring θ , must be done fast enough to prevent significant drying due to evaporation or continued water movement. The column holder visible in Fig. 2 was designed to do this in about 15 sec. Figure 3 shows a cross-sectional view through the center of the soil column and the restraining plate. Since all parts are transected except the rods in the corners, the customary crosshatching is omitted. The soil surface is prepared outside the frame but on the base plate with the restraining plate supporting the brass sleeve over part of its circumference (Fig. 2). The restraining plate is 4.75-mm (3/16-in) thick and slides snugly through a slot in the base plate. The raised center section of the base plate supports, with or without spacers, the soil column. The prepared soil column is slipped in the frame over a sliding plate that covers an inflatable rubber bag, of the type used for blood pressure measurements. The ceramic plate fits inside a small depression in the top surface of the frame and is held in place by a swinging and hinging arm. At the same time the syringe pump is started, the rubber bag is inflated, which presses the soil surface against the ceramic plate. The soil surface can be compressed up to 0.25 mm to ensure good contact. After that the pressure of the inflated bag is carried by the arm via the restraining plate, the brass sleeve, and the ceramic plate. When a run is terminated, the pressure in the inflatable bag is quick-

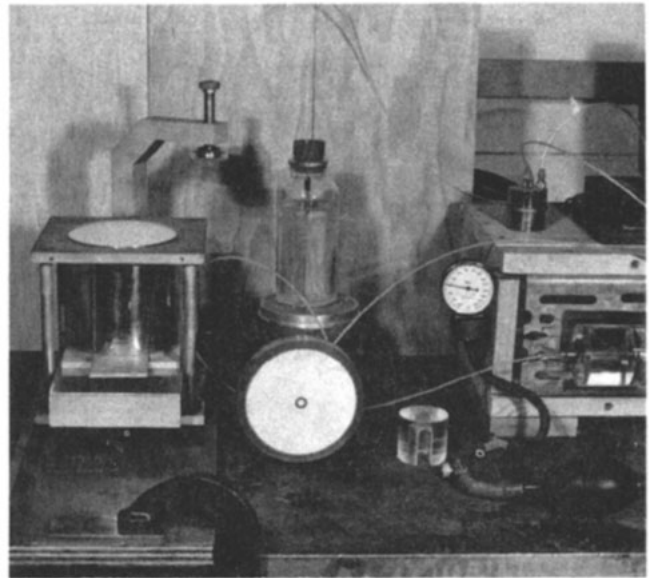


Fig. 2—Column holder for gravimetric sampling. Also shown are ceramic plate with small tensiometer and servo pressure sensor.

ly reduced so that the arm can be dropped below the top surface, the ceramic plate is removed, and the restraining plate is pulled out from under the brass sleeve. When the bag is gradually reinflated, the brass sleeve can, because of the stop, rise only until it is even with the top surface, but the soil column can be pushed up to any desired height beyond the top surface by the raised center section of the base plate. The soil protruding above the top surface is quickly cut with a large, sharp knife and put in a soil moisture can. In the wet range, the thickness of the soil sample can be of the order of 5 mm, but in the dry range it should be no more than about 1 mm. Since the entire soil surface is sampled, this still gives enough soil for an accurate determination of θ_1 .

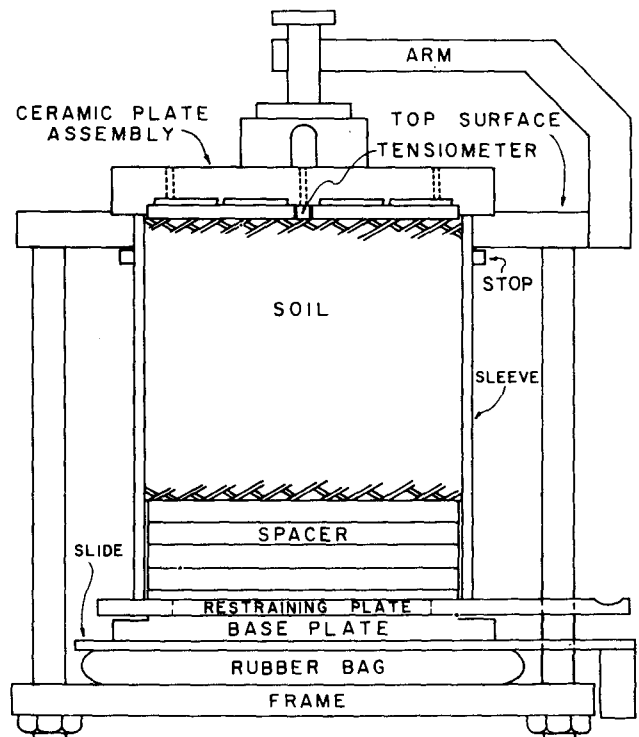


Fig. 3—Column holder for gravimetric sampling. Cross-section through center.

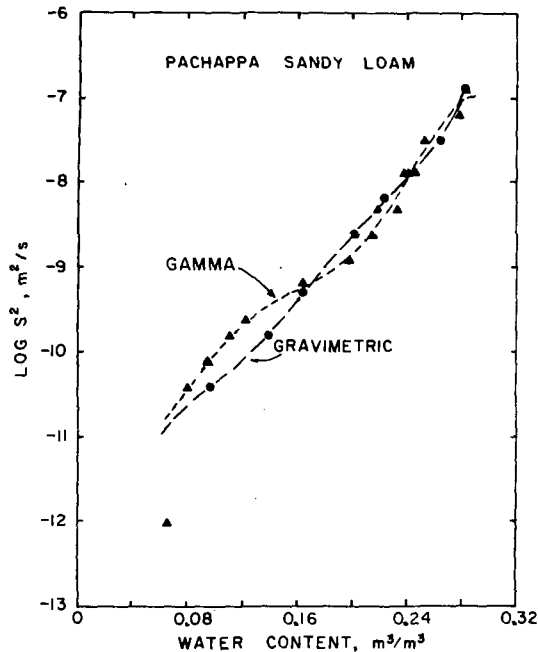


Fig. 4—Log S^2 vs. θ_1 for Pachappa sandy loam. Data points: \blacktriangle gamma attenuation, \bullet gravimetric. Dashed lines show 5th degree polynomial regressions.

A similar set-up was used with the 1-mm collimated gamma beam. The shielded source and detector and a soil column anchor were all mounted on a strong base to preserve alignment. The soil column was again placed on a small, inflatable rubber bag and raised against a flat aluminum surface that fixed the 1-mm gamma beam slightly beneath the soil surface to avoid surface irregularities. After gamma counts were obtained through dry soil, the aluminum plate was replaced by a porous ceramic plate and the cam rotation was started. Gamma counts were taken until they stabilized, which was often after a few minutes. This was true even with very dry soil, in which steady-state methods often take 1 week or more to yield results. If gamma counts do not stabilize in 15 min, this is probably the result of drift in the gamma equipment, changes in soil packing, or soil heterogeneities encountered near the wetting front.

After we finished a sorptivity run, the wetting front was immediately inspected for uniformity. While preparing a new soil surface, spacers of various thickness were used inside the brass sleeve to compensate for discarded wet soil. Depending on soil type, sorptivities used, and duration of individual measurements, from 5 to 10 measurements could be made on each soil core. It is most efficient to alternate two cores, especially with the gamma method if gamma counts or rates can be recorded. While a sorptivity run is made on one core, the surface of the other can be prepared. Using this procedure, one person can obtain a sorptivity function, and thus a diffusivity function, for routine characterization of soils within 4 hours.

Figure 2 also shows a small tensiometer flush with the surface in the center of the ceramic plate and the pressure transducer with which h_1 was recorded. With this design, the wetting soil-water-retentivity function and the hydraulic conductivity function can be determined at the same time. Since the time available to measure h_1 is very short, the pressure transducer must require essentially zero water displacement for a full-scale reading. This is the case with Servo Pressure Sensors (Model 314D, Sundstrand Data Control, Inc., Redmond, Washington)³ which operate on the force-balance principle. The tensiometer must be small, because it forms an impermeable spot for the water supply through the ceramic plate and depends for its functioning on lateral flow through the soil. Measured steady pressure heads are not affected by contact resistances.

³The mention of particular products or companies is for the convenience of the reader and does not imply endorsement, guarantee, or preferential treatment by the USDA.

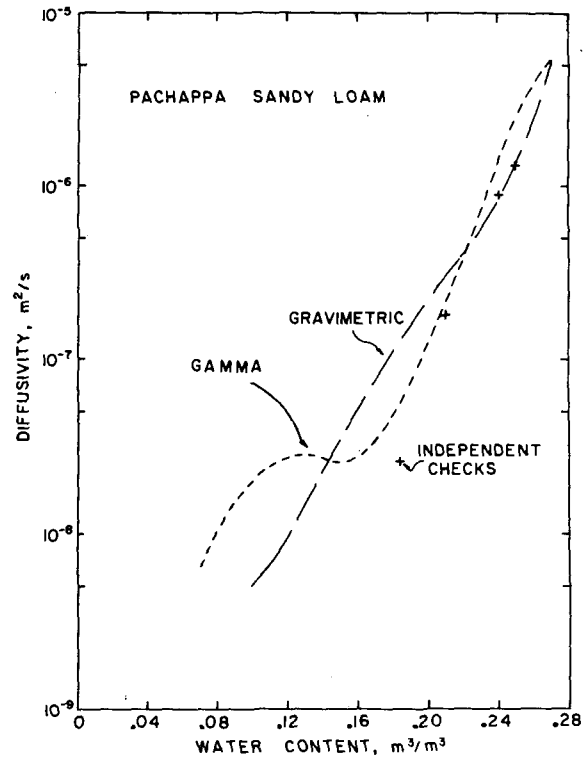


Fig. 5—Diffusivity functions of Pachappa sandy loam derived from polynomial regressions in Fig. 4 and Eq. [3]. Independent checks are listed by +.

RESULTS

Figure 4 shows plots of $\log S^2$ vs. θ_1 for Pachappa fine sandy loam (Mollic Haploxeralfs). Sixteen measurements were made by gamma attenuation, but only seven gravimetrically. The fifth degree polynomial regression of the gamma data does not include the one point at $\theta_1 = 0.066 \text{ m}^3/\text{m}^3$ to make the equation follow better the data points around $\theta_1 = 0.13 \text{ m}^3/\text{m}^3$. The diffusivity curve in Fig. 5 derived from this polynomial exhibits a relative maximum (peak) at about $\theta_1 = 0.12 \text{ m}^3/\text{m}^3$. Similar peaks have been reported in the literature. Most of these (Jackson, 1964; Philip, 1969; Scotter, 1976) are lower and occur at lower water contents, but those reported by Scotter (1974) are quite comparable. It is not clear at this point whether the peak has real physical meaning, associated with vapor transport "liquid islands," etc. (Philip and deVries, 1957; Scotter, 1976). The shape of the diffusivity curve seems to be more sensitive to errors in θ_1 at low than at high water contents. The gravimetric sorptivity (Fig. 4) and diffusivity functions (Fig. 5) were reasonably consistent with those for the gamma data, although they were based on less than half as many data point. The diffusivity curve in this case did not exhibit a relative maximum. The lowest water content was only $0.097 \text{ m}^3/\text{m}^3$.

The Pachappa sandy loam cores were taken out of a larger laboratory column on which alfalfa was grown for 2 years (see Dirksen and Huber, 1980).⁴ Just be-

⁴C. Dirksen and M. J. Huber. 1980. Interaction of alfalfa with matric and osmotic soil water potentials non-uniform in space and time. I. Experimental. (in preparation).

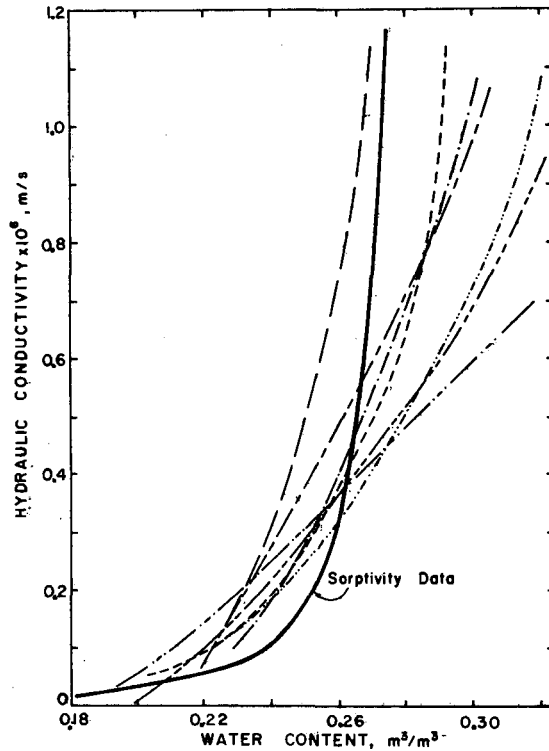


Fig. 6—Hydraulic conductivity function (solid line) of Pachappa sandy loam based on seven gravimetric sorptivity runs and simultaneous tensiometer measurements. Other lines are independent determinations on similar columns.

fore the coring, after the alfalfa plants had been removed, hydraulic conductivities of the larger column were determined by following drainage after saturation, while preventing evaporation. The four diffusivity values indicated in Fig. 5 were calculated from these data as the product of $k[\theta]$ and $dh/d\theta$ of the wetting characteristic, which was determined separately. The agreement was excellent, especially at the higher water contents where the hydraulic conductivity data were most accurate.

Pressure heads were also measured in the Pachappa sandy loam with the tensiometer shown in Fig. 2 during the seven sorptivity runs with gravimetric water content determinations. All pressure heads agreed well with the other data, except that the driest run at $\theta_1 = 0.097 \text{ m}^3/\text{m}^3$ did not quite reach a steady value in 12 min. Figure 6 shows the hydraulic conductivity function solely derived from these seven runs, which lasted from 6 to 12 min each. This function compares well with the other curves that were derived for seven columns before the alfalfa was planted. The latter measurements took weeks, however, and still yielded values only down to about $\theta = 0.20 \text{ m}^3/\text{m}^3$. The sorptivity measurements and simultaneous tensiometer measurements gave hydraulic conductivity values down to $\theta = 0.10 \text{ m}^3/\text{m}^3$.

Figure 7 gives $\log S^2$ vs. θ_1 data from a fine sand fraction. In contrast with results from the Pachappa sandy loam, there was a major discrepancy between the gamma data and the gravimetric data. Also, the scatter of the gamma data points was much larger. Figure 8 gives the diffusivity functions derived via the polynomial regressions in Fig. 7. Since diffusivity

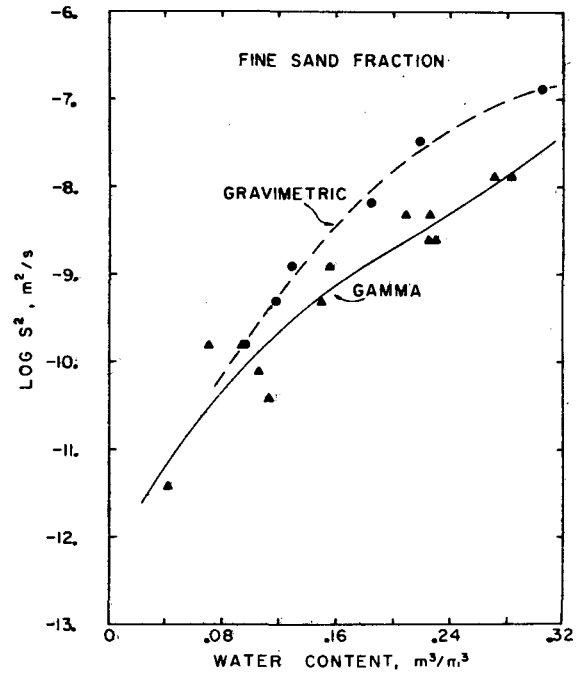


Fig. 7— $\log S^2$ vs. θ_1 , for fine sand fraction. Data points: \blacktriangle gamma attenuation, \bullet gravimetric. Lines show 3rd degree polynomial regressions.

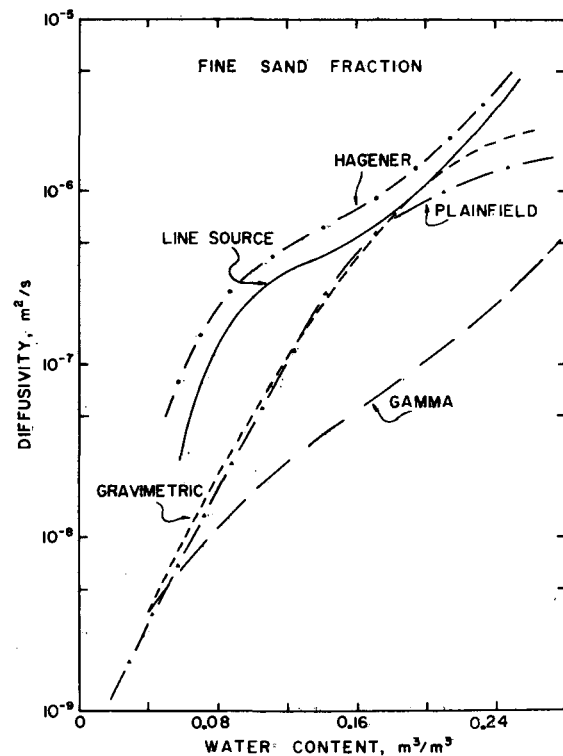


Fig. 8—Diffusivity functions of fine sand fraction derived from polynomial regressions in Fig. 7. Also shown are diffusivity functions from independent data on same sand (line source) and two other sands.

values generally increase with increasing coarseness of the soil texture, the gamma data were suspect. This was confirmed by the later gravimetric data and their comparison with other data. The fine sand fraction

was the same as that used in a laboratory model study of flow from buried line sources (Dirksen, 1978). The diffusivity function marked "line source" in Fig. 8 was derived from hydraulic conductivity and soil water retentivity functions obtained in that study. The gravimetric diffusivity agreed well with the line source data between $\theta = 0.16$ and $0.23 \text{ m}^3/\text{m}^3$. Moreover, the gravimetric data virtually coincides with diffusivity data for Plainfield sand (Typic Udipsamments) reported by Black et al. (1969). In turn, the line-source data coincided with the wet-side envelope of data for Hagener sand (cumuli Hapludolls) (Selim et al., 1970, Fig. 5a). Figure 8 shows the average of their data. Thus the gravimetric data compared well with other independent data, but the gamma data showed the sand much too wet, except at the very dry end. This must have been caused by a tightening of the sand pack during the (initial) wetting, which caused an increase in gamma ray attenuation. Such an increase cannot be distinguished from an increase due to water content. It was almost impossible to pack the fine sand tightly, especially at the surface. This was not a problem with the undisturbed cores. This shows clearly that gamma ray attenuation measurements will be in error if any changes, other than in the water content occur in the soil.

Figure 9 gives the diffusivity functions of three other soils, which were measured with the gamma beam on undisturbed field cores. One of these is Dateland fine sandy loam (Duric Haplargids) from the site of the U.S. Salinity Laboratory minimum leaching experiments on citrus at Tacna, Arizona (U. S. Salinity Laboratory Staff, 1977). It displays a plateau similar to that of the Pachappa sandy loam at about $\theta = 0.10 \text{ m}^3/\text{m}^3$. Earlier, three large cores (0.125 m diam, 0.9

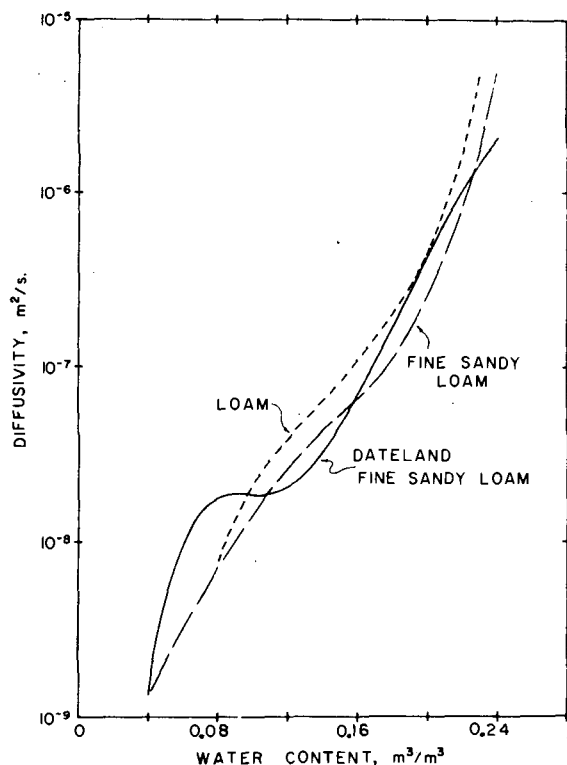


Fig. 9—Diffusivity functions of Dateland fine sandy loam and two other soils derived from sorptivity data.

m long) were taken near the same site and hydraulic conductivities were determined by imposing various steady fluxes until pressure heads became steady. The bars in Fig. 10 give the range of observed pressure heads for the various imposed fluxes. The solid line was drawn by sight. The diffusivity function of Fig. 9 was combined with a drying soil-water retentivity function to calculate the hydraulic conductivity curve, which is also shown in Fig. 10. A wetting soil-water retentivity function should have been used, but was not available. Since $d\theta/dh$ is smaller for wetting than drying at the lower pressure heads and larger at the higher pressure heads, the correction would swing the sorptivity curve towards the steady-state curve. Thus, the agreement between the two sets of data was excellent. Obtaining the steady-state data required over 6 months; the sorptivity data took about 4 hours. Moreover, the lowest water content of the steady-state data was only about $\theta = 0.17 \text{ m}^3/\text{m}^3$, whereas sorptivity data were taken down to $\theta = 0.04 \text{ m}^3/\text{m}^3$. The sorptivity data did not include the very wet end, because these measurements were made before the 6-rpm motor was put in use and an extra large syringe (Fig. 2) was made. The other two soils in Fig. 9 were taken from different sites; one was also a fine sandy loam soil, the other was more like a loam. Their diffusivity values were very similar, but no further measurements were made on these two soils.

DISCUSSION

Once the necessary apparatus has been assembled, the sorptivity method for measuring soil hydraulic property functions is very simple and fast. In sharp contrast with the potential-controlled boundary, the

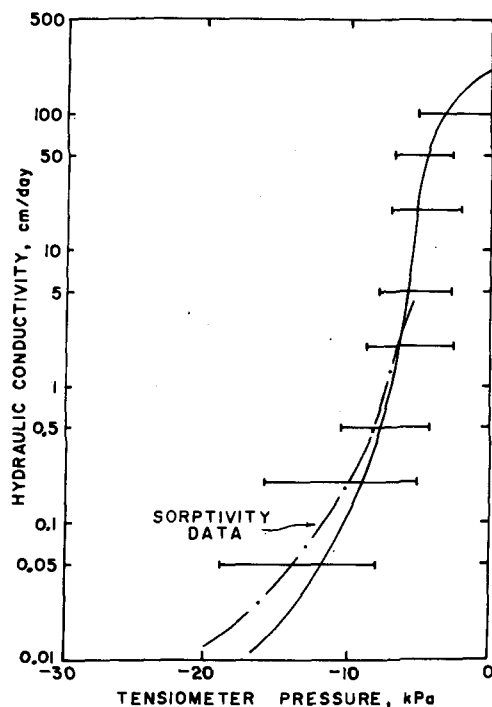


Fig. 10—Hydraulic conductivity function of Dateland fine sandy loam derived from sorptivity data. The solid line was drawn by sight through bars indicating range of pressure heads observed during steady flows.

flux-controlled boundary results in uniform wetting in nearly every run. The preparation of a new soil surface takes about 5 to 10 min. The column holder (Fig. 3) simplifies the gravimetric sampling enough that it can be done in about 15 sec, eliminating errors due to continued water movement and evaporation.

Initially, it seemed preferable to measure θ_1 by gamma attenuation while the sorptivity runs were in progress, but the results with the fine sand fraction showed clearly that gravimetric measurements were inherently more accurate. They measured the water content of the entire surface layer directly, rather than measuring just a small fraction indirectly and introducing errors due to electronics drift, counting statistics, calibration, and dissimilarity between dry and wet conditions. Also, gravimetric measurements require far less expensive equipment and they are faster (no reference counts). The gamma measurements, on the other hand, were useful in that they showed that the water contents stabilized after a few minutes, even at low water contents. Also, gamma results were obtained immediately, and the choice of sorptivities could be adjusted along the way. This feature could be retained with gravimetric measurements by using a microwave oven (Miller et al., 1974).

Except for the gamma data with the fine sand fraction, the data were all very consistent. The regression coefficients R^2 of the polynomial regressions of $\log S^2$ with respect to θ_1 ranged between 0.984 and 0.999. Whenever the results were checked against independent data the agreement was good, certainly for hydraulic diffusivities and conductivities. It is not possible to give absolute accuracies, because no independent check was made on the same soil columns, and so they were also subject to unknown errors. With the change to the flux-controlled boundary, S has become the independent variable and the errors are mainly in θ_1 . Therefore, least-squares regressions were also made of θ_1 on $\log S^2$. However, the resulting polynomial $\theta_1 = f[\log S^2]$ could not be reversed to $\log S^2 = f^{-1}[\theta_1]$ with equations for "reversion of series" (Hodgman et al., 1959), in order to perform the differentiation in Eq. [3]. Since the two types of regressions should yield essentially the same results (judging from the regression coefficients), this was not pursued further. However, if more variability in the data is encountered, it would be desirable to find a convenient way to obtain $D[\theta]$ via the proper polynomial regression.

The gravimetric sampling could be adapted for in situ measurements. However, since sorptivity is measured in only a thin layer of soil, the sorptivity method is unsuitable for conditions which require in situ measurements because of large scale effects such as cracks and nonuniform soil profiles. This method is most attractive, however, for determining the spatial variability of soils that are uniform on a small scale. This can be done on undisturbed cores in the laboratory rather than sacrificing the convenience and accuracy of working in the laboratory for in situ measurements. Moreover, in the field, one can only make measurements in the range wetter than field conditions, and the initial water contents may not be all the same. In the laboratory, cores can first be dried uniformly and the hydraulic property functions determined over the entire water-content range. Since the measurements

take only a short time and water is continually being pushed (sucked) out of the syringe through the ceramic plate, the low water potentials in the soil have no time to cause air-entry normally encountered beyond the tensiometer range. To derive hydraulic conductivities, the retentivity function must be determined separately at low water contents, for example with a pressure membrane apparatus. The sorptivity at saturation may be easier to measure without using the cam and ceramic plate, by inundating the soil surface with a thin layer of water and regulating the water level with a device, such as that described by Shalhevet et al. (1976), or by using the procedures of Talsma (1969). The sorptivity method was not tested on swelling soil, but the gravimetric vs. gamma attenuation results for the fine sand fraction suggested that acceptable results may be obtainable with slightly swelling soils by gravimetric sampling. The same principles have already been used to determine the diffusivity for desorption of clay suspensions in terms of the void ratio (Smiles and Harvey, 1973; Smiles, 1976).

The sorptivity method combines the speed of transient methods with the resolution and accuracy of steady-state methods over the entire water content range. Steady-state methods are prohibitively slow, and, therefore, often also inaccurate, at the lower water contents. Even at higher water contents, steady-state methods often take months. Sorptivity measurements are equally easy and fast at low and high water contents. Probably, the most similar and most extensively used method is the crust method (Bouma et al., 1971; Baker, 1977). It is considered to have reached a steady state only after the flux has been constant for at least 4 hours, and it takes more time to make a new crust than to make a sorptivity measurement (Bouma and Denning, 1972). Also, it seems to work only for pressure heads higher than about -10 kPa. Gardner and Miklich (1962) stated that with their constant-flux method 3 weeks was usually sufficient to obtain data for the tensiometer range. Even one transient run usually takes more time than a series of very short sorptivity runs. For example, Passioura (1976) estimates that his simplified one-step pressure-plate-outflow method takes about 5 weeks in a 6-cm long column and 1 week in a 3-cm long column. To cover the entire water content range, he advocated running one long and one short column. Instantaneous profile methods (Klute, 1972) are also very time-consuming and laborious, while they still can cover only the wetter part of the water content range. With sufficient instrumentation they can measure hydraulic conductivities of nonuniform soil profiles.

The sorptivity method has a high degree of resolution, since water contents are determined accurately in an area where they change little with distance. In other methods, for example that of Bruce and Klute (1956), one must obtain all the information for the lower water contents by sampling in an area where the water content varies fast with distance. This is inherently inaccurate. With the modification as proposed by Whisler et al. (1968), and further tested by Selim et al. (1970), the change in water content with time is measured at a fixed position with gamma attenuation. The disadvantages of gamma measurements have already been listed. A method equivalent to the Bruce and Klute (1956) method — but then for

evaporation rather than infiltration — is the forced air method introduced by Arya et al. (1975). It is simple and fast and is beginning to be used routinely (Ehlers, 1976; Bouma, 1977). It has the same disadvantage of low resolution. The sampling takes much longer than with the sorptivity method, increasing the opportunity for water redistribution and evaporation. In addition, the conditions are distinctly nonisothermal. The effect of this on the results needs further investigation.

Other methods that can compare with the sorptivity method in terms of speed are those that calculate the hydraulic conductivity function from the soil-water-retentivity function (Brutsaert, 1967; Klute, 1972), and possibly those that fit a prefixed functional relationship (Ahuja and Swartzendruber, 1972; Stroosnijder and Bolt, 1974; Miller and Bresler, 1977; Bressler et al., 1978). Both methods leave something to be desired in accuracy. The calculation methods generally use one measured value (often the saturated hydraulic conductivity) with which the calculated value is matched, and the rest of the function is then adjusted in the same proportion. These matching factors have varied over three orders of magnitude, without being predictable. Often, the adjusted functions still differ appreciably from directly measured values. The fitting methods use several relatively simple measurements, such as advance of wetting front, θ_1 at saturation, and air-entry value. Air-entry measurements are inherently inaccurate and many soils do not follow the prefixed relationships. The sorptivity method can, in principle, determine any kind of relationship. For some soils, this may require a few extra measurements, but these measurements take little time.

In conclusion, this improved sorptivity method seems to be the only available method that is fast and simple enough to be used routinely and on a large scale, while at the same time being capable of establishing any kind of functional relationships based on actual measurements of a high accuracy. The method was already shown to have a sound theoretical basis (Dirksen, 1975a, b). This study shows that with the flux-controlled boundary and gravimetric water content determinations, it is experimentally sound, as well.

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