

SOIL SCIENCE SOCIETY OF AMERICA JOURNAL

VOL. 45

MAY-JUNE 1981

No. 3

DIVISION S-1—SOIL PHYSICS

Hydrodynamic Dispersion During Constant Rate Absorption of Water by Soil¹

D. E. SMILES, K. M. PERROUX, S. J. ZEGELIN, AND P. A. C. RAATS²

ABSTRACT

An analysis of hydrodynamic dispersion accompanying constant flux absorption of KCl solution by an initially relatively dry soil, is developed for the case when the hydrodynamic dispersion coefficient is pore water velocity-independent. It is shown that in this process both the water content and the soil water salt concentration are uniquely defined by $\theta(X,T)$ and $C(X,T)$, where $X = v_0 x$ and $T = v_0^2 t$ are space- and time-like coordinates, and v_0 is the constant surface flux of water.

Quasi-analytical methods based on the flux-concentration relation predict $\theta(X,T)$ while an error-function solution, based on a material coordinate Q labeling parcels of water, predicts the salt profile.

The analysis is demonstrated using a chemically inert sandy soil. The results show that during transient, unsaturated flow a simple piston-flow model described the process over a range of water contents. The method may be extended to explore dispersion in structured and chemically reactive soils.

Additional Index Words: piston-flow model, material coordinates.

Smiles, D. E., K. M. Perroux, S. J. Zegelin, and P. A. C. Raats. 1981. Hydrodynamic dispersion during constant rate absorption of water by soil. *Soil Sci. Soc. Am. J.* 45:453-458.

SMILES ET AL. (1978) and Smiles and Philip (1978) examined the hydrodynamic dispersion of potassium chloride during absorption of water into an initially uniform, horizontal soil column. The experiments were performed with the soil surface maintained at constant water potential $\Psi \sim -0.01$ m. It was found that both the soil water content, θ , and the solution salt concentration, C , preserved similarity in terms of distance x divided by the square root of time, t . These observations implied that, for the flow regime

imposed, the dispersion coefficient D_s was effectively independent of v , the volumetric flux of the water. In addition, the movement of the salt was found to be explicable in terms of a model based on piston-like displacement of the initial water by the invading water. It is important to note that this behavior is an immediate consequence of transverse equilibration of solute concentration, even if there is, in reality, a wide range of pore water velocities. Our use of the piston-like model must be interpreted within this context.

Elrick et al. (1979), examining both sorption and infiltration into a clay-loam, found the piston model less satisfactory; they invoked an immobile water component to explain the failure of the "salt front" to correspond with the notional plane that would separate initial and invading water, were the displacement of water complete.

In aggregated and reactive soils, one would expect such a phenomenon although its magnitude must depend on the reaction rates, on the relative fluid velocities within and around the aggregates, and on the aggregate dimensions. Passioura (1971) and Passioura and Rose (1971) discussed the aggregate problem for the case of saturated flow, but the effects of aggregation and immobile water in unsaturated soils still appear to present problems. It is therefore valuable to obtain reliable data on simple systems to establish the range of validity of the piston-flow model and to extend the analytical and experimental techniques in ways that may permit more critical experiments in unsaturated soil.

The group of experiments described in this paper explore the process of hydrodynamic dispersion during transient flow in substantially unsaturated columns at whose surface water is supplied at a constant rate v_0 . At the core of the problem lies the prediction of the transient water-content profile. White et al. (1979) presented an analysis using the space- and time-like variables $X = v_0 x$ and $T = v_0^2 t$, which permit generalization of the problem to encompass all values of v_0 . We superimpose on this analysis a dispersion equation with a dispersion coefficient, $D_s(\theta)$, that may

¹Contribution from C.S.I.R.O., Australia. Received 13 Aug. 1980. Accepted 26 Jan. 1981.

²Chief, Experimental Officer, Technical Assistant, and Visiting Research Fellow, respectively. C.S.I.R.O. Division of Environmental Mechanics, P.O. Box 821, Canberra City, ACT 2601, Australia. Dr. Raats' permanent address is: Institute for Soil Fertility, P.O. Box 30003, Haren-Gr., The Netherlands.

be water-content dependent but is velocity-independent.

The analysis shows how space- and time-like variables, which reveal the essential patterns of water flow during constant flux absorption, also permit simplification of the salt flow problem, provided some physically reasonable simplifying assumptions relating to salt flow are made. The analysis also demonstrates the analytical simplification that results when the dispersion equation is recast in terms of a material coordinate, Q , labeling parcels of water.

In addition, the analysis provides a potentially useful technique for studying dispersion during transient water flow where the soil surface is unsaturated. In this respect, the constant-flux boundary condition is relatively simply imposed. The soil surface water potential, which critically affects the water-content profile, is not affected by transmission properties of sintered plates or the soil-plate interface. These problems greatly complicate the realization of the constant-potential boundary condition, at substantial, negative-supply potential.

THEORY

Water Movement

Experiments and theory related to constant-flux absorption of water by a horizontal column of soil with initially-uniform water content were presented by White et al. (1979). The flow of water may be described in this process by the balance of mass,

$$\partial\theta/\partial t = -\partial v/\partial x, \quad [1]$$

and Darcy's Law,

$$v = -D(\theta) \partial\theta/\partial x. \quad [2]$$

In these equations, t is time, x is the horizontal coordinate, θ is the volumetric water content, v is the volumetric flux of the soil water, and D is the water-content dependent soil-water diffusivity.

The initial and boundary conditions are

$$\theta = \theta_0, \quad t = 0, \quad x \geq 0 \quad \text{and} \quad [3]$$

$$-D(\theta) \partial\theta/\partial x = v_0, \quad t > 0, \quad x = 0. \quad [4]$$

It is useful to introduce the space- and time-like variables defined by

$$X = v_0 x \quad \text{and} \quad [5]$$

$$T = v_0^2 t, \quad [6]$$

and the reduced flux F given by

$$F = v/v_0. \quad [7]$$

Then, introducing X , T , and F into Eq. [1]-[4] gives

$$\partial\theta/\partial T = -\partial F/\partial X, \quad [8]$$

$$F = -D(\theta) \partial\theta/\partial X, \quad [9]$$

$$\theta = \theta_0, \quad T = 0, \quad X \geq 0, \quad \text{and} \quad [10]$$

$$-D(\theta) \partial\theta/\partial X = 1, \quad T > 0, \quad X = 0. \quad [11]$$

As White et al. (1979) point out, Eq. [8]-[11] imply that water-content profiles $\theta(X)$ at any value of T , are unique. In particular, the surface-water content θ_0 and the "position of the wetting front" are uniquely defined by T .

Introducing the reduced water content Θ defined by

$$\Theta = (\theta - \theta_0)/[\theta_0(T) - \theta_0], \quad [12]$$

White et al. (1979) showed that the time-independent relationship between F and Θ given by

$$F = \Theta^{2-1/\alpha} \quad [13]$$

gives satisfactory prediction of $\theta(X, T)$, at least if T is not too small. In the present work the even simpler relationship

$$F = \Theta \quad [14]$$

is used. Further explanation of the analysis is given by White et al. (1979), and White (1979) discusses in more detail the relationships between F and Θ .

Salt Movement

Following Smiles et al. (1978), we write the equation describing the transport of the salt during unsaturated water flow in soil as

$$\frac{\partial(\theta C)}{\partial t} = \frac{\partial}{\partial x} \left[D_s(\theta) \frac{\partial C}{\partial x} \right] - \frac{\partial(vC)}{\partial x}, \quad [15]$$

which we seek to solve for the initial and boundary conditions

$$C = C_0, \quad t = 0, \quad x \geq 0, \quad [16]$$

$$C = C_0, \quad t > 0, \quad x = 0. \quad [17]$$

If we introduce the reduced variables X and T defined by Eq. [5] and [6] in Eq. [15], expand the derivatives of the products, ∂C and $F C$, use Eq. [8] to cancel two terms, and divide by θ , we obtain

$$\frac{\partial C}{\partial T} + \frac{F}{\theta} \frac{\partial C}{\partial X} = \frac{1}{\theta} \frac{\partial}{\partial X} \left(D_s(\theta) \frac{\partial C}{\partial X} \right). \quad [18]$$

In addition, the initial and boundary conditions [16] and [17] become

$$C = C_0, \quad T = 0, \quad X \geq 0 \quad [19]$$

$$C = C_0, \quad T > 0, \quad X = 0. \quad [20]$$

Because F and θ are functions only of X and T , the solution $C(X, T)$ of Eq. [18], subject to [19] and [20], will be unique if $D_s(\theta)$ is velocity-independent.

The convective term on the left hand side of the equation makes the solution of the set of Eq. [18] to [20] very difficult in terms of analytical functions. We therefore reconsider the solute transport in terms of a material coordinate labeling parcels of water. In the first instance, we base our approach on the assumption that piston-flow of water occurs and individual parcels of water retain their nearest neighbors, so the initially-present water is displaced in its entirety by the invading solution.

To describe the convective transport of the solutes in such a framework one wishes to know the location X of individual parcels of water at successive times T . The function $Q(X, T)$ defined by

$$\theta = \partial Q/\partial X \quad \text{and} \quad [21]$$

$$F = -\partial Q/\partial T, \quad [22]$$

effectively labels all members of a collection of parcels of water participating in a one-dimensional flow. Substitution of Eq. [21] and [22] in Eq. [8] shows that this definition of Q is valid.

Equations [21] and [22] define Q up to an arbitrary constant. It is convenient to define this constant by setting the "parcel function" Q at zero at the soil surface at $T = 0$, i.e.,

$$Q[T = 0, X = 0] = 0. \quad [23]$$

From the defining Eq. [21] and [22] and from Eq. [23] it follows that Q at any time T and distance X may be written as

$$Q(X, T) = \int_0^T \left(\frac{\partial Q}{\partial T} \right)_{x=0} dT + \int_0^X \left(\frac{\partial Q}{\partial X} \right)_T dX. \quad [24]$$

That is,

$$Q(X, T) = \int_0^X \theta dX - T. \quad [25]$$

The first term on the right side of Eq. [25] represents the cumulative amount of water in the profile between the surface and X ; the second term, T , is the total cumulative inflow, i.e.,

$$T = \int_{\theta_n}^{\theta_0} X d\theta \quad [26]$$

Equation [25] implies that the element of water initially at the soil surface will be found at a later time T at $X = X^*$ given by

$$\int_0^{X^*} \theta dX = T \quad [27]$$

Alternatively, by changing the independent variable of the integral in Eq. [25] and substituting for T from Eq. [26], we obtain

$$Q(X, T) = \theta X - \int_{\theta_n}^{\theta} X d\theta \quad [28]$$

In terms of Eq. [28], the coordinate X^* of the parcel of water labeled $Q = 0$ will be given by

$$\theta^* X^* = \int_{\theta_n}^{\theta^*} X d\theta \quad [29]$$

The content of Eq. [25] and [27] and Eq. [28] and [29] is shown diagrammatically in Fig. 1a, b, and c. It should be noted that the formal definition of $Q(X, T)$ presented as Eq. [28] is analogous to the function $g(\lambda)$ defined by Eq. [12] of Smiles et al. (1978) and Eq. [6] of Smiles and Philip (1978). In addition, it should be recognized that if the piston-flow model is valid, then at any time T , the center-of-mass of the salt front, X' , given by

$$\int_{X'}^0 \theta(C_0 - C) dX = \int_{\infty}^{X'} \theta(C - C_n) dX \quad [30]$$

must coincide with $Q = 0$. Note that X' so defined does not correspond exactly with a similar position defined by Smiles et al. (1978). That definition was incorrect, although the error incurred in its use is not significant.

The differentiation and manipulation of the functions $C(X, T)$ and $X(Q, T)$, together with Eq. [21] and [22], permit us to rewrite Eq. [27] as

$$\left(\frac{\partial C}{\partial T}\right)_a = \frac{\partial}{\partial Q} \left[\theta D_s(\theta) \frac{\partial C}{\partial Q} \right] \quad [31]$$

For the data to be presented in this paper, it is reasonable to assume that in the region where dispersion is significant (close to $Q = 0$), the water content θ depends on T but at fixed time T varies only slightly with Q . If then we assume (Smiles and Philip, 1978)

$$D_s(\theta) = \theta D_m \quad [32]$$

where D_m is the molecular diffusion coefficient of the salt in water, then Eq. [31] may be written

$$\frac{\partial C}{\partial \tau} = \frac{\partial^2 C}{\partial Q^2} \quad [33]$$

with the time-like coordinate τ given by

$$\tau = \int_0^T (\theta^*)^2 D_m dT \quad [34]$$

This integral, [34], requires that we determine the water content at $Q = 0$ as a function of T .

Then ignoring complications near $X = 0$ and $T = 0$, the solution of Eq. [33] for the step change in solution concentration at $Q = 0$ implied in Eq. [28] and [29] can be written

$$\frac{C - C_0}{C_n - C_0} = \frac{1}{2} \left[\operatorname{erf} \left(\frac{Q\tau^{-1/2}}{2} \right) + 1 \right] \quad [35]$$

To summarize: if the hydrodynamic dispersion coefficient is velocity-independent, then for dispersion accompanying absorption of water by soil subject to conditions [3] and [4], and [16] and [17], both the water-content profiles $\theta(X, T)$, and the solution salt-concentration profiles $C(X, T)$ are unique. Furthermore, if the piston-like model of water flow is valid, we may introduce a material coordinate Q whose origin must coincide with the center-of-mass of the advancing salt front. The Q -coordinate simplifies the dispersion equation to permit

calculation of the salt profile using conventional error-function analysis.

EXPERIMENTAL

The experiments were designed to realize the conditions on water movement set out as Eq. [3] and [4] and salt movement defined by Eq. [16] and [17]. They were intended to demonstrate the use of the constant-flux boundary condition as a means for examining hydrodynamic dispersion in an unsaturated soil. They demonstrate the use of material coordinates during unsteady, saturated water flow. They also test the following propositions for a simple system: (i) that $D_s(\theta)$ is practically velocity-independent; (ii) that piston-like behavior of the water occurs; and (iii) that the error-function solution in terms of Q and τ gives acceptable prediction of the salt profiles.

Bungendore fine sand was used in all experiments. It revealed $D(\theta)$ and the wetting moisture characteristic $\Psi(\theta)$ shown in Fig. 2 and 3. $\Psi(\theta)$ was determined by equilibrating a vertical column of the sand with its base in a reservoir of water or 0.1M KCl solution. The $D(\theta)$ relation was measured using the method of Bruce and Klute (1956).

The absorption experiments were performed in sectioned plexiglass columns forming a cylinder of internal diameter 19.3 mm. A similar column is shown by Elrick et al. (1979). The columns were packed to a bulk density of about $1.6 \times 10^3 \text{ kg m}^{-3}$ with sand moistened with 0.1M KCl to produce an initial volumetric solution content of approximately 0.1. One molar KCl solution was then supplied to one end of the column, through a porosity-2 sintered glass plate, using a constant-rate syringe pump.

At the conclusion of each experiment, columns were sectioned, the solution content determined by oven-drying at 105°C, and the solution-salt concentration measured after dilution using a Radiometer DCM3 conductivity meter. Dilution varied within

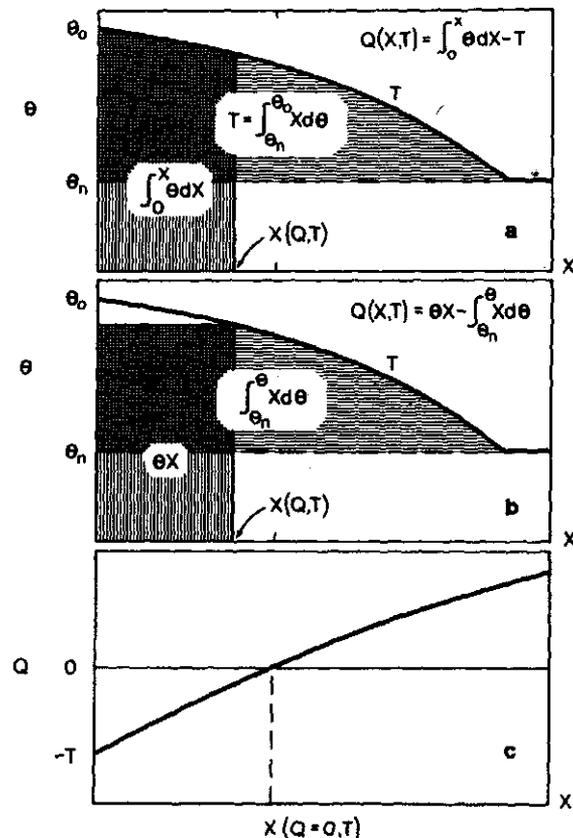


Fig. 1—Diagrams demonstrating the physical significance of the material coordinate Q labeling parcels of water during absorption into a soil. Fig. 1a shows how $Q(X)_r$ is defined by Eq. [25]; Fig. 1b demonstrates Eq. [28]; Fig. 1c shows $Q(X)_r$. $X(Q = 0, T)$ is X^* defined by Eq. [27] and [29].

Table 1—Summary of the important features of the constant-flux absorption and dispersion experiments.

$T, \text{m}^2/\text{sec}$	$v_o, \text{m}/\text{sec}$	t, sec
1.00×10^{-4}	3.283×10^{-7}	9.348×10^4
	4.377×10^{-7}	5.232×10^4
	6.566×10^{-7}	3.322×10^4
	8.754×10^{-7}	1.308×10^4
5.40×10^{-4}	8.754×10^{-7}	7.050×10^4
	1.313×10^{-6}	3.132×10^4
	1.751×10^{-6}	1.758×10^4
	3.283×10^{-6}	5.010×10^3
3.97×10^{-7}	3.939×10^{-6}	2.556×10^4
	5.253×10^{-6}	1.440×10^4
	7.879×10^{-6}	6.420×10^3
	1.051×10^{-5}	3.600×10^3

$\theta_n = 0.1 \quad C_n = 0.1M \text{ KCl} \quad C_o = 1M \text{ KCl}$

the range 1:5 to 1:15, depending on solution content and sample size.

Three sets of experiments were performed. Different experiments in each set had different imposed surface fluxes but were terminated at times such that the set was parametrized by a particular value of $v_o t = T$. The theory of White et al. (1979) then requires that each set reveals a unique water-content profile $\theta(X)$. The observed salt profiles for each set were also presented in terms of this space-like coordinate, as $C(X)_T$. Table 1 summarizes the important features of all experiments. The profiles of water content and solution-salt concentration are presented in Fig. 4, 5, and 6.

The experimental variability in the $\theta(X)_T$ data is evident in the scatter of points on the figures. Because of small variations in measured C_n and C_o between experiments, the salt profiles have been normalized in terms of C_n and C_o for each experiment, and are presented in terms of the variable $(C-C_o)/(C_n-C_o)$. The experimental variability is therefore obscured somewhat, although we present standard deviations in our estimates of C_n and C_o for each set of data. We note here also that C was effectively constant close to $X = 0$, justifying our use of the salt boundary condition Eq. [17] rather than the alternative "flux boundary condition" discussed, for example, by Parlange and Starr (1975) and Smiles et al. (1978).

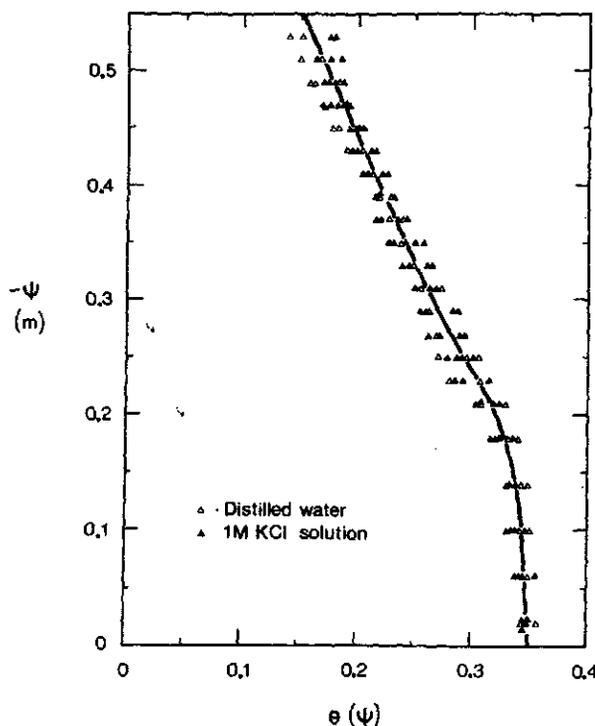


Fig. 2—Wetting water content $-\psi$ water potential relation, $\theta(\psi)$, of Bungendore fine sand with $\theta_n = 0.1$.

DISCUSSION

Water-Content Profiles

As we point out above, the prediction of the water-content profile is central to any study of hydrodynamic dispersion during transient flow. These profiles were calculated by the theory of White et al. (1979), using the material properties shown in Fig. 2 and 3 and the approximate flux-concentration relation $F = \Theta = (\theta - \theta_n)/(\theta_o - \theta_n)$. The curves were predicted from first principles and have not been matched to experimentally observed values of θ_o ; they are shown as smooth curves superimposed on the experimental $\theta(X)_T$ data of Fig. 4, 5, and 6. As with other comparisons using this approach, the correspondence between prediction and observation is excellent. It then follows that we have in principle a useful tool for studying dispersion, since we can, by suitably controlling v_o and t , achieve a great range of unsaturated, unsteady water-content profiles. By implication we could also design experiments in aggregated soils that might quite precisely partition water flow within and between the aggregates.

Soluble Salt Profiles

The experimental $C(X)_T$ data presented in Fig. 4, 5, and 6 confirm the unifying role of the variables X and T . The concentration profiles collapse to a single curve for each set of experimental data parametrized by a particular value of T , despite the wide ranges of v_o and times at which the experiments were terminated. This behavior confirms, for these experiments, the assumption that D_s is practically velocity-independent. This observation is consistent with statistical models of Saffman (1959) and Pfannkuch (1963) in so far as our Péclet numbers, Pe , never exceeded 2. Pe was calculated using the equation

$$Pe = vl/\theta D_m \quad [36]$$

in which the characteristic length, l , was identified

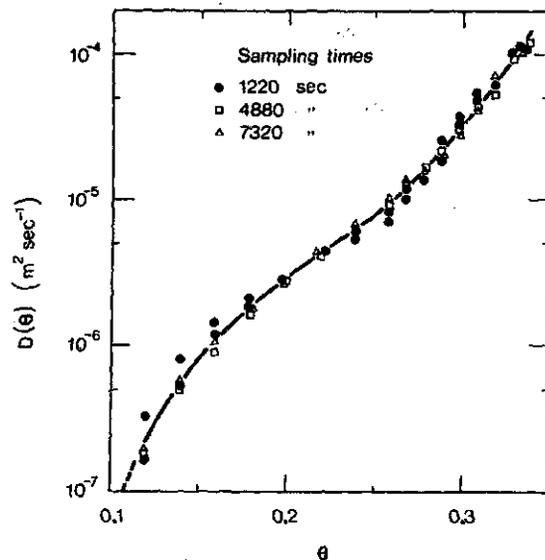


Fig. 3—Soil-water diffusivity, $D(\theta)$, for the fine sand samples used. This function was determined using the method of Bruce and Klute (1956) with $\psi_o = -0.02 \text{ m}$.

with the modal particle size (100 μm), and D_m , the diffusion coefficient for KCl in water, was taken as $1.9 \times 10^{-9} \text{ m}^2 \text{ sec}^{-1}$.

In fact it is difficult to see this experimental method permitting critical examination of the velocity-dependence of D_s , because for v_0 large, the experiments are necessarily of short duration and the accurate resolution of the salt profile correspondingly more difficult. This is not a very serious qualification on the method, however, if one accepts the contention of Smiles and Philip (1978) that, in field situations involving sustained, unsaturated flow, Péclet numbers are unlikely to exceed unity for any significant length of time.

Piston-like Behavior of Absorbed Water

In the $\theta(X)_T$ diagrams we identify $Q = 0$ as X^* defined by Eq. [29]. The location X^* and the location of the center-of-mass of the salt front, X' , defined by Eq. [30], are shown as vertical dashed lines in Fig. 4, 5, and 6. It will be observed that, despite the fact that the initial water-content is quite large, and that each of the three sets of water content profiles represent results of transient flows in substantially unsaturated soils, these planes coincide within experimental error. We therefore conclude that the piston-flow notion again provides a useful basic model for

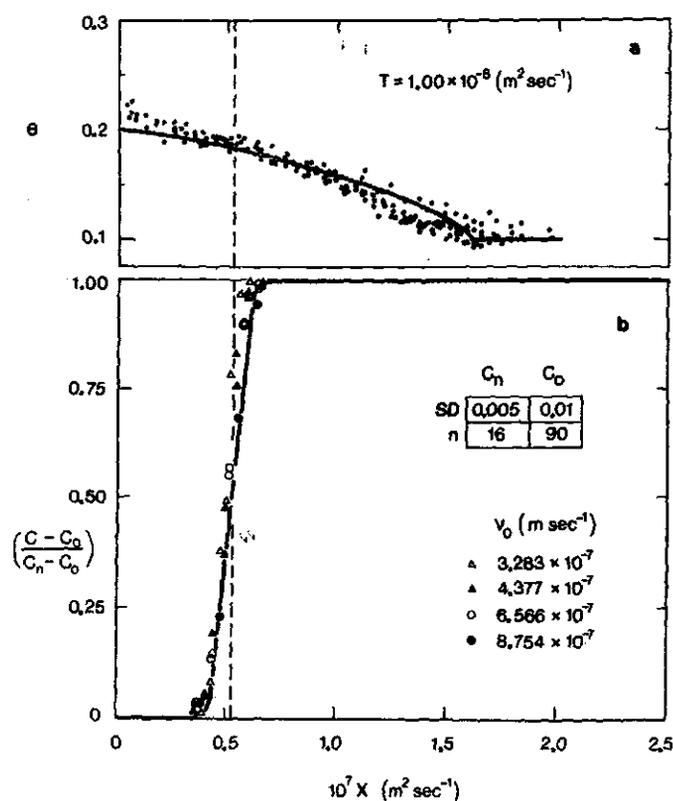


Fig. 4—Experimental water-content (a) and solution salt-concentration (b) profiles presented in reduced form $\theta(X)$ and $(C-C_0)/(C_n-C_0)$ vs. X for the set of experiments parametrized by $T = 1.0 \times 10^{-8} \text{ m}^2 \text{ sec}^{-1}$. The vertical broken lines identify the planes X^* and X' defined in the text. The curves on both $\theta(X)$ and $C(X)$ diagrams are predicted according to the theory, the latter using $D_s = 1.9 \times 10^{-9} \text{ m}^2 \text{ sec}^{-1}$.

salt movement accompanying transient water-flow in this unsaturated soil.

Calculation of the Salt Profile

The calculated salt-concentration profiles are shown as smooth curves in Fig. 4, 5, and 6. They were derived for particular values of T by evaluating τ at that value, using Eq. [34] and $\theta^*(T)$, determined by solving the water-flow equation. The reduced concentration $(C-C_0)/(C_n-C_0)$ vs. $Q|_T$ was then recast in terms of $X|_T$ using Eq. [25]. The predicted curves clearly correspond well with data for each experimental set and confirm that the approach is valid.

These observations must, however, be extended to other systems with care. Strictly speaking the diffusivity θD_s in Eq. [31] is a function not only of T but also of Q , and it contains information on the evolutionary history of the water profiles $\theta(X)_T$ within which the salt profiles evolved. In some circumstances, this Q -dependence of θ close to $Q = 0$ may not be relatively small. It might then be necessary to solve Eq. [31] numerically, recognizing in full the dependence of θ on Q and T .

In the context of the evolution of the $\theta(X)_T$ profiles, it is worth noting that as θ^* increases with increasing T , the continuity equation for the salt works against the process of dispersion in space. This phenomenon is not observed in Q -space.

In addition, we note that since τ is the time integral of the dispersion coefficient up to the current time T , it recognizes the evolutionary history of the current salt profile. We can compare τ/T with the dispersion coefficient, $(\theta^*)^2 D_m$, at the current time T , in terms of the $\tau/[(\theta^*)^2 D_m T]$. This ratio is initially small but increases towards unity with increasing time T ; thus

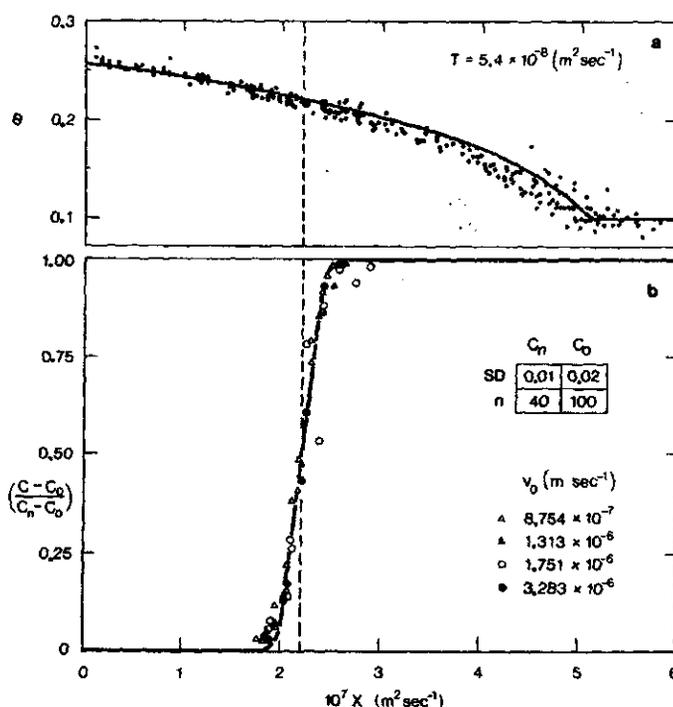


Fig. 5—Moisture (a) and salt-concentration (b) profiles similar to those shown in Fig. 3 but parametrized by $T = 5.4 \times 10^{-8} \text{ m}^2 \text{ sec}^{-1}$.

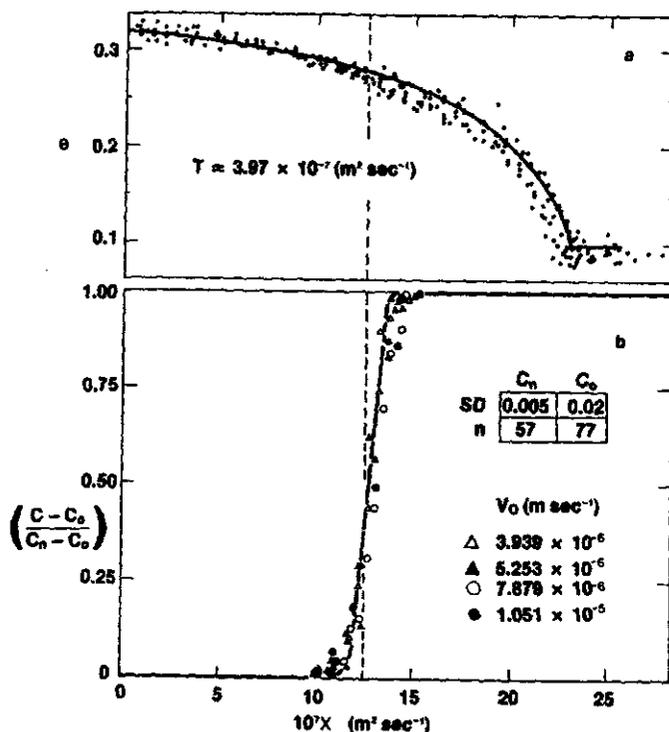


Fig. 6—Moisture (a) and salt-concentration (b) profiles similar to those shown in Fig. 4 but parametrized by $T = 3.97 \times 10^{-7} \text{ m}^2 \text{ sec}^{-1}$.

for $T = 5.4 \times 10^{-8} \text{ m}^2 \text{ sec}^{-1}$ the ratio is approximately 0.71, and it has increased to 0.79 when $T = 3.9 \times 10^{-7} \text{ m}^2 \text{ sec}^{-1}$.

Finally, we observe that $Q(X, T)$ is introduced in this paper on the assumption that the piston-flow model is valid. The presence of an immobile "layer" of water represented by θ_i , that is, inaccessible to the salt (for example, by virtue of anion exclusion), presents only a minor complication, provided that θ_i is constant. The complications that arise when θ_i is dependent on the ambient salt concentration C in a chemically reactive soil are being explored at present.

CONCLUSION

The variables $X = v_0 t$ and $T = v_0^2 t$ yield a common family of water-content profiles during constant-flux absorption of water into an initially-uniform soil. In particular, $\theta(X)_T$ is unique for all values of the flux v_0 .

If the dispersion coefficient is velocity-independent, then these variables also reduce the salt profiles to a single family when the absorbed solution has a salt concentration that differs from that originally present in the soil.

As a result, the calculation of dispersion during constant-flux absorption, and the early stages of constant-flux infiltration, is simplified to the extent that $C(X)_T$ is unique irrespective of v_0 . The calculation of dispersion is then simplified further if one introduces a material coordinate Q labeling parcels of water participating in the convective flow, and if one makes some physically reasonable assumptions about the water-content dependence of the dispersion coefficient.

The method may provide useful insights into the nature of hydrodynamic dispersion in unsaturated soil during transient flow. In the case of the experiments described here on a very simple material it confirms that the piston-flow model is valid during transient flow at relatively low soil-water contents. The method may also provide a useful basis for study of the process in structured soils and when chemical reaction is important; this is because the constant-flux boundary condition provides better control of the hydrodynamics than is possible using a constant (negative) potential boundary condition. The latter boundary condition is often difficult to realize because of its dependence on the transmission properties of the water-supply membrane, and the membrane-soil surface.

ACKNOWLEDGMENTS

The authors wish to acknowledge helpful discussion with Drs. B. E. Clothier, J. R. Philip, and I. White of this Division and J. H. Knight of the C.S.I.R.O. Division of Mathematics and Statistics. Financial assistance from the Australian Water Resources Council is also acknowledged.

LITERATURE CITED

1. Bruce, R. R., and A. Klute. 1956. The measurement of soil water diffusivity. *Soil Sci. Soc. Am. Proc.* 20:458-462.
2. Elrick, D. E., K. B. Laryea, and P. H. Groenevelt. 1979. Hydrodynamic dispersion during infiltration of water into soil. *Soil Sci. Soc. Am. J.* 43:856-865.
3. Parlange, J.-Y., and J. L. Starr. 1975. Linear dispersion in finite columns. *Soil Sci. Soc. Am. J.* 39:817-819.
4. Passioura, J. B. 1971. Hydrodynamic dispersion in aggregated media. 1. Theory. *Soil Sci.* 111:339-344.
5. Passioura, J. B., and D. A. Rose. 1971. Hydrodynamic dispersion in aggregated media. 2. Effects of velocity and aggregate size. *Soil Sci.* 111:345-351.
6. Pfannkuch, H. O. 1963. Contribution à l'étude des déplacements des fluides miscibles dans un milieu poreux. *Rev. Inst. Fr. Petrol.* 18:215-270.
7. Saffman, P. G. 1959. Dispersion due to molecular diffusion and macroscopic mixing in flow through a network of capillaries. *J. Fluid Mech.* 7:194-208.
8. Smiles, D. E., and J. R. Philip. 1978. Solute transport during absorption of water by soil: laboratory studies and their practical implications. *Soil Sci. Soc. Am. J.* 42:537-544.
9. Smiles, D. E., J. R. Philip, J. H. Knight, and D. E. Elrick. 1978. Hydrodynamic dispersion during absorption of water by soil. *Soil Sci. Soc. Am. J.* 42:229-234.
10. White, I. 1979. Measured and approximate flux-concentration relations for absorption of water by soil. *Soil Sci. Soc. Am. J.* 43:1074-1080.
11. White, I., D. E. Smiles, and K. M. Perroux. 1979. Absorption of water by soil: the constant flux boundary condition. *Soil Sci. Soc. Am. J.* 43:659-664.