MEASUREMENT AND EVALUATION OF SOIL WATER TRANSMISSION COEFFICIENTS IN SOME HAWAIIAN LATOSOLS

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INTRODUCTION

The efficient use of available water for increased crop production is of primary concern in agriculture. Since water is generally regarded as the media of transport within the soil system, the principles underlying moisture retention and flow must be placed under careful scrutiny. Prediction, and ultimately control, of water losses from infiltration, evaporation, evapotranspiration, and drainage are only possible once the principles involved are thoroughly understood.

The parameters that describe the dynamic moisture regime in the soil are the transmission coefficients, capillary conductivity, $K(\theta)$, and soil water diffusivity, $D(\theta)$, while the relationship between moisture content, θ, and suction, ψ, provides information on the retentive capacity of the soil under certain specified conditions.

Previous investigations of these parameters have been confined primarily to soils of the continental or temperate regions, and the results are hardly expected to apply to tropical soils where strong aggregation of soil peds is the rule rather than the exception. The recent work of Cagauan (1963) and Cagauan and Uehara (1965) to quantify the relation of aggregate stability with soil anisotropy and the previous studies of Thorne (1949), Cornelison (1954), and Ekern (1966) led to the recent work of Sharma (1966). Sharma initiated studies on the influence of soil structure on moisture retention in Hawaiian latosols. Lin and Chen (1963), working with Taiwan latosols, were the first to utilize pressure plate outflow techniques on tropical soils to determine transmission coefficients of soil water. The current study is a continuing effort to evaluate the dynamic as well as the static behavior of soil water in the suction range of zero to one bar.

The need for a rapid method that gives an accurate evaluation of the transport or transmission coefficients for soil water prompted the examination of existing transient outflow methods. Although these methods have come under considerable criticism for repeated failure of experimental measurements to correlate with theory, Green, *et al.* (1964) demonstrated that field infiltration rates could be predicted from transmission coefficients measured in the laboratory. It appears likely that methods which are not exact with respect to theoretical considerations may nevertheless be exceedingly useful in characterizing the hydrological properties of soils. The transient outflow methods appear to provide a rapid and reasonably accurate means of characterizing water transmission. Transient outflow methods employed in this study are referred to as the one-step and the multi-step methods.

The objectives of this study are three-fold and are enumerated as follows:

1. To study the reliability of the one-step method for measurement of soil water diffusivity on latosolic soils.
2. To compare the results of the one-step method with another transient method, a steady-state method, and calculation methods based on water retention.
3. To evaluate the soil-water transmission coefficients of selected soils classed as Low Humic Latosols.

REVIEW OF LITERATURE

Many reviews on the historical development of water movement in soils in the early period of this century have been presented by others (Bazargani, 1964; Sharma, 1966), and will not be discussed further. Reviews of work leading up to the present study will be of primary concern.

Studies on transport phenomena in porous media have been dealt with in considerable detail by Muskat (1946), and Scheidegger (1960). When the porous medium is the soil, the most important transport phenomenum is the flow of water. The transport or transmission coefficients that describe the movement of water in soils are better known as capillary conductivity (K), and soil water diffusivity (D). Both terms, conductivity and diffusivity, are analogous to those used in physics and chemistry or in any of the physical sciences.

A. Theoretical Development of Flow Equations

The movement of water may occur under both saturated and unsaturated conditions but in almost all agricultural soils the latter condition is predominant and hence, the more important. The force that translocates water in the soil system can be described by the potential theory and by the diffusion theory. The difference between the two is that movement of water is effected by a potential gradient in the former and by a concentration gradient in the latter.

The primary use of the potential theory is in steady-state problems where the moisture content, suction, suction gradient, and flux do not change. A steady-state system is not necessarily a saturated system;

therefore, both transmission coefficients, capillary conductivity, $K(\theta)$, and water diffusivity, $D(\theta)$, must be considered to be a function of moisture content since some of the pores are only partially filled with water. Klute (1952) stated that the presence of air in the pore space has the effect of reducing the effective cross-sectional area available for flow, thus showing evidence of the dependence of K on moisture content, θ. The movement of water in saturated soils can be best expressed mathematically by Darcy's law as:

$$v = - K \frac{\partial\psi}{\partial x} \qquad (1)$$

where v is the flow velocity per unit cross-sectional area, K is the conductivity, x the unit distance, and ψ the suction or pressure. The potential gradient is represented as a partial derivative of suction with respect to distance in equation (1). For unsaturated soils, K is replaced by $K(\theta)$ implying that K is a function of moisture content. In an unsaturated medium, Darcy's law is not invalidated since only the volume available for flow is reduced, not its continuity.

In most instances, however, a transient state, where moisture content, suction, and flux change with time, rather than a steady-state, exists in soils. Hence, in addition to Darcy's law of motion, the flow of water into and out of a unit mass of soil must obey the law of conservation of matter and is consequently represented by an equation of continuity (Klute, 1952),

$$\frac{\partial\theta}{\partial t} = - \nabla \cdot v \qquad (2)$$

where ∇ is the Laplacian operator $\left(\frac{\partial}{\partial x} + \frac{\partial}{\partial y} + \frac{\partial}{\partial z}\right)$, v, the mass of fluid flowing through unit cross-sectional area per unit time, and, θ, the moisture content on a volume basis.

Substituting equation (1) into equation (2) and considering one-dimensional flow, we have the following,

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial x}\left[K(\theta)\ \frac{\partial\psi}{\partial x}\right] \tag{3}$$

Equation (3) can only be evaluated with great difficulty because the partial derivative of ψ with respect to x is difficult to determine experimentally. By assuming a continuous relationship between ψ and x, it is possible to render this derivative in a more useful form by applying the chain rule. Thus,

$$\frac{\partial\psi}{\partial x} = \frac{\partial\psi}{\partial\theta}\ \frac{\partial\theta}{\partial x} \tag{4}$$

where it is now possible to evaluate experimentally the terms on the right-hand side of equation (4). Substituting equation (4) into equation (3) we have,

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial x}\left[K(\theta)\ \frac{\partial\psi}{\partial\theta}\ \frac{\partial\theta}{\partial x}\right]. \tag{5}$$

But the diffusivity, D, is expressed as,

$$D(\theta) = K(\theta)\ \frac{\partial\psi}{\partial\theta} \tag{6}$$

So the diffusion equation is thus written as,

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial x}\left[D(\theta)\ \frac{\partial\theta}{\partial x}\right] \tag{7}$$

which is a non-linear partial differential form of one-dimensional transport by a horizontal semi-infinite column of soil. Klute (1952), and many later workers (Gardner and Mayhugh, 1958; Rawlins and Gardner, 1963), transformed equation (7) by use of the Boltzman transformation. The validity of the differential equation was then tested by having a straight line relationship between $t^{\frac{1}{2}}$ and x (distance). The salient point of their solution of equation (7) resulted in a numerical solution that gave a good qualitative method for predicting the movement of a wetting front in a soil at a previously constant moisture content. It is now evident that in the transient state, the diffusion equation or the diffusivity function is dependent on the concentration gradient expressed as the partial derivative of θ with respect to x. In the present study, the transient state is of primary concern. Equation (7) represents a horizontal case of flow where the effect of gravity is negligible. Evaluation of equation (7) will be shown later.

In the foregoing development of defining the transmission coefficients, $K(\theta)$ and $D(\theta)$, it was assumed that both are single-valued functions of both θ and ψ if one considers either a case of desorption or sorption. Miller and Miller (1955) presented a physical theory on the nature of hysteresis and listed some of the basic assumptions needed to interpret hysteresis effects. They found that the transmission coefficients could best be described if the past history regarding wetting and drying were known. These transmitting functions were thus labeled "hysteresis functions" (Miller and Miller, 1955). Klute (1952) neglected the effects of hysteresis in analyzing flow in unsaturated soils, but in a later work he (Klute, 1965) showed the effects of hysteresis in using radial outflow cells. Staple (1962) found that at a suction of 25-400 cm of water, diffusivity was at least twice as great in a drying soil than in a wetting one. He also generalized that at all tensions, more water is held by the soil during desorption rather than sorption. In his studies of Hawaiian latosols, Sharma (1966) showed that the phenomena of hysteresis had a pronounced effect on $K(\theta)$ as demonstrated by the wetting and drying cycle for selected aggregate sizes analyzed. Philip (1955) concluded that the effect of hysteresis can be minimized if the correct hysteresis loops of the moisture content-suction plot are used.

Difficulties arising in the use of Darcy's law led to the recent work of Swartzendruber (1962a) in which he proposed a modification of Darcy's law to include a departure from the norm called non-Newtonian flow which is dependent on the shear-rate of the system. This non-Newtonian flow implies, then, that the modified equation will include the effects of clay-water interaction which leads to non-Darcy behavior. His modified equation can be written as,

$$v = M(i - I) \tag{8}$$

which is analogous to equation (1) where M is the hydraulic conductance, i the hydraulic gradient, and I the non-Newtonian index. Methods to evaluate M and I are presented in his work and will not be presented here. In his later works, Swartzendruber (1962b, 1963) applied his modified equation to both saturated and unsaturated medias and concluded that in evaluating diffusion-type processes, non-Darcy behavior showed that diffusivity and conductivity are not only a function of θ but also of the moisture gradient.

Other factors that affect the determination of capillary conductivity and soil water diffusivity can be enumerated as follows: (1) temperature, (2) salt content, and (3) pore size distribution. Gardner (1959), Klute (1962), and Jackson (1963b) have demonstrated that diffusivity increases with increasing temperature, and Gardner, *et al.* (1959) showed that diffusivity decreased significantly with an increase in electrolyte concentration. Miller and Miller (1955), Marshall (1958), Jackson (1963a), Amemiya (1965), Sharma (1966), and others have demonstrated the importance of porosity, aggregate size, and pore size distribution in evaluating the water transmitting properties of soils.

B. Outflow Methods

Methods to evaluate transmission coefficients of soil water are described as either steady-state or transient state. A separate discussion of each is presented in the following pages.

1. Steady-State.

In order to evaluate the transmission coefficients, $K(\theta)$ and $D(\theta)$, experimentally, the relationship between suction and moisture content must be known. Recall that the potential theory predicts movement of water in terms of $K(\theta)$ and the diffusion theory in terms of $D(\theta)$. The relationship between the potential theory and the diffusion theory can be expressed as in equation (6).

Initial attempts to measure conductivity and diffusivity were conducted by Richards (1931, 1948) by means of steady-state flow through porous media using porous ceramic plates.

Richards considered horizontal flow and determined the potential gradient within the system by tensiometers. Nielsen and Biggar (1961) used a similar steady-state concept but a simpler apparatus that utilized fritted glass-bead plates to reduce the effect of plate impedance. Later, Elrick (1963), using sintered plates of known impedance, simplified matters by eliminating the use of tensiometers. Head differences of the inflow and outflow burets were used as a measure of potential after applying a given tension. An even simpler method was proposed by Davidson, et al. (1963) to evaluate $K(\theta)$. Soil water content distribution in vertical profiles was plotted to determine $K(\theta)$ after a constant volumetric infiltration rate was established. Capillary conductivity values, assumed equal to the constant infiltration rate to depths of 75 cm, compared favorably with $K(\theta)$ values obtained by Nielsen and Biggar (1961).

2. Transient State

a. Multi-Step Methods. The following methods to be discussed shall be grouped as "multi-step" methods, implying the use of successive increments of applied pressure on the soil sample to determine $K(\theta)$ and $D(\theta)$.

The inherent limitation of steady-state methods (the suction range of the tensiometers) is eliminated by using pressure cells. In 1956, Gardner proposed a transient outflow method that utilizes pressure plate outflow data to calculate capillary conductivity assuming negligible plate impedance. By making simplifying assumptions, Gardner was able to solve the non-linear differential equation expressed in equation (7). These assumptions are as follows: (1) the change or differential of suction is sufficiently small such that $K(\theta)$ remains constant within the increment and (2) the suction, ψ, is a linear function of θ such that $\theta(\psi) = a + b\psi$, where a and b are constants. By applying assumption (1), $K(\theta)$ can be removed from the differential operator and by substituting the differential of assumption (2), equation (7) reduces to,

$$\frac{\partial\psi}{\partial t} = D\,\frac{\partial^2\psi}{\partial x^2} \qquad \text{or} \qquad \frac{\partial\theta}{\partial t} = D\,\frac{\partial^2\theta}{\partial x^2}$$

since θ and ψ are assumed to be complementary linear functions.

To obtain a unique solution for equation (9), boundary conditions must be imposed. The initial boundary condition requires that at $x = 0$ and at any time, the pressure or suction at the boundary between the porous plate and sample is atmospheric at all times or,

$$\psi(0,\ t) = 0 \tag{9a}$$

Secondly, no flow of moisture occurs at the upper boundary of the sample implying the absence of any potential gradient. To make this condition mathematically convenient to handle, it is assumed that at $x = 2L$ and at any time, t, the pressure is also atmospheric or,

$$\psi(2L,\ t) = 0 \tag{9b}$$

Condition (9b), hence, satisfies the assumption that,

$$\left(\frac{\partial\psi}{\partial x}\right)_{x=L} = 0$$

or a pressure gradient is absent.

The third and final boundary condition requires that at any distance, x, within the sample, $(0 < x < L)$ at $t = 0$, there is a difference in pressure or suction,

$$\psi(x, 0) = \Delta \psi \tag{9c}$$

Equation (9) can now be solved explicitly by utilizing the method of separation of variables where ψ is a function of both distances, x, and time, t, or $\psi(x, t) = X(x)\,T(t)$. By Fourier sine series expansion of odd integers, the solution of equation (9) is simplified and can be expressed as,

$$\ln(Q_o - Q) = \ln(8\,Q_o/\pi^2) - \alpha^2 D\, t \tag{10}$$

where Q_o is the total outflow for a pressure increment, Q the outflow at any time t, $\alpha = \frac{n\pi}{2L}$, a constant, and D the diffusivity. Only the values of Q_o, Q, and t are experimentally determined. By plotting $\ln(Q_o - Q)$ against t on a semi-log scale, Gardner (1956) obtained a linear relationship where the slope was equal to $\alpha^2 D$ and the intercept equal to $\ln 8Q_o/\pi^2$.

Using small increments of pressure to comply with assumption (2), Gardner was able to plot capillary conductivity versus suction and moisture content. He attributed deviations from the straight line to be the result of taking too large a pressure increment, thus rendering assumption (2) invalid.

Miller and Elrick (1958) improved Gardner's method by taking into account plate impedance in determining capillary conductivity. Using reduced variables of time and discharge, they found that all possible transient curves can be described by a single parameter, $\{\alpha_1/\tan \alpha_1\}$. This parameter also serves to distinguish between different transient curves and ranges from zero for negligible membrane impedance to unity for negligible slab impedance, where α_n is the n^{th} solution of the equation $a\alpha \equiv \cot\alpha$ and "a" (reduced membrane impedance) = $\frac{Z_m \text{ (impedance of membrane)}}{Z_s \text{ (impedance of soil slab)}}$. The principal limitation of their method is the uncertainty in the determination of flow impedance because if good contact between plate and soil is not established, an unknown contact impedance may exist.

It wasn't until 1963 that Lin and Chen compared the methods of Gardner (1956), and Miller and Elrick (1958). In all determinations, using natural cores and sieved samples, the capillary conductivity that compensates for plate impedance yielded higher values. They also found that in using Gardner's method, deviation from the straight line occurred near $t = 0$ and at the end of the outflow process, indicating the possible effects of plate impedance and error in assuming $K(\theta)$ to be constant.

Rijtema (1959) proposed a method that has an advantage over that of Miller and Elrick in that both membrane and soil contact impedance can be calculated from experimental data. He was able to show by plotting $(Q_o - Q)/Q_o$ against time on a log-log scale, the intercept gave a good indirect estimate of the parameter, $\alpha \tan \alpha$ to determine plate impedance. The simplified method of Kunze and Kirkham (1962) eliminated the necessity of experimentally determining plate impedance and also reduced the number of steps required to calculate $K(\theta)$ and $D(\theta)$. Their method also incorporated the effect of contact impedance without determining it separately. Starting with the equation of Miller and Elrick (1958),

$$\frac{Q_o - Q}{Q_o} = \sum_{n=1}^{\infty} \frac{2 \exp(-d_n^2 Dt/L^2)}{\alpha_n^2 (a + \csc^2 d_n)} \tag{11}$$

they modified it slightly and made it computationally easy by the following,

$$\frac{Q}{Q_o} = 1 - \sum_{n=1}^{\infty} \frac{2 \exp(-\alpha_n^2 Dt/L^2)}{\alpha_n^2 (a + \csc^2 \alpha_n)} \tag{12}$$

With this modified equation, it is possible to accurately plot initial outflow at small values of time on a log-log scale. Kunze and Kirkham (1962) considered the matching of the initial outflow curve with the theoretical more meaningful than matching the final phases of outflow. By plotting Q/Q_o against actual time, t, and at the same time against reduced time $\alpha_1^2 Dt/L^2$, the values of the parameters, α, and t_{RP}, can be determined by matching experimental outflow curves with theoretical curves calculated from a ratio of plate impedance to impedance of soil slab, a. Actual time is related to reduced time by a factor of 10^{-3}, i. e. 1, 10, 100, 1000 actual time correspond to 0.001, 0.01, 0.1, 1,0 reduced time. Diffusivity is then calculated by using the following equation,

$$D = L^2/\alpha_1^2 \, t_{RP} \tag{13}$$

where t_{RP} is the reference point obtained by observing the actual time value that coincides with a value of 1.0 for the reduced time when matching the experimental and theoretical outflow curves. Capillary conductivity, K(θ), then is calculated by using the relationship,

$$K(\theta) = D(\theta)\,\frac{\Delta\theta}{\Delta\psi} = D(\theta)\,\frac{\Delta\theta}{\psi_2 - \psi_1} \qquad (14)$$

where $\Delta\theta = \frac{Q_0}{V}$, V = bulk volume of soil slab. The advantages of Kunze and Kirkham's method are two-fold: (1) separate determination and calculation of plate and contact impedance are eliminated and (2) the coordinates of time and outflow become sensitive, after modifying equation (11), to initial outflow data, thus, emphasizing the importance of early outflow rather than later results in determining the parameters, α and t_{RP}. Even with these advantages, however, they found that the entire experimental curve did not fit the theoretical. Whenever the experimental curve transversed the theoretical, theory no longer applied to that portion crossing over. Further evidence presented by Elrick (1963), Bruce and Klute (1963), Jackson _et al_. (1963), and Bazargani (1964), clearly demonstrated that theoretical discrepancies resulted in a poor match of experimental and theoretical curves. Kunze and Kirkham (1962) attributed the deviation from theoretical to be a direct result of wrongly assuming K(θ) and D(θ) to be constants.

Although these discrepancies arise, the transient-state methods have more appeal than steady-state because $D(\theta)$, $K(\theta)$, and $\psi(\theta)$ can be determined simultaneously on a single soil sample over a given range of suction (Swartzendruber, 1965).

Recently, Kunze and Peters (1966) were able to fit almost all of their experimental points on a theoretical curve of Miller and Elrick (1958) by plotting Q/Q_o against time on a log-log scale by the method proposed by Kunze and Kirkham (1962). By previous methods, values of Q/Q_o usually fell below the theoretical during the later stages of outflow when the initial stages were matched with the theoretical. To correct this, Kunze and Peters (1966) assumed that the ratio of experimental time, t_{exp}, and theoretical time, t_{theo}, at specific Q/Q_o values could be represented as,

$$\frac{t_{theo}}{t_{exp}} = \frac{D_{exp}}{D_{theo}} = \frac{D}{Di} \qquad (15)$$

where D_i is the initial diffusivity assumed constant for a particular pressure increment. By plotting log D/D_i against log Q/Q_o, they were able to account for changes in $D(\theta)$ during the late stages of outflow, whereas the initial work (Kunze and Kirkham, 1962) accounted for $D(\theta)$ in the early stages. They hypothesized that changes in $D(\theta)$ are affected by three factors: (1) inherent structural instability of non-consolidated pores; (2) inability of pores to drain instantaneously with pressure change; and (3) failure to satisfy boundary conditions required by theory.

b. One-Step Method. As the name of the method implies, only a single increment of pressure is required to determine $X(\theta)$ and $D(\theta)$ instead of successive increments.

In 1962 Gardner simplified the outflow method considerably by deriving a relationship in which only a single equilibration step is required. He considered the one-dimensional outflow equation (equation [7]) but assumed $D(\theta)$ to be a variable function of x and t. Again, the method of separation of variables was employed where $\theta(x, t) = X(x)T(t)$ to obtain

$$\frac{1}{T}\frac{dT}{dt} = \frac{D}{X}\frac{d^2X}{dx^2} = -\alpha^2 = \text{const} \qquad (16)$$

It is significant that at this point, Gardner makes an important assumption. He assumes $D(\theta)$ to be separable into functions of x and t since $D = f(\theta)$ and $\theta = f(x, t)$. Hence, $D(\theta)$ can be written as $D(x, t) = D_x(x)D_t(t)$. Using this relationship and $\theta(x, t) = X(x)T(t)$, equation (16) reduces to two ordinary differential equations,

$$\frac{1}{D_tT}\frac{dT}{dt} = \frac{1}{X}\frac{d}{dx}\left[D_x\frac{dX}{dx}\right] = -\alpha^2 = \text{const} \qquad (17)$$

In considering equation (17), Gardner (1959) was able to show by numerical methods that water content remained uniform throughout the sample except at the lower boundary in the region of the porous plate. Hence, it was assumed that D_x is constant at any distance, x, within the sample and can be removed from within the differential operator. D_t, however, changes with time.

By imposing boundary conditions (9a) and (9b) of the original diffusion equation, Gardner was able to show that the solution of the X-dependent differential equation yielded a constant,

$$\alpha^2 = \frac{n^2\pi^2\, Dx}{4L^2} = \text{const} \qquad (18)$$

where $n = 1$ since the series solution of the Fourier sine expansion converges rapidly at this point. By equating the left-hand side of equation (17) with α^2, we have

$$D = -\frac{4L^2}{\pi^2(W - W_f)}\frac{dW}{dt} \qquad (19)$$

where W is the volumetric moisture content at any time, t, when the instantaneous outflow rate dW/dt is determined, W_f, the final equilibrium volumetric moisture content, and L, the height of the core sample. For values of $\alpha^2 Dt > 0.3$, the approximations of D should be reasonably accurate since the second term of the series expansion would be less than one per cent of the first (Gardner, 1956). Thus, $D(\theta)$ can be evaluated directly from instantaneous outflow rates, moisture content, and the dimensions of the sample.

Doering (1965) was the first to test this method, subsequently called the one-step method. In general, the results compared very well and at times, the one-step method gave better approximations than the other three methods (negligible and non-negligible plate impedance methods and a constant flux method).

Doering (1965) stated, although both the one-step and the multi-step methods have the same theoretical foundation, the significant experimental difference between the two methods can be attributed to the dependence of flow properties to small gradients. In the case of the multi-step, this results in a scatter of points, while in the case of the one-step method, relatively smooth curves were obtained, suggesting that calculated experimental values resulting from the large gradient caused by applying a single, large pressure increment gave a more reliable estimate of diffusivity and conductivity.

The major and most apparent advantage of the one-step method is the reduction of time required to obtain data necessary to calculate $D(\theta)$ and $K(\theta)$. Also, the experimental procedure does not require (1) soil moisture content to be a linear function of suction within the applied pressure increment and (2) diffusivity to be constant throughout the water content changes.

Kunze and Peters (1966) also demonstrated that results obtained by the one-step method gave good correlative results when compared with their modified version of the multi-step method based on the previous work of Miller and Elrick (1958) and Kunze and Kirkham (1962). The size of the pressure increment used was small, from 8 to 13 millibars.

c. Infiltration Methods. Methods similar to the constant flux procedure developed by Childs and Collis-George (1950) have been employed to study infiltration.

Davidson, et al. (1963), using a hand-packed column of soil, were able to determine the moisture distribution within the column by clamping off the moisture supply at a preselected time and by observing the movement of the wetting front. Diffusivity values were calculated using the equation developed by Bruce and Klute (1956), where

$$D(\theta) = \frac{1}{2t_o} \frac{dx}{d\theta} \int_{\theta_n}^{\theta} x d\theta \qquad (20)$$

Although imposing some stringent assumptions to their theoretical procedure, Davidson, et al. (1963) were able to obtain a good comparison of moisture distribution with depth between a silt loam and a sandy loam. However, in most instances, their results demonstrated a departure from the theory set forth by the diffusion equation.

Hanks and Bowers (1962) used the method of Bruce and Klute (1956) to determine diffusivity values for use in numerical calculations of infiltration. Their calculations showed that the diffusivity of water in the near-saturated zone near the wetting front controlled the infiltration rate. Unfortunately, the "wetting front" method of Bruce and Klute is least accurate in the wet range.

C. Calculation Methods

In evaluating the transmission coefficients by means of calculation methods, the usual theoretical development of this subject starts with Kozeny's equation,

$$K = \frac{\varepsilon^3}{S^2K} \qquad (21)$$

where K is the permeability, ε, the water-filled porosity, S, the surface area, and k, an empirical constant, and Poisseuille's equation,

$$\mu = -\ (r_t/8\eta)\ d\psi/dl \qquad (22)$$

where μ is the mean velocity, η the viscosity of the fluid, r_t the radius of the cross-section of tubes, and $d\psi/dl$ the gradient that causes flow.

One of the first attempts to determine conductivity by calculated means was presented by Childs and Collis-George (1950) to be abbreviated here as CCG. They developed an equation that accounts for sequences of different sized pores when arranged randomly in a porous system. Utilizing Kozeny's and Poisseulle's equations and two compensating assumptions, they were able to derive the following,

$$K = m \sum_{\rho=0}^{\rho=R} \sum_{\sigma=0}^{\sigma=R} \sigma^2 f(\rho)dr\, f(\sigma)dr \qquad (23)$$

where K is the total permeability, m a constant called the matching factor, σ^2 the square of the radius of the smaller (replaced by ρ^2 if $\rho < \sigma$), $f(\sigma)$, and $f(\rho)$ the distribution functions that describe the pore space of uncemented sand and $f(\sigma)dr$ and $f(\rho)dr$, the fraction of the total apparent volume which is occupied by pores whose radii range from ρ to $\rho + dr$ or σ to $\sigma + dr$, respectively. The summation is stopped at the pore size, R, the largest pore remaining full of water at a specific moisture content. The two compensating assumptions used were: (1) all effective resistance to flow in the sequence is confined to the smaller of the pores, resulting in an over-estimation of total permeability and (2) negligence of by-passing sequences of several pores results in an under-estimation when only a direct sequence of pore size is considered to contribute to permeability.

Experimentally, the required data are the moisture retention curve and a K value. The critical determination of this method lies in the selection of m, the matching factor. This factor is calculated by matching a measured value of K at some moisture content or suction.

Similarly, capillary conductivity, K(θ), can also be calculated from retention data by matching K(θ) values instead of permeability, K. The major limitation of the CCG method is that it considers only a cross-sectional factor and neglects the effect of pore length.

Marshall (1958) derived an equation to calculate both K and K(θ) by assuming: (1) that the flow rate is controlled by the cross-sectional area of the channels connecting the pores and (2) that in an isotropic medium, where a random distribution of pore space or voids exist, the fractional pore area is $\varepsilon(\mathrm{cm}^2/\mathrm{cm}^2)$ in any plane.

By considering these two assumptions and expanding Poisseulle's equation for the average cross-sectional area of channels, Marshall obtained the following equation after equating it to Darcy's law,

$$K = \varepsilon^2 n^{-2}[r_1^2 + 3r_2^2 + 5r_3^2 + \ldots + (2n - 1)r_n^2]/8 \qquad (24)$$

where ε is the fractional pore area filled with water, n the number of pore classes up to moisture content of interest (the value of n is different in calculating K for each moisture content chosen), and r_1, r_2, ... r_n, the radii of pores occupied by water--r_1 being the largest and r_n the smallest. To evaluate pore radius experimentally, the following relationship between pore radius, r, and suction, h(cm of water), was utilized,

$$r = 2\gamma/\rho g h \qquad (25)$$

where γ is the surface tension of water, ρ the density of water, and g the acceleration due to gravity.

Equation (24) is simplified by using the inverse relationship of r and h, where h is determined directly from moisture retention data, to obtain,

$$K = 2.8 \text{ X } 10^{-3} \varepsilon^2 n^{-2} [h_1^{-2} + 3h_2^{-2} + 5h_3^{-2} + \ldots + (2n - 1)h_n^{-2}] \qquad (26)$$

which is applicable at 20 C. The constant value of $2.8 \text{ X } 10^{-3}$ is obtained by evaluating the constants γ, ρ, and g. It is apparent h_1, h_2, ... h_n represent suction in equal classes where h_1 corresponds to r_1, hence, belonging to the class with the largest pore, and h_n corresponds to r_n, the class with smallest pores. To calculate capillary conductivity, the same equation is used except a constant value of $2.7 \text{ X } 10^2$ at 20 C replaces $2.8 \text{ X } 10^{-3}$ simply by multiplying the latter by g/η, where η is the viscosity of water. The advantage of Marshall's method is that it eliminates the tedious determination of a matching factor. Marshall obtained good correlations when he compared the measured and calculated permeability of slate, sandstone, and quartz sand. Any deviation from the expected was attributed to (1) the material or porous media not being isotropic, (2) the material containing lengthy conducting channels, and (3) swelling of materials containing moderate to high clay content.

Millington and Quirk (1961), abbreviated MQ, developed a similar equation to Marshall's except that ε^2 is replaced by $\varepsilon^{4/3}$. This equation is thus written as,

$$K = 2.8 \times 10^{-3} \varepsilon^{4/3} n^{-2}[h_1^{-2} + 3h_2^{-2} + 5h_3^{-2} + \ldots + (2n - 1)h_n^{-2}] \quad (27)$$

The assumptions of the MQ method can be enumerated as follows: (1) isotropic distribution of total pore space, (2) random distribution of pore size classes, (3) reliable pore size distribution data, and (4) stable and inert materials. A comparison of the CCG and M methods by Millington and Quirk with their method revealed that both the CCG and M methods yielded higher approximated values than the experimentally determined values of permeability.

In 1964, Brooks and Corey, to be abbreviated BC, presented a theoretical development of functional relations among saturation, suction or pressure, and permeabilities of air and water in which the evaluation of three parameters, S_r, residual saturation, P_b, bubbling pressure and λ, pore size distribution index, is required. The bubbling pressure, P_b, is related to the maximum pore size forming a continuous network of flow channels within the porous media, and λ evaluates the size distribution of the flow channel. All three of the parameters can be determined from the moisture retention curve. The first approximation of S_r is determined at a point where the curve of moisture content (θ) versus suction (ψ) becomes asymptotic to the vertical θ axis. S_r values are obtained as fractions of complete saturation and are used to evaluate S_e, effective saturation, by the relationship,

$$S_e = \frac{S - S_r}{1 - S_r} \qquad S_r < S \leq 1 \qquad (28)$$

where S is the saturation value at any specified suction.

The calculated S_e is then plotted against suction on a log-log graph. Successive approximations of S_r are tried until a value that yields a linear relationship of the log-log plot at high suction or pressure is obtained. The intercept, P_b, and slope, λ, give the necessary values of the parameters to evaluate the relative conductivity, K_r. K_r is calculated by means of the relationship,

$$K_r = \left(\frac{P_b}{P_c}\right)^{3\lambda + 2} \qquad (29)$$

where P_c is any suction (notation of BC method). The straight line relationship is valid as long as $P_c > P_b$.

The permeability K is then determined by,

$$K = K_r K_{p_b} \qquad (30)$$

where K_{p_b} is the permeability at saturation. Comparison of calculated and measured values of K_r and K gave good correlation for sand according to Brooks and Corey (1964).

Nielsen, et al. (1960) compared the CCG and M methods to calculate conductivity of two loess soils and two glacial soils. In all determinations, the M method gave higher results than measured values, which was attributed to failure of soil material to meet the assumptions specified by the methods. They speculated that if experimentally determined constants were used in conjunction with the M equation, a better fit would result. The CCG method compared favorably with measured values of $K(\theta)$ for the loess soil but poorly for soils derived from glacial material.

Jackson, et al. (1963) compared measured $K(\theta)$ of 50 - 500 μ sand with calculated values obtained by the CCG, M, and MQ methods. All of the calculated values were higher than measured. They found, however, that the curve obtained by the MQ method had a similar shape to that of measured conductivity and thus prompted the employment of a constant or matching factor to obtain good correlation of results. The matching factor was obtained by taking the ratio of measured to calculated saturated conductivities.

Sharma (1966) found that by employing the CCG, M, MQ, and BC methods, calculated curves of conductivity were significantly higher than measured at all suction values. However, by incorporating the MQ and BC methods, he obtained a modified equation to calculate capillary conductivity.

$$K = 2.7 \times 10^{2}(\varepsilon - S_T)^{2}\, n^{-2}[h_1^{-2} + 3h_2^{-2} + 5h_3^{-2} + \ldots + (2n - 1)h_n^{-2}] \quad (31)$$

Using this equation, he obtained good correlative results in comparing measured and calculated conductivities of selected aggregate sizes of Hawaiian latosols for a suction range of 0 - 200 cm of water.

MATERIALS

A. Soils

The soil samples used in this study were from the Molokai and the Wahiawa series of the Low Humic Latosols. Previous studies by Sherman and Alexander (1959) have shown that the soils of this group possess similar mineralogical and chemical properties. However, Uehara, et al. (1962) found that there were significant micromorphological differences among Hawaiian Latosols. Later Cagauan (1963), and Cagauan and Uehara (1965) showed that the Molokai and the Wahiawa series exhibited extremes in a structural sequence of the Low Humic Latosols, with the Molokai at the lower end and the Wahiawa at the higher.

The Low Humic Latosols, as described by Sherman and Alexander (1959), are characterized by the development of a weak A_1 horizon overlying a weathered B horizon containing as much as 55 to 80 per cent kaolin clay. Another important attribute of these soils is the relatively high free iron content. They concluded that the uniformity in chemical and mineralogical composition throughout the solum is the result primarily of uniform weathering of the parent material.

The mineralogical composition of the Low Humic Latosols is predominantly kaolin group minerals, kaolinite and halloysite, and hydrated oxides of iron and aluminum with small amounts of soil mica, montmorillonite, and other secondary minerals. The low cation exchange capacity (10 to 25 milli-equivalents per 100 gm of soil with base saturation of 30 to 100 per cent) tends to reflect the mineralogy of the soil. Detailed information on the chemical properties of these two soils have been given by previous workers (Kanehiro and Chang, 1956; Matsusaka, 1952), and are available; therefore, chemical analyses were omitted in this study.

The high aggregate stability is an outstanding physical property of these soils. After 70 years of cultivation, these soils showed practically the same percentage of stable aggregates as a virgin soil. Uehara, et al. (1962) attributed this salient physical property to be the result of two factors: (1) the nature of the soil minerals (kaolin minerals and free iron oxides); and (2) the arrangement of these mineral constituents within soil pedological units and the process by which these arrangements came about. Soil scientists have generally believed that stable aggregates were the result of the free iron oxides acting as cementing agents to consolidate kaolin minerals to form stable aggregates (Lutz, 1936; McHenry and Russell, 1943; Kroth and Page, 1946; McIntyre, 1956).

Since the mineralogy of the Low Humic Latosols consists predominantly of kaolinite and free iron oxides, these constituents are not the limiting factors in the development of water stable aggregates. Thus the degree of arrangement and orientation of these mineral constituents are not the limiting factors in the development of water stable aggregates. Thus the degree of arrangement and orientation of these mineral constituents must be the index used to differentiate soils with stable aggregates and well-developed structure and those without. This index is called soil anisotropy as used by Cagauan (1963) and Cagauan and Uehara (1965). With increasing anisotropy, there is a curvilinear increase in aggregate stability. This relationship is generally true for Hawaiian latosols.

Sharma and Uehara (1967) studied the influence of stability of these soils in their investigation of the relationship between moisture retention and soil structure. The strong aggregation of peds in both the Wahiawa and the Molokai was apparently responsible for the sand-like water retention of these soils (Ekern, 1966). This sand-like behavior is possibly due to the similarity of inter-aggregate pore geometry of these peds with sand grains. Further discussion on this subject will be presented in a later section.

All of the soil samples used were collected on the island of Oahu. Both soils are under extensive agricultural use, particularly for sugar cane and pineapple culture.

The Molokai soil samples were collected at 20-40 inch depths from a road cut on land belonging to the Oahu Sugar Company. The sample site is located approximately one mile above the H-1 defense highway and about 50 yards to the left of Kunia Road. The vegetation is primarily sugar cane with an extensive undergrowth of nut grass. The soils of the Molokai family are usually found at elevations ranging from 0-1500 feet with a mean annual temperature between 70-78 F and an average rainfall of 10-40 inches per year (Cline, 1955).

The Wahiawa soil samples were collected at 37-43 inch depths in a eucalyptus grove in an area between the Leilehua Golf Course and the 298th Artillery Group National Guard Armory. The sample site is located approximately 150 yards above Dole Field 4523. The Wahiawa soils are typically found at elevations between 250-1200 feet with an average temperature range of 70-73 F and an average annual rainfall of 30-60 inches (Cline, 1955).

B. Apparatus

1. Tempe Pressure Cells.

Outflow determinations for the one-step method were obtained by extracting moisture from undisturbed natural cores and hand-packed disturbed cores enclosed in the Tempe pressure cells. Fig. 1 shows a cross-sectional view of the pressure cell that was designed by Reginato and van Bavel (1962) of the U. S. Water Conservation Laboratory at Tempe, Arizona. The pressure cells are commercially available at the Soil-moisture Equipment Company of Santa Barbara, California, catalog No. 1400. The cells are constructed completely of clear lucite plastic and accept either ceramic (about 1400 mb bubbling pressure) or monel (about 200 mb bubbling pressure) porous plates depending on the tension range being studied. The retainer cylinders to hold the soils are made of 19 gauge brass and are designed so that they can be readily used in conjunction with a soil core sampler (catalog No. 200). Air pressure is introduced at the tubular opening at the top of the pressure cell and outflow is measured by the accumulation of water drops in a 25 ml Pyrex graduated cylinder located beneath the tubular opening at the base of the cell. Laboratory line pressure (60 psi) was used as the source of pressurized air. Air was first allowed to pass through a drain filter (Type 12-005, C. A. Norgren Co., Littleton, Colorado) then through a pressure regulator of 125 psi capacity (Type 11-002, C. A. Norgren Co.,

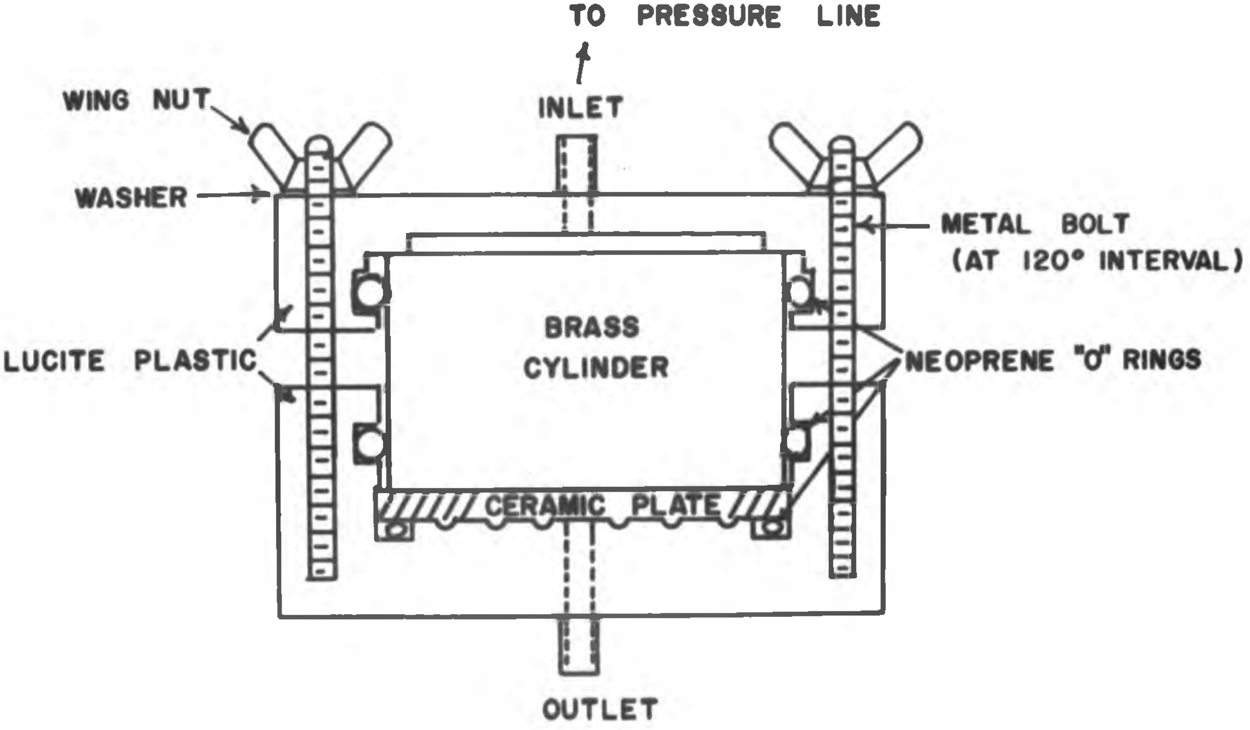

Fig. 1. Cross-sectional Diagram of the Tempe Pressure Cell

Littleton, Colorado) for rough adjustment of air inflow and finally through a nullmatic pressure regulator (Model 40, Moore Prod., Philadelphia, Pennsylvania) for fine adjustment of air pressure that was indicated on a 15 psi pressure gauge. The pressure line was then connected to a series of needle valves which in turn were attached to the individual Tempe pressure cells. Precise measurement of pressure was accomplished with a mercury manometer.

2. Pressure-Plate Extractor.

The pressure plate extractor shown diagrammatically in Fig. 2 was used to measure outflow by the multi-step method. The entire unit was constructed of clear lucite plastic and the design[1] was similar to that used by Tanner and Elrick (1958), Kunze and Kirkham (1962), and Bazargani (1964). The base of the unit was constructed to accept a porous ceramic plate, 1/4-inch thick and 4 1/4-inch diameter, with an air-entry value of 1 bar. One of the outlet tubes at the base was connected to a 100 cc buret and to a 1 cc graduated pipet with 0.01 graduations. Tygon tubing (3/16 inch I. D., 5/16 inch O. D., 1/16 inch wall) was used for connections. Applied air pressure was controlled by a wall-mounted manifold unit similar to that used for the Tempe cells. Instead of connecting the air pressure line to a series of needle valves, the lines were connected in series to a set of nullmatic air regulators, each connected to a manometer for precise measurement of pressure in the whole range from zero to one bar. The pressure

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[1] Dr. R. E. Green, Assistant Professor, Department of Agronomy and Soil Science, University of Hawaii.

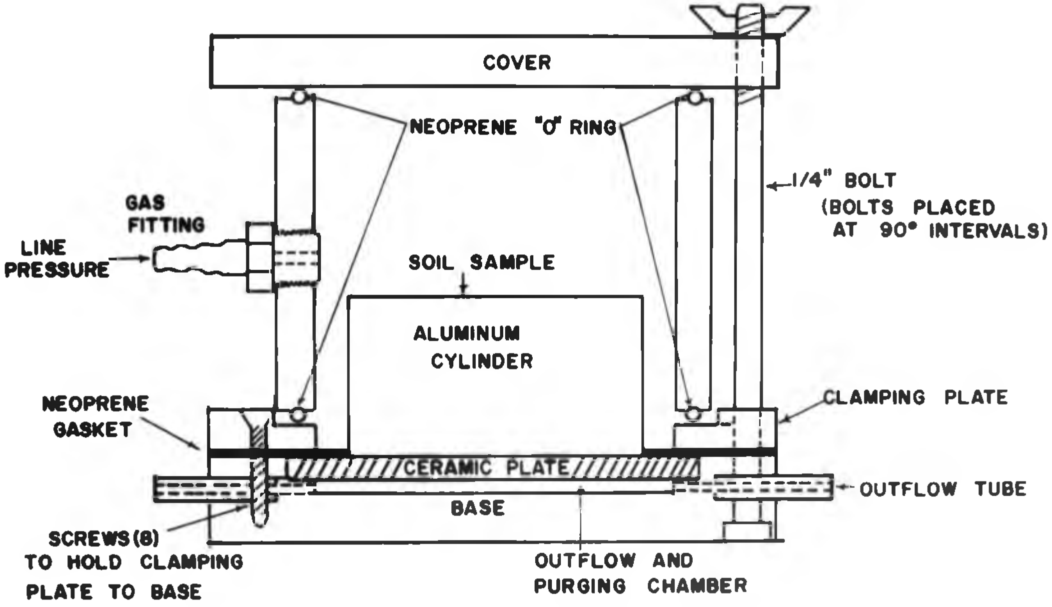

Fig. 2. Cross-sectional Diagram of the Pressure Plate Extractor

line was connected to the gas fitting by the same tygon tubing. The tubing connecting the extractor and the buret was of a length such that it could be easily inserted into a manually operated water circulating pump. The pump was constructed of clear lucite plastic except for the handle which was made of steel. A cross-sectional diagram of the pump is shown in Fig. 3. The primary purpose of the pump was to introduce water into the system beneath the porous plate and to remove all entrapped air bubbles remaining below the porous plate before the flow measurements. There was no accumulation of bubbles beneath the ceramic plate during the measurements.

An aluminum cylinder of dimension 7.4 cm diameter by 4.1 cm height was placed in the center of the plate and packed with soil before saturating both the plate and the soil.

3. Infiltration Column.

A column of hand-packed soil was used in an attempt to determine conductivity, $K(\theta)$, by direct means. The column was constructed of clear lucite plastic cylinders which measured 2.0 cm in height and 4.5 cm in diameter. The unit is shown diagrammatically in Fig. 4. The entire unit measured 80 cm in height. Tensiometers were placed in alternating cylinders to measure differences in suction within the column. The tensiometers were constructed of lucite tubes sealed with Porvic membrane of low impedance. Ethylene dichloride was used as the sealant.

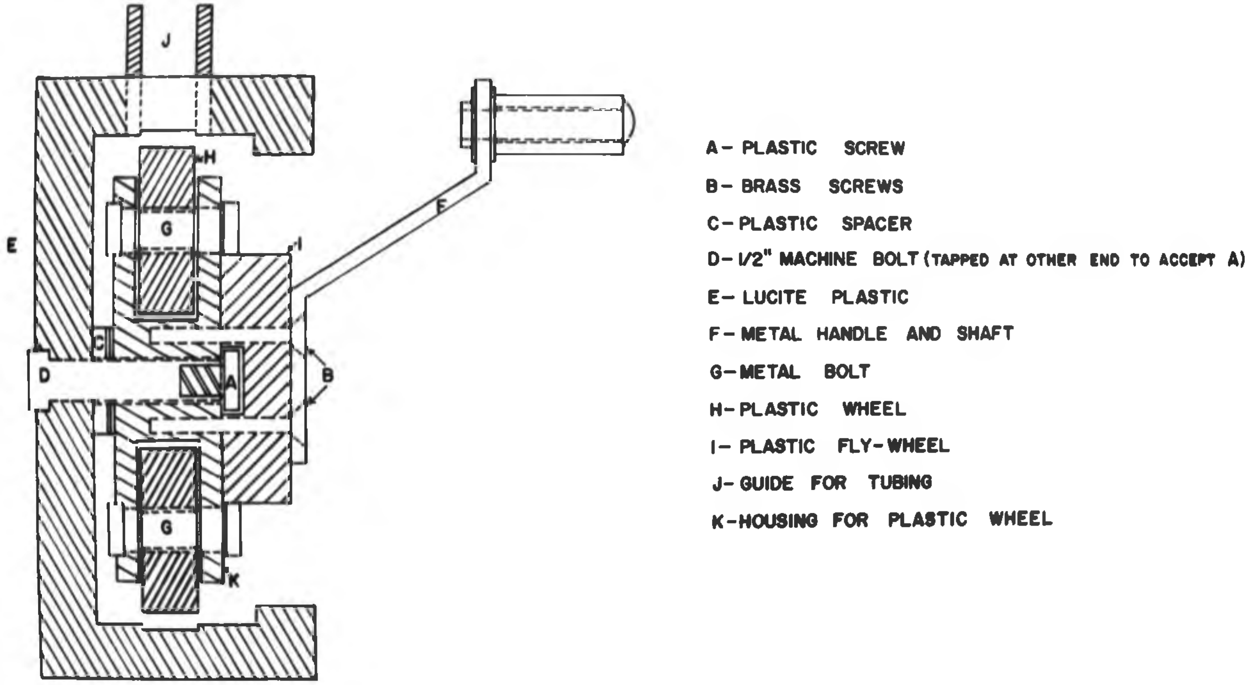

Fig. 3. Cross-sectional Diagram of the Manually-Operated Water Circulating Pump

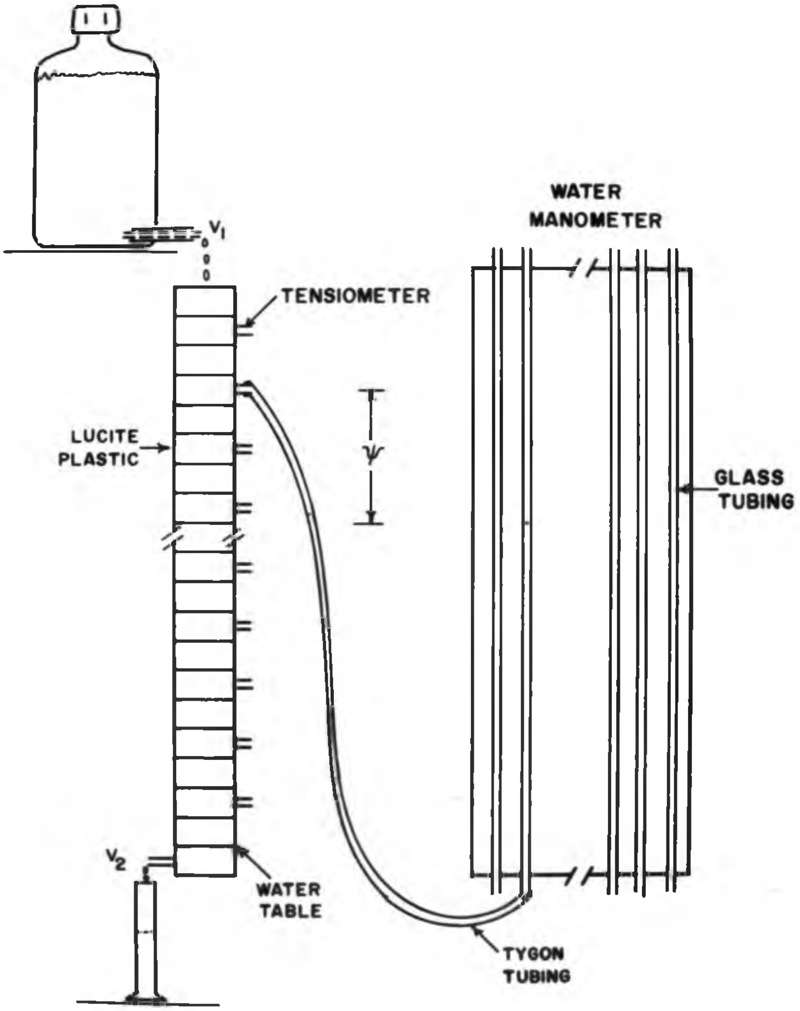

Fig. 4. Schematic Diagram of the Infiltration Column

METHODS

A. Mineralogical Analysis

Mineralogical constituents of both the Wahiawa and the Molokai soils were identified with the Norelco X-ray diffractometer and the Stone automatic differential thermal analysis (DTA) instrument.

X-ray diffraction analyses were run on randomly oriented powdered whole soil (100 mesh) and on the clay fraction oriented to the 001 axis on glass slides. The clay fraction was separated from the sand and silt by a rapid method described by Dr. L. D. Swindale[2] for tropical soils with high free iron content. After adding sodium carbonate to disperse the soil, the clay fraction was separated from the sand and silt fraction by centrifuging the suspension at 750 r.p.m., deferrated in a 0.3 M sodium citrate solution at 80 C with sodium dithionite, and divided into two equivalent parts for saturation with K^+ and Mg^{++} ions before being X-rayed.

Samples run on the Stone DTA were air-dried, ground to pass through a 100 mesh sieve, and equilibrated to a uniform moisture in a 50 per cent relative-humidity chamber before being analyzed.

B. Collection of Aggregates

The different sized water-stable aggregates for flow measurements separated from soil previously passed through a 10 mesh sieve. A 50 gm sample was placed on the top sieve of a nest of sieves with the following sizes in order: 10, 20, 40, 60, and 140 mesh. The sieves

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[2] Routine Test Method, Department of Agronomy and Soil Science, University of Hawaii.

were agitated in water at a frequency of 30 oscillations per minute for 30 minutes on an aggregate stability machine. The samples were then air-dried and stored in labeled jars for later use.

C. Specific Gravity

The specific gravity or particle density of both soils (< 2 mm samples) were determined by using glass pycnometers as described by Blake (1965). A vacuum dessicator was inadequate for complete removal of gases from the stable peds of both soils; hence, the samples were gently heated on a hot plate set at 60 C. Failure to remove entrapped air led to low values. Values obtained by using the vacuum dessicator ranged from 2.30-2.45 while those obtained by gentle heating ranged from 2.85-3.00. An average value of 2.90 was obtained for the Wahiawa while the Molokai had an average value of 2.97.

D. Outflow Techniques

All outflow determinations were determined under nearly isothermal conditions at a temperature of 22.0 ± 0.5 C.

1. One-Step Method

a. Undisturbed Cores. Natural undisturbed cores were obtained for the Molokai soils with a core sampler[3] that utilizes the brass cylinders used in the Tempe pressure cell unit. Samples were immediately stored in moisture cans for later use. Undisturbed cores for the Wahiawa soils were difficult to obtain because core samples usually crumbled when separated from the sampler.

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[3] Catalog No. 200, Soilmoisture Equipment Co., Santa Barbara, California.

b. Disturbed Hand-packed Cores. Disturbed samples were obtained by first air-drying bulk field samples and crushing to pass through a 2 mm sieve. The < 2 mm samples were then bottled and stored for later use. Disturbed samples were hand-packed by placing the brass core on the base of the Tempe cell and filling the core with the < 2 mm sample. In an attempt to obtain close packing rather than loose, the core was gently tapped and then compacted with a size 10 or 11 rubber stopper each time after adding about a tablespoon of sample. This procedure was followed in an attempt to obtain uniform bulk density.

The different sizes of water stable aggregates were packed in a similar manner as the < 2 mm samples.

c. Sorption and Desorption. The ceramic and monel plates to be used in outflow determinations were immersed in de-ionized water and placed in a vacuum dessicator to remove all air bubbles from the plates. The base of the Tempe cell was filled with water to a level that was maintained by connecting the cell to a constant head reservoir. The saturated plate was then inserted into the base of the pressure cell and kept in place by means of an "O" ring. For determination on undisturbed field samples, the core samples were removed from moisture cans previously used for storage. The core samples were placed directly onto the porous plate and saturated by capillarity. For disturbed hand-packed samples of < 2 mm, 20, 40, 60, and 140 mesh, the empty brass cylinders were first placed on the porous plate before packing the sample. Again the sample was saturated by capillarity. Saturation was usually attained within 24 hours. Upon saturation, the level of the reservoir was lowered 1.5 cm below the top of the brass core for a uniform starting point before applying the pressure for desorption. At equilibrium, the top cover of the Tempe cell was screwed tightly into place to obtain an air-tight system. A pressure of one bar was then applied and the accumulating outflow volume measured in a 25 ml graduated cylinder (0.01 graduation) with time. The sample was not desaturated and saturated again as suggested by Doering (1965) in an effort to minimize hysteresis effects. The procedure proposed by Doering was unacceptable because on resaturating the sample by capillarity, there was a 10 to 15 per cent difference between the total water holding capacity at the assumed saturated state and the calculated porosity. Time was recorded accurately to 0.01 minute on a Lab-chron 1401 timer. All runs were made at a temperature of 21 ± 0.5 C.

Equilibrium was usually attained at the end of 48 hours or when there was no measurable outflow. The final volumetric water content was obtained by oven-drying the sample at 105 C.

Moisture release curves for pressure up to one bar were obtained with the Tempe cells by applying successive small increments of pressure and measuring the accumulated outflow after each increment. From the one-step method, the diffusivity-water content relationship (D vs. θ) was calculated directly.

Capillary conductivity was calculated indirectly by using the graphically determined slopes of the moisture release curve (θ vs. ψ) and the relationship expressed by equation (6).

2. Multi-step Method

The multi-step method utilizes a pressure outflow unit similar to that described by Kunze and Kirkham (1962) with a few modifications, as shown in Fig. 3.

Before saturating the sample (disturbed < 2 mm), the ceramic plates were air dried, then secured in the base of the outflow unit. An aluminum cylinder was placed in the center of the plate and the soil sample was packed in the core in the same manner as for the one-step method. The entire system, soil sample and ceramic plate, was then saturated with CO_2 before saturating with H_2O to assure complete dissolution of trapped gases. The porous ceramic plate and soil sample were then saturated by capillarity for 48 hours due to large sample size. Water was pumped under the plate by means of the manually operated water circulating pump. To attain complete saturation a vacuum pump was connected to the gas fitting outlet and allowed to run until free water was visible at the surface of the sample.

Initial outflow was then determined by measuring the rate at which an introduced air bubble moved in the graduated 1 cc pipet (0.01 graduations). Equilibrium was attained when cumulative outflow, measured in a 100 cc buret, remained constant.

The outflow variable, Q/Q_o, was plotted vs. time on Dietzgen 34 ODR-L23 log-log paper. The parameters of the diffusion equation were then determined by matching the experimental outflow curve with theoretical curves developed by Kunze and Kirkham (1962).

3. Infiltration Column

Clear lucite plastic cylinders were sealed together by first applying melted paraffin between the cores and then wrapping the joints with electrical tape to add strength to the column. The column was secured on a wooden rack before packing with soil. The soil was added in increments of 100 gm in an attempt to maintain uniform packing. After packing was complete, a No. 5 Whatman filter paper and "fine" monel screen were cut to the diameter of the core and placed on top of the column. Saturation of the column was effected by controlling the rate of water dripping on the screen and filter paper. When inflow and outflow became constant, implying a steady-state system, the difference in potential at each tensiometer was determined. Conductivity was then calculated directly by employing Darcy's equation.

E. Calculated Conductivity

Methods of calculating capillary conductivity from moisture retention curves were suggested by Childs and Collis-George (1950), Marshall (1958), Millington and Quirk (1960), Brooks and Corey (1963), and Sharma (1966). These methods will be known as CCG, M, MQ, BC, and S, respectively. Previous investigations by Sharma (1966) have shown that his method of calculation gives a good approximation of capillary conductivity for selected aggregate sizes of Hawaiian latosols. The $K(\theta)$ values calculated from the moisture release curves by the S method were compared with the experimentally obtained $K(\theta)$ from the one-step method.

RESULTS AND DISCUSSION

A. Mineralogy

The mineralogical results obtained from X-ray diffraction and differential thermal analysis (DTA) are presented in Figures 5 and 6. X-ray diffractograms of the whole soils demonstrate that the mineralogical compositions of both soils are practically the same. The predominant clay mineral is kaolinite (7 Å) with small amounts of soil mica (10 Å). A comparison of the X-ray diffractograms of the K^+-saturated clay fractions indicates the mica content is greater in the Wahiawa than in the Molokai. This conclusion is supported by the findings of Juang (1964) who made a careful study of the mica content of soils classed as Low Humic Latosols. Juang found that the mica content tended to increase with a sequential increase in elevation and rainfall. The kaolinite content of the Wahiawa is seen to be greater than that of the Molokai by qualitatively comparing the area under the curve of both differential thermograms of the whole soils at the 550 C endotherm and the 900 C exotherm in Fig. 6.

A strong endotherm appears at 300 C in the Molokai thermogram implying the presence of a greater amount of gibbsite in this soil than in the Wahiawa. A broad band appearing between 600 and 800 C in the Molokai thermogram suggests the presence of amorphous-aluminum mineraloids or insufficient nitrogen pressure in the DTA chamber to suppress the oxidation of organic matter. The absence of this band in the Wahiawa is the only noticeable mineralogical difference between the two soils.

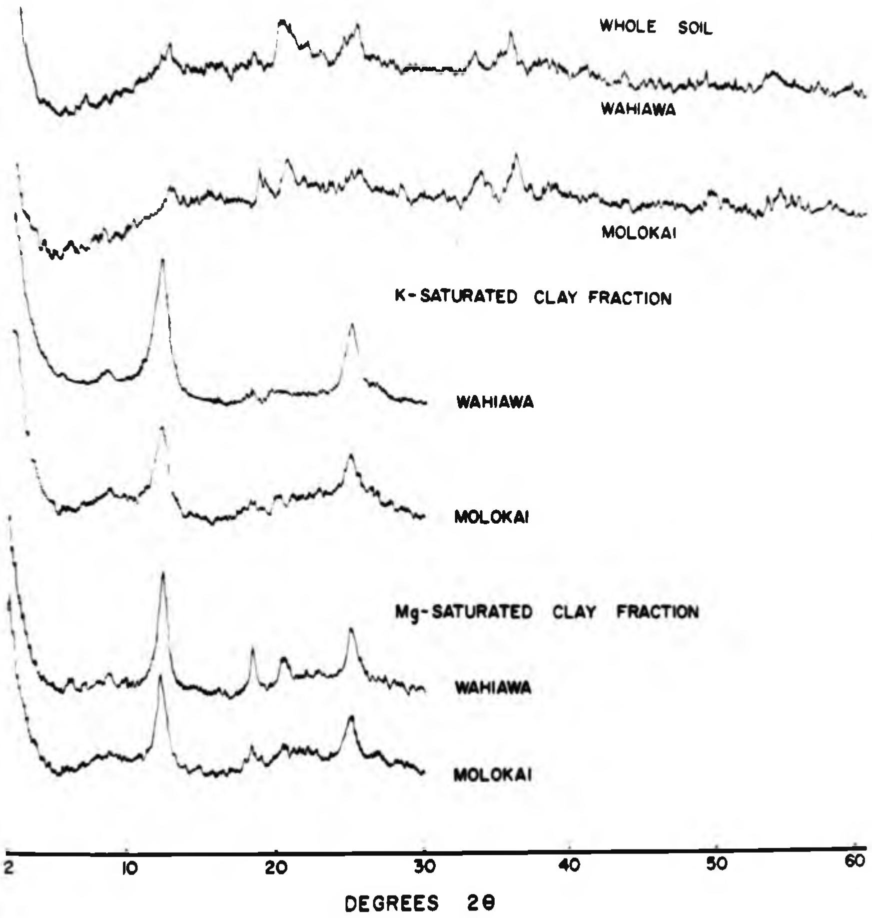

Fig. 5. X-ray Diffractograms of the Molokai and Wahiawa Soils

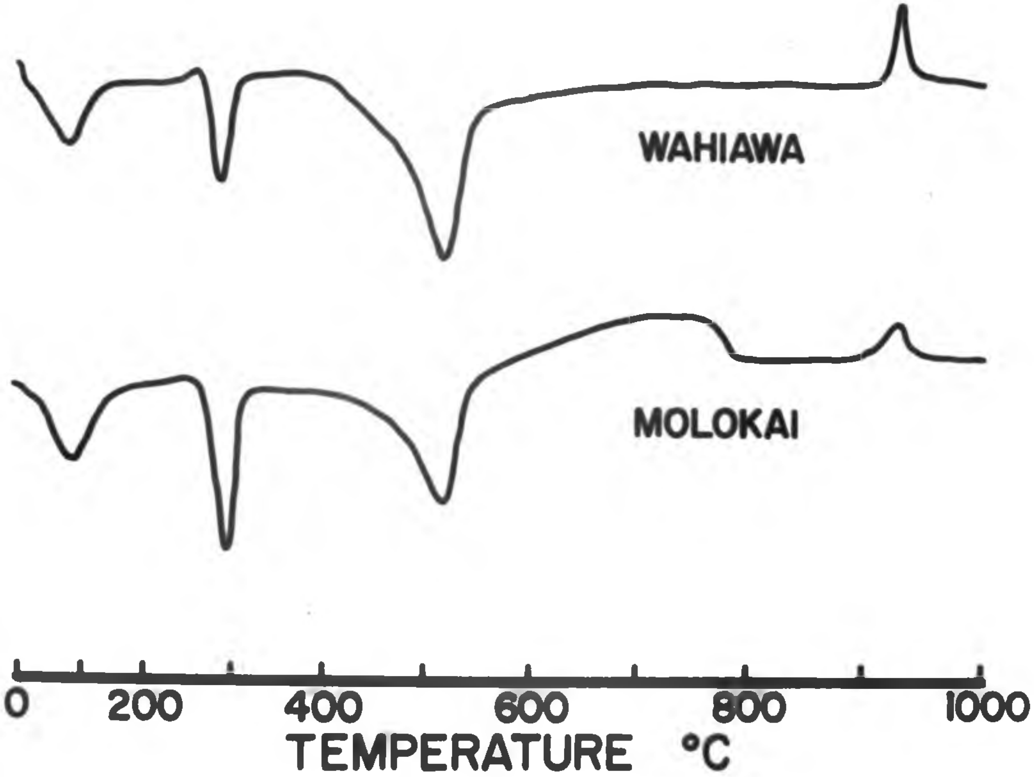

Fig. 6. Differential Thermograms of the Molokai and Wahiawa Soils

Also, the absence of any expanding 2:1 minerals is verified by the results obtained by Mg^{++}-saturating and glycolating the clay fraction of both soils.

B. Moisture Retention

Before undertaking a discussion of moisture retention, it is necessary to define some of the terms, particularly those dealing with soil structure, to be used in this and in subsequent sections.

A strict definition of structure is not available since it has been defined in numerous ways in the past. Perhaps the best is that of Brewer and Sleeman (1960): structure is defined as the physical constitution of a soil material as expressed by the size, shape, and arrangement of the solid particles and associated voids to form compound particles and the compound particles themselves. Soil fabric is the structural element that deals with the arrangement of the solid particles and associated voids. In this study, as in Sharma's (1966), soil fabric is subdivided into: (1) macro-fabric (inter-ped fabric) and (2) micro-fabric (intra-ped fabric). The macro-fabric is the arrangement of peds within the soil matrix while the micro-fabric deals with the internal arrangement of primary particles within the ped. The intra-ped fabric is the fundamental and characteristic property of the soil matrix and is not as easily subjected to change as the inter-ped fabric.

All of the moisture retention data presented in Figures 7 to 11 were obtained from samples equilibrated at pressures of zero to one bar. The volume of moisture (θ_V) held in a soil at any suction can be attributed to such factors as temperature, salt concentration, organic matter, mineral composition, and pore size distribution. Pore size distribution is the factor of primary concern in determining moisture retention if one assumes isothermal conditions, negligible amounts of salt and organic matter, and constant mineralogical content. Since Sharma (1966) has shown by mechanical analysis that both soils have similar texture, it follows that differences between soils can be attributed to soil structure and aggregate stability in particular. Water stable aggregates, disturbed < 2 mm samples, and undisturbed samples were used in an attempt to evaluate the effects of pore size distribution on moisture retention.

1. Moisture retention by different aggregate sizes.

The moisture retention curves of the hand-packed aggregate sizes, ranging from 10-20, 20-40, 40-60, and 60-140 mesh, which corresponds to fine gravel (2.0-1.0 mm), coarse sand (1.0-.5 mm), medium sand (.5-.25 mm), and fine sand (.25-.10 mm), for both soils were measured under nearly isothermal conditions and are presented in Figures 7 and 8.

The largest aggregate size had the highest saturation value as demonstrated in Fig. 7 for the Molokai soil. This evidence is supported by calculating the total pore space or per cent porosity from the experimentally determined bulk densities which decreased with an increase in particle size, i.e. (10-20) < (20-40) < (40-60) < (60-140). The bulk densities averaged 0.840 ± 0.03 g/cm^3. The total water holding capacity, determined by summing the total outflow volume and the residual volume of moisture released on oven-drying, was used in the saturation value. It usually was not less

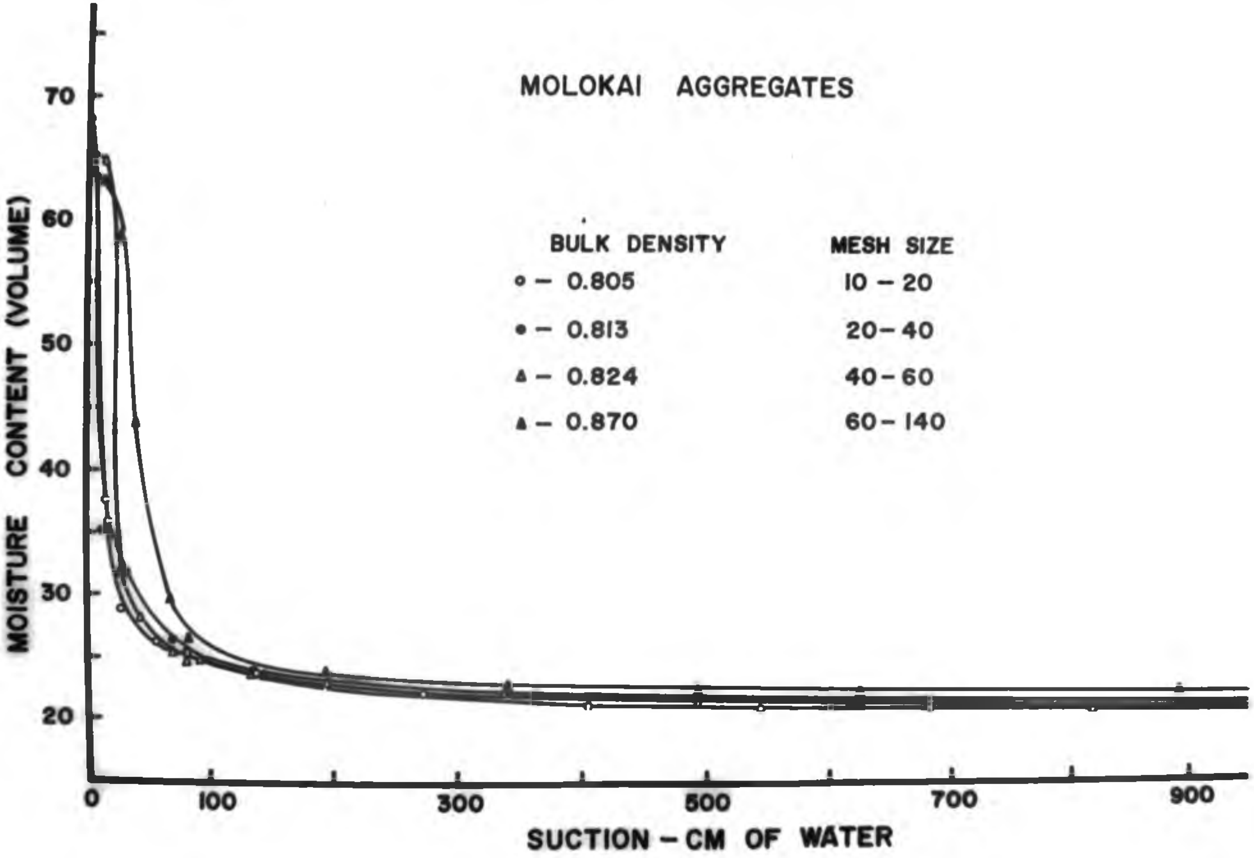

Fig. 7. Moisture Retention Curves for Water Stable Aggregates of the Molokai Soil

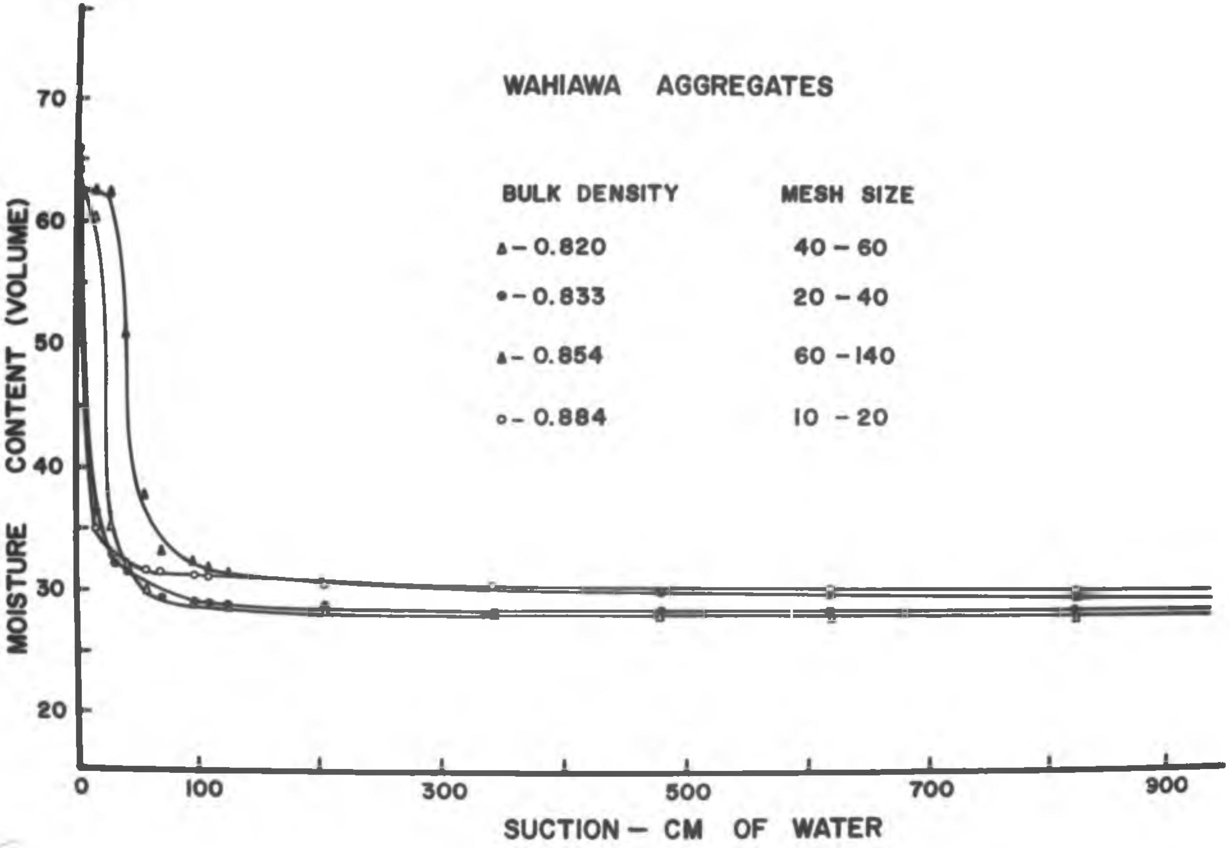

Fig. 8. Moisture Retention Curves for Water Stable Aggregates of the Wahiawa Soil

than 5 per cent of the calculated per cent porosity.

For the Wahiawa soil (Fig. 8), however, the 20-40 mesh size had the highest saturation value. This was expected, also, since the 20-40 mesh size had a lower bulk density of 0.833, while the 10-20 mesh size had a higher bulk density of 0.884. Bulk densities for the Wahiawa aggregates averaged 0.850 ± 0.03 g/cm^3 and increased as follows: (40-60) < (20-40) < (60-140) < (10-20). According to this arrangement, the 40-60 mesh size should have had the highest saturation value, decreasing to the 10-20 mesh size. This discrepancy can be attributed either to experimental error in packing or incomplete saturation of the inter-aggregate pores resulting from entrapped air. Very recently, Jackson and Klute (1967) proposed a method to estimate the dead-end pore volume in soils. They found that dead-end pore volume made up a small percentage of the total pore volume and is usually negligible.

In the suction range of 0-100 cm, there was a rapid release of water from both the Molokai and Wahiawa aggregates as demonstrated in Figures 7 and 8. In the Molokai the decrease in moisture volume was about 40 per cent, while in the Wahiawa it was about 30 per cent. For both soils, the 40-60 and 60-140 mesh sizes show definite impedance to flow in the low suction range. Intuitively, one might conclude that the impedance to flow is probably the result of the surface tension with which water is held in the smaller inter-aggregate voids. On the other hand, there is a rapid drainage of inter-aggregate water for aggregate sizes of 10-20 and 20-40 mesh, indicating a smaller surface tension on water molecules in the larger inter-aggregate pores. In the tension range of 100-1000 cm, the change in the volumetric moisture content was nil, implying the almost complete drainage of water from the inter-aggregate pores. The final moisture content for all aggregate sizes of both soils showed that the sample with the highest bulk density retained the greater amount of moisture in the suction range of 100-1000 cm of water.

A comparison of the retention curves of the aggregates for each soil shows that the Wahiawa soil retains as much as 5 per cent more water than the Molokai in regions of higher suction. The results confirm the earlier work of Cagauan (1963), and Cagauan and Uehara (1965), dealing with anisotropic behavior of strongly structured soils. The Wahiawa soil, because of its well-developed intra-ped fabric, is able to retain more water than the Molokai with its weakly developed intra-ped fabric. It is evident from the moisture retention results that the contribution of water held in the intra-aggregate pores to liquid flow can be considered negligible. The aggregates of both soils tend to retain from 25-30 per cent moisture which appears to be unavailable for liquid flow.

2. Moisture retention of < 2 mm whole soil.

The moisture retention curves of the disturbed samples of the Molokai and Wahiawa whole soil (< 2 mm) obtained from two different outflow apparatus are shown in Figures 9 and 10. The moisture retention curve (Fig. 9) obtained from the Tempe pressure cell gives a good indication of the aggregate stability of the soils. Previous measurements of aggregate stability have clearly shown

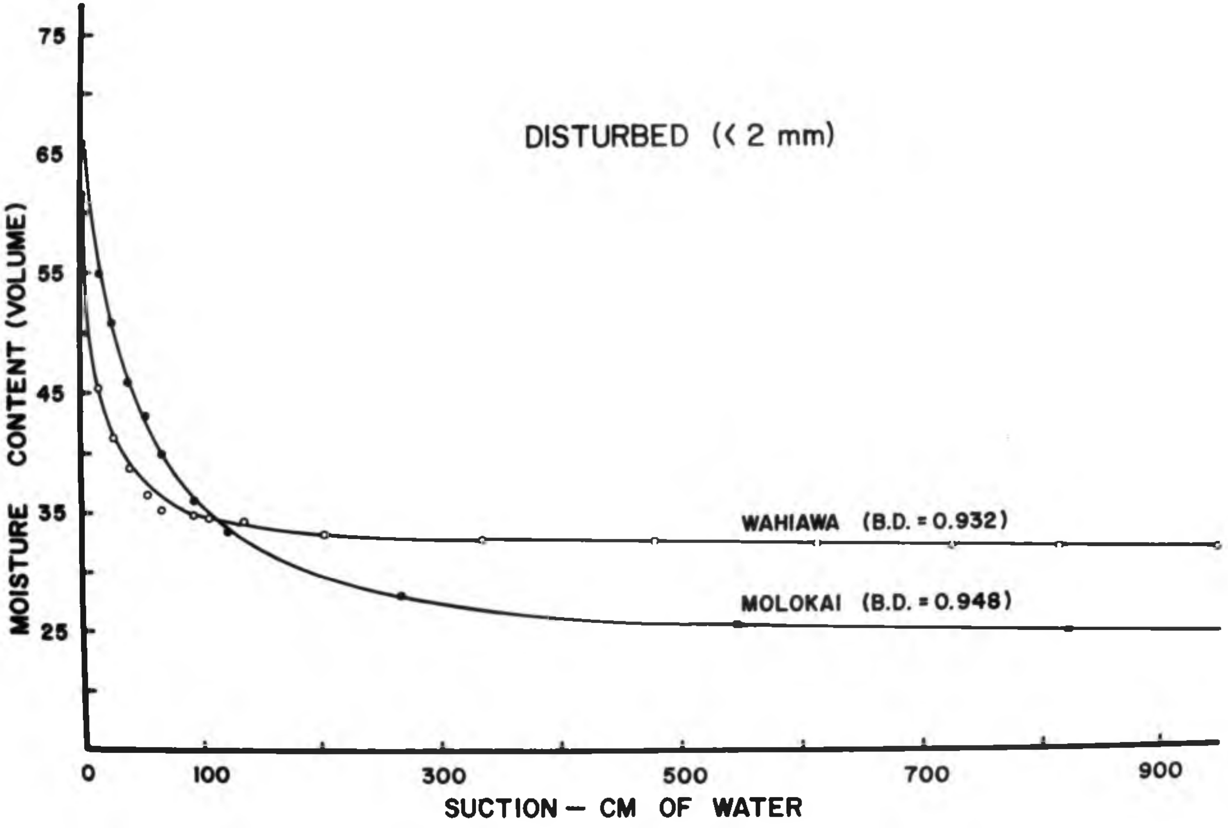

Fig. 9. Moisture Retention Curves for Whole Soil (<2 mm) of the Molokai and Wahiawa Soils Using the Tempe Pressure Cell

that the Wahiawa has a greater percentage of larger sized aggregates than the Molokai. As a result of this, the Wahiawa soil releases water held in the inter-aggregate pores at much lower suctions than the Molokai. However, the Wahiawa practically ceased to release water in the suction range of 100-1000 cm of water, indicating a nearly complete removal of water held in the pores of the interped fabric. The Molokai, on the other hand, continues to release water after 100 cm suction although the amount is small in comparison to the volume extracted at suctions below 100 cm of water. This is the result of smaller sized aggregates with smaller inter-aggregate pores releasing tightly held water when the suction or pressure on the system overcomes the capillary force of the system. As in the case of separate aggregate sizes, the Wahiawa soil retained up to 5 per cent more water than the Molokai.

Fig. 10 shows the moisture retention data obtained from the pressure plate extractor; the curves are similar to those obtained with the Tempe pressure cells. The only difference is that the curves in Fig. 10 have inflection points at moisture contents higher than those obtained from the Tempe pressure cells. The difference in moisture content at the inflection point was about 5 per cent. Accompanying differences in $K(\theta)$ and $D(\theta)$ for the two soils should be expected.

3. Retention data of Molokai < 2 mm and undisturbed core samples.

A comparison of the moisture retention curves of the Molokai hand-packed (< 2 mm) and undisturbed field cores are presented in Fig. 11. At zero suction, the < 2 mm sample had a higher

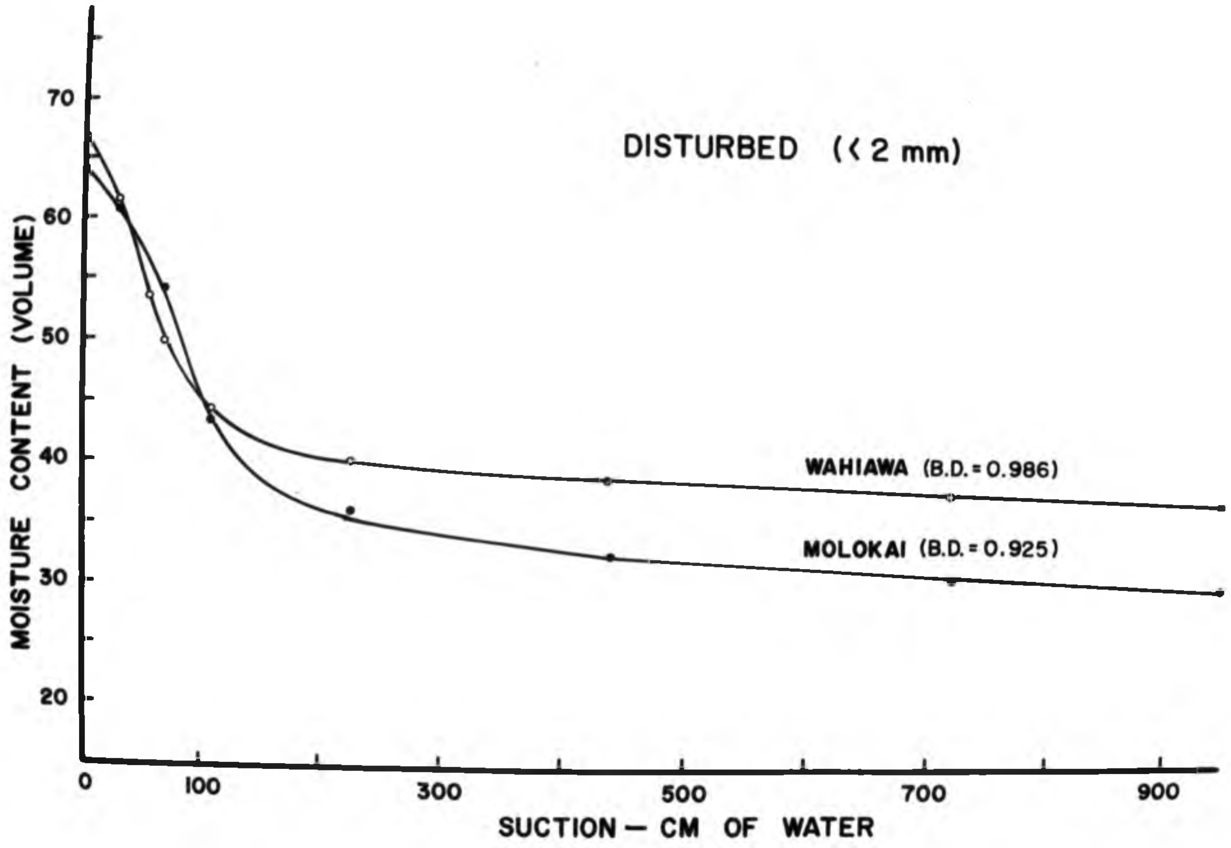

Fig. 10. Moisture Retention Curves for Whole Soil (<2 mm) of the Molokai and Wahiawa Soils Using the Pressure Plate Extractor

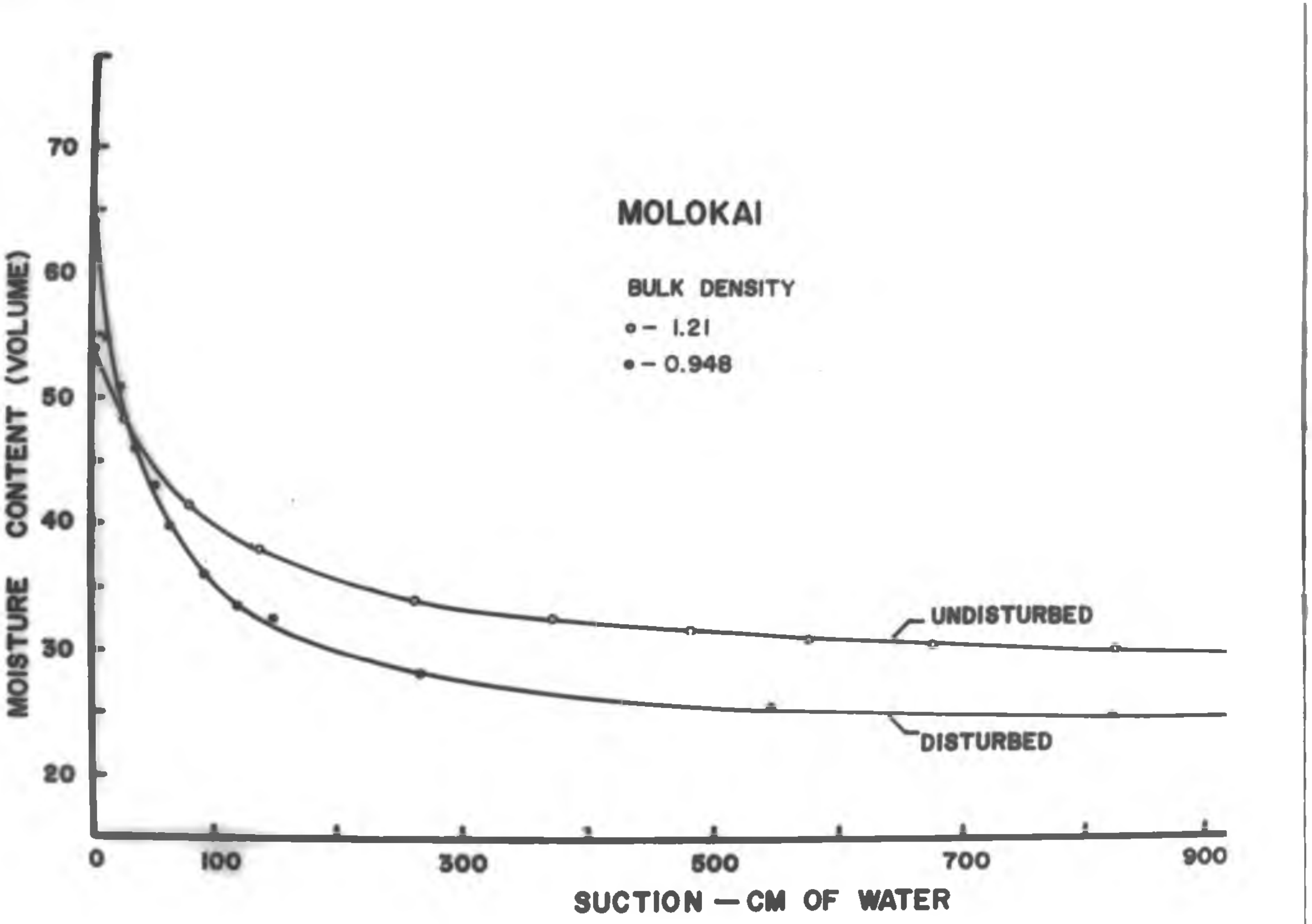

Fig. 11. Moisture Retention Curves for Disturbed (<2 mm) and Undisturbed Samples of the Molokai Soils

saturation value than the undisturbed core--a direct result of differences in bulk densities.

Examination of Fig. 11 reveals that the hand-packed sample releases water rapidly up to a suction of 100 cm while the undisturbed core released water gradually up to a suction of 300 cm. Beyond the suction range mentioned for both samples, outflow volume was almost negligible, as indicated by the general leveling off of the retention curve. The natural undisturbed core, with a higher bulk density, will obviously have smaller inter-aggregate pores than the hand-packed < 2 mm sample. Hence, drainage of the undisturbed core is expected to be complete at higher suctions.

The greater retention of water at higher suctions in the undisturbed core sample can be also attributed to aggregate sizes larger than 2 mm, whereas the undisturbed sample is composed of particles < 2 mm. Ekern (1966) reported that at least 47.3 per cent of the natural aggregates in the field are greater than 2 mm in diameter; hence, the crushed laboratory samples actually represent only one-half of the whole soil.

Further evidence dealing with the difference in pore size distribution between the two soils is presented in Figures 12 and 13. Using the method described by Brooks and Corey (1964) significant differences between the bubbling pressure, P_b, and the pore size distribution index, λ, were found in comparing the hand-packed samples of the Molokai and Wahiawa and the undisturbed sample of the Molokai. The P_b values for both Molokai samples were at least twice as great as that of the Wahiawa. The results suggest that

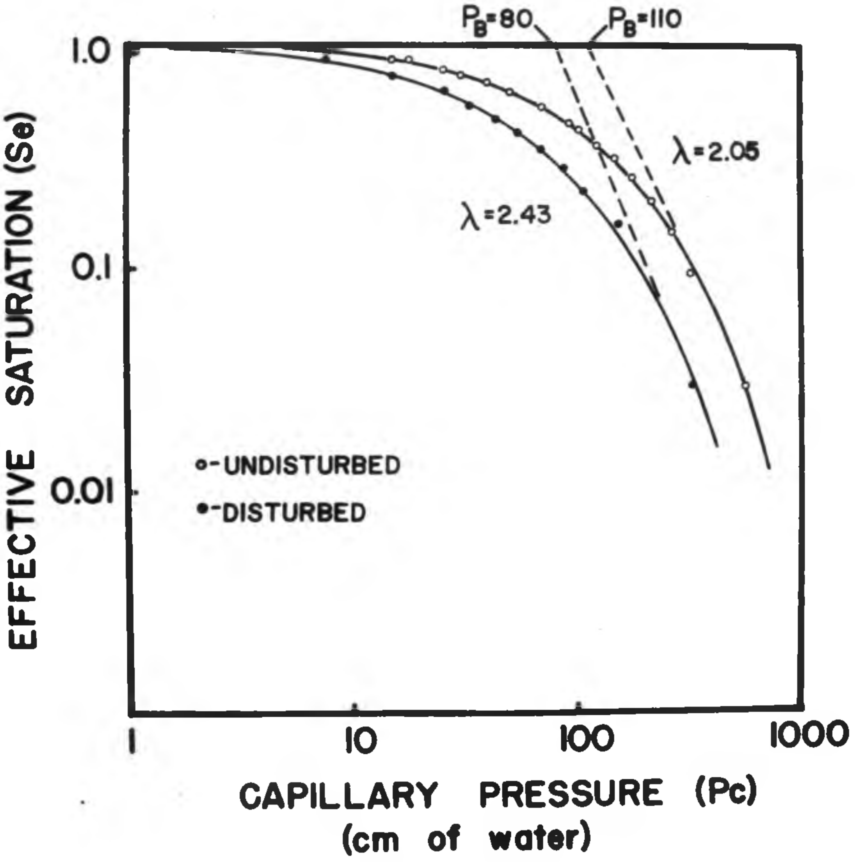

Fig. 12. Effective Saturation and Capillary Pressure Relationships for Disturbed (<2 mm) and Undisturbed Samples of the Molokai Soil

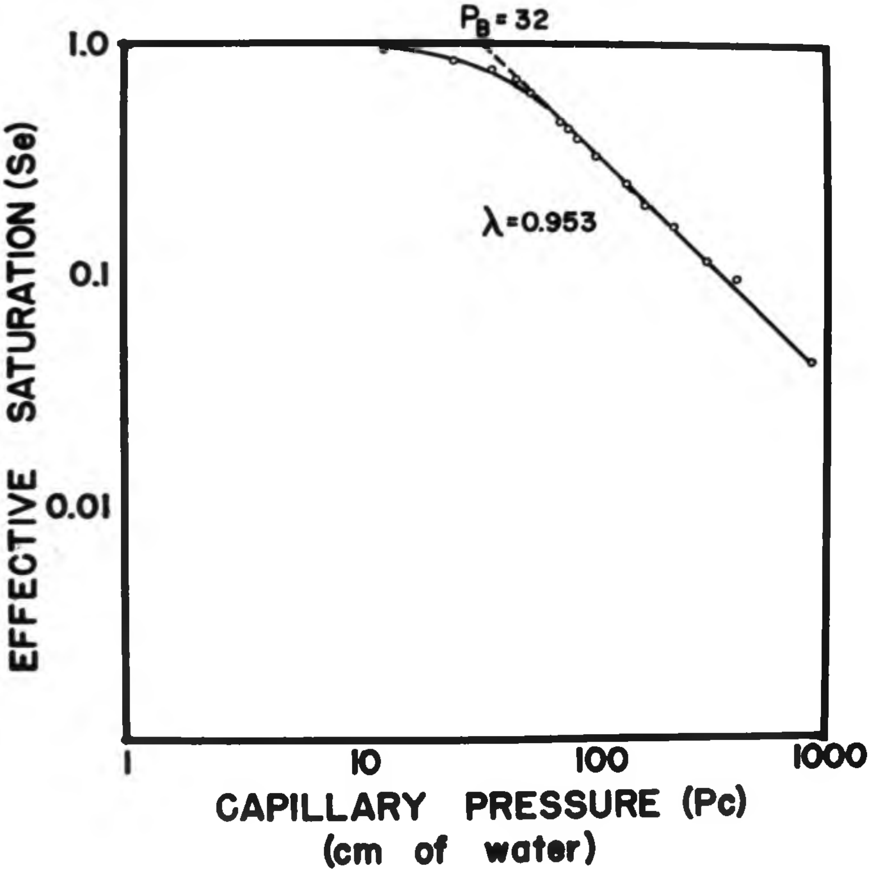

Fig. 13. Effective Saturation and Capillary Pressure Relationships for Disturbed (<2 mm) Samples of the Wahiawa Soil

continuous flow channels of water are present and exist at higher suctions in the Molokai than in the Wahiawa. The pore size distribution index, λ, as shown in the plot of effective saturation as a function of suction (Figures 12 and 13), is smaller for the Wahiawa than for both Molokai samples. The small value of λ indicates that the Wahiawa has a wider range of pore sizes while the Molokai tends to have a more uniform pore size distribution. The reliability of the P_b and λ values can be evaluated by carefully examining and comparing the retention curves of both soils in the suction range of 50 to 150 cm of water.

C. Evaluation of Transmission Coefficients

1. One-step Method. The results obtained by employing the one-step method are presented in Figures 14 to 24.

To study the effect of pore size distribution on $D(\theta)$, the one-step method was first applied to the different aggregate sizes. The results are shown in Figures 14 and 15. For all aggregate sizes, a significant linear relationship exists when $D(\theta)$ is plotted as a function of θ on a semi-logarithmic scale. Instead of plotting $D(\theta)$ as a function of θ "by eye," the method of least squares was employed to obtain the best-fit line through the experimental points. Although a regression equation results, it must be used with reservation, since θ is not a true independent variable measured without error, and the measured values of D at a given value of θ have associated errors which are not random (the error for a given measurement is usually in the opposite direction of the error for the previous measurement). Even so, the least-squares fitting

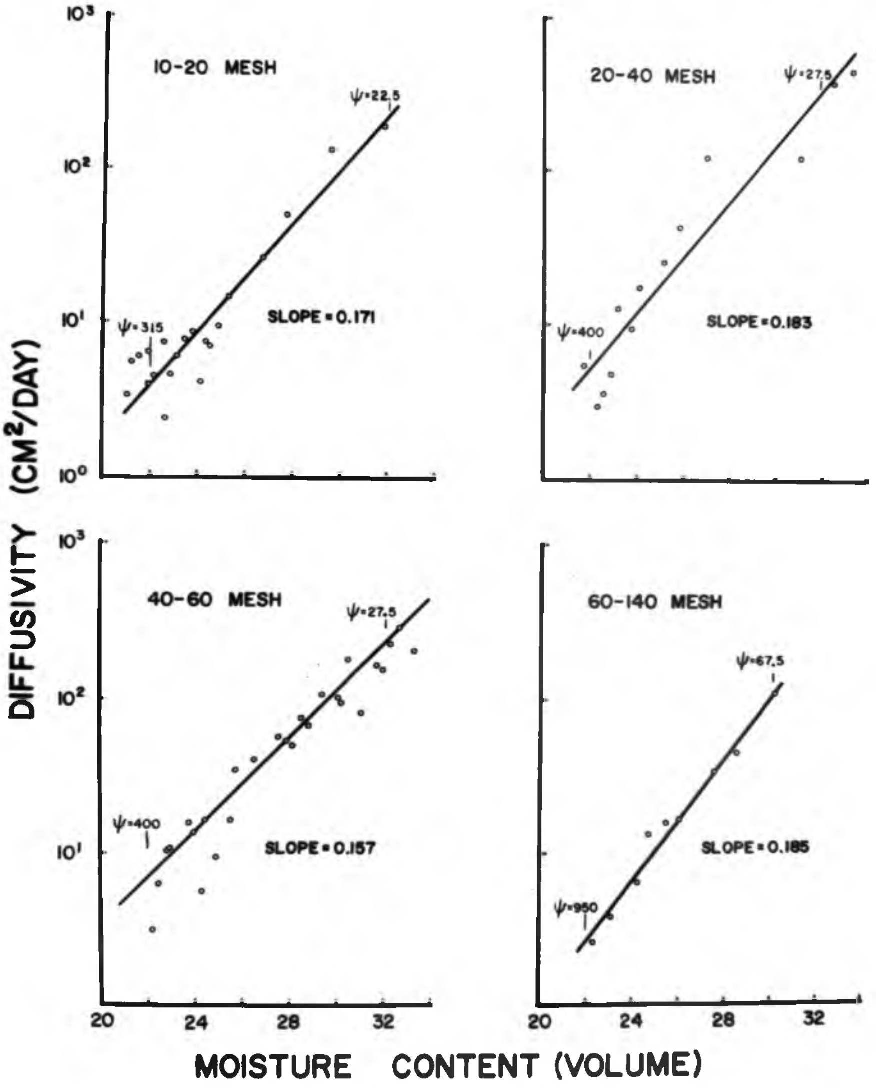

Fig. 14. Soil Water Diffusivity as a Function of Volumetric Water Content for Water Stable Aggregates of the Molokai Soil

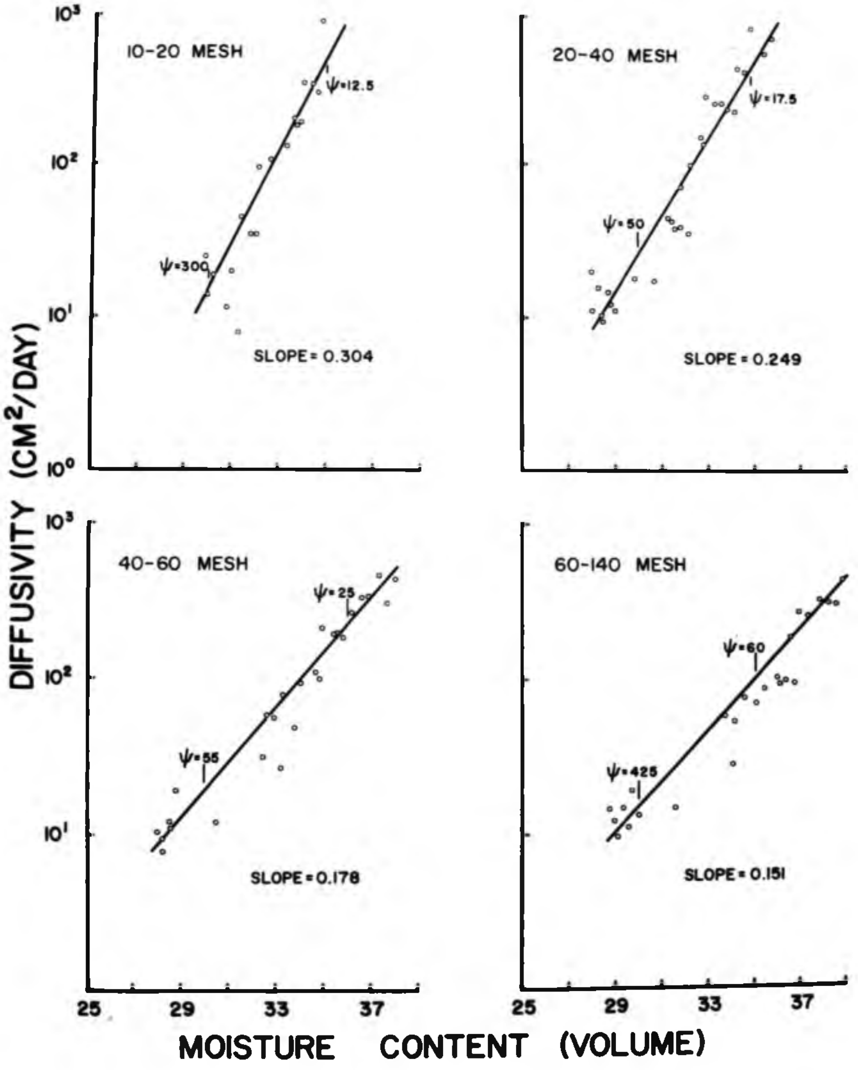

Fig. 15. Soil Water Diffusivity as a Function of Volumetric Water Content for Water Stable Aggregates of the Wahiawa Soil

procedure appears to provide an acceptable means of representing the D(θ) data with a single curve and a slope for comparative purposes.

For the Molokai aggregates (Fig. 14), the slopes (regression coefficient) for the different aggregate sizes were not significantly different. The coefficients had an average value of 0.170. The Wahiawa aggregates had a range of values from 0.304 for the 20 mesh sample decreasing to 0.151 for the 140 mesh sample. Small differences in slopes of duplicate runs can be attributed to variability in bulk density during packing. For the Molokai, the size of aggregate does not appear to affect the rate at which D(θ) changes with θ, indicated by the nearly identical slopes of the four aggregate sizes. The results appear to be valid in the moisture content range of 0.38 to 0.20 cm^3/cm^3 which corresponds to the suction range of 15 to 1000 cm of water.

The increase in slope with a corresponding increase in aggregate size implies that the size of aggregate plays an important role in water transport for the Wahiawa soil for the moisture range of 0.37 to 0.29 cm^3/cm^3, corresponding to suctions of 27 to 1000 cm of water. Contrary results from both soils can be attributed to differences in structure--intra-ped fabric in particular.

In light of the work of Cagauan and Uehara (1965) the size of aggregate should not play an important role in the Molokai soil because of its weakly developed or nearly isotropic structure. Because of the well-developed intra-ped fabric of the Wahiawa soil, differences in slopes of each aggregate size were anticipated.

Capillary conductivity was calculated for each of the aggregate sizes by first selecting a $D(\theta)$ value at a specific moisture content, then calculating the slope of the moisture retention curve at that moisture content. Capillary conductivity was then obtained from equation (6). The results of the calculations are presented in Figures 16 and 17, where $K(\theta)$ is plotted as a function of θ, and in Figures 18 and 19, where $K(\theta)$ is plotted as a function ψ, for selected aggregate sizes.

In all instances, there was a simultaneous decrease of $K(\theta)$ with a decrease of θ and an increase in ψ. As predicted by the moisture retention curves, $K(\theta)$ for different aggregate sizes converged as θ decreased and ψ increased.

Similar significant linear relationships were obtained when $D(\theta)$ was plotted as a function of θ for the < 2 mm samples of both the Molokai and Wahiawa and the natural undisturbed core of the Molokai. The results are presented in Figures 20, 21, and 22. Slopes of the Molokai, < 2 mm and undisturbed core, were not significantly different from each other but considerably lower than the previously obtained values of the separate aggregates. The lower values of slope are probably due to the presence of small particles (> 140 mesh) and to the dispersion of water-unstable particles. Similarly, the slope of the Wahiawa < 2 mm sample was lower than those of the separate aggregates. In Fig. 20, the curve can be extended to higher moisture contents by using monel plates (lower impedance). Ceramic plates yield values that do not agree with assumption, $\alpha^2\ Dt > 0.3$.

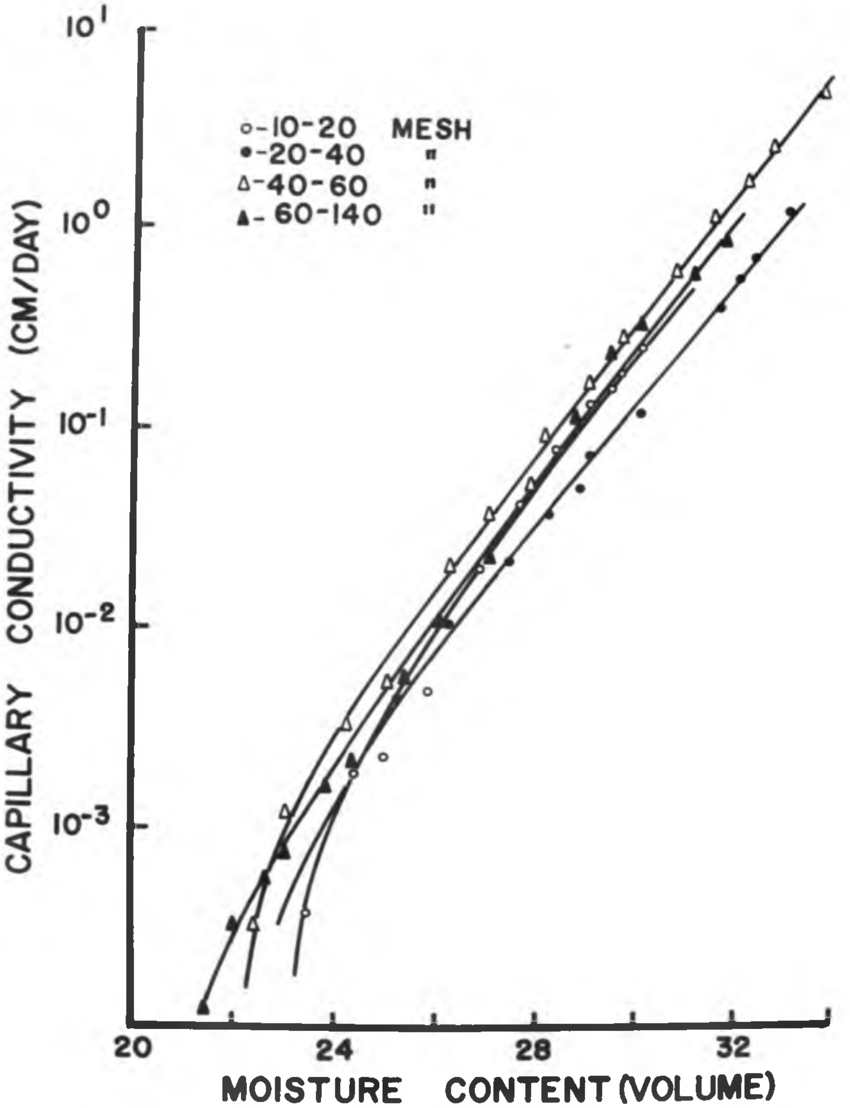

Fig. 16. Capillary Conductivity as a Function of Volumetric Water Content for Water Stable Aggregates of the Molokai Soil Calculated from Figures 7 and 14

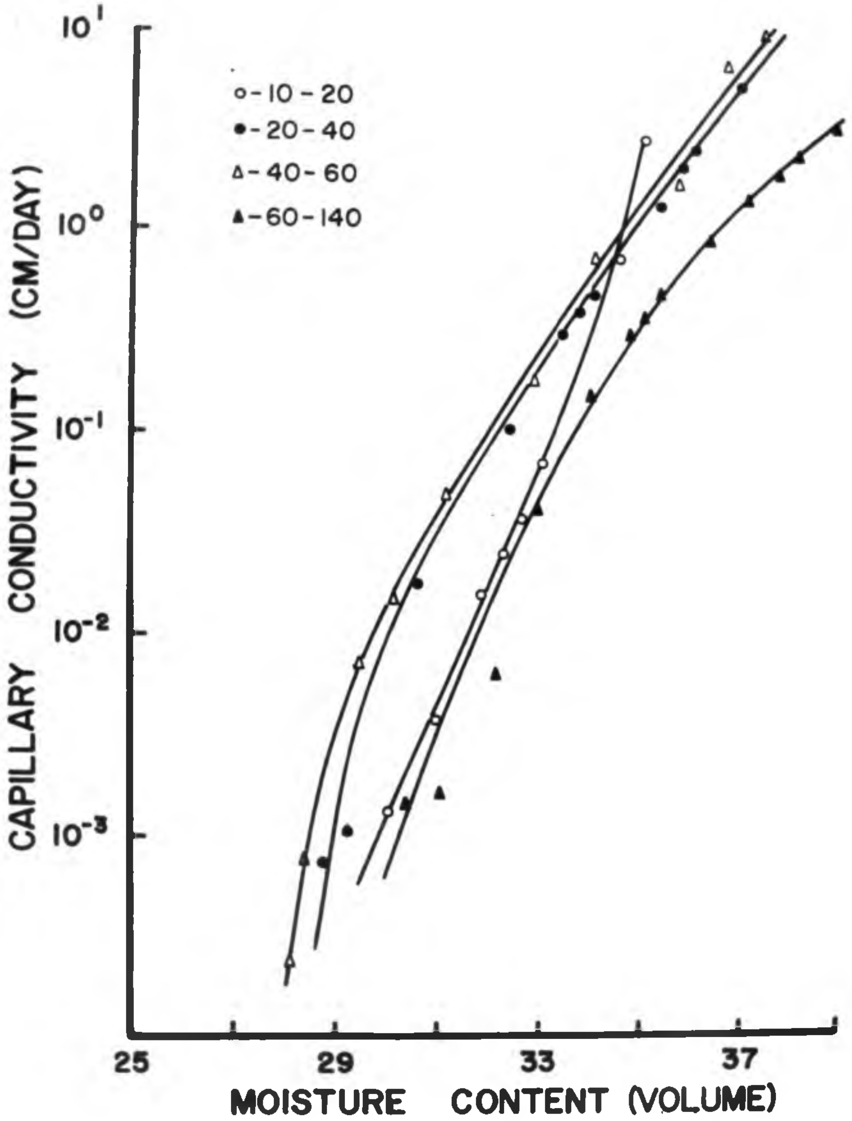

Fig. 17. Capillary Conductivity as a Function of Volumetric Water Content for Water Stable Aggregates of the Wahiawa Soil Calculated from Figures 8 and 15

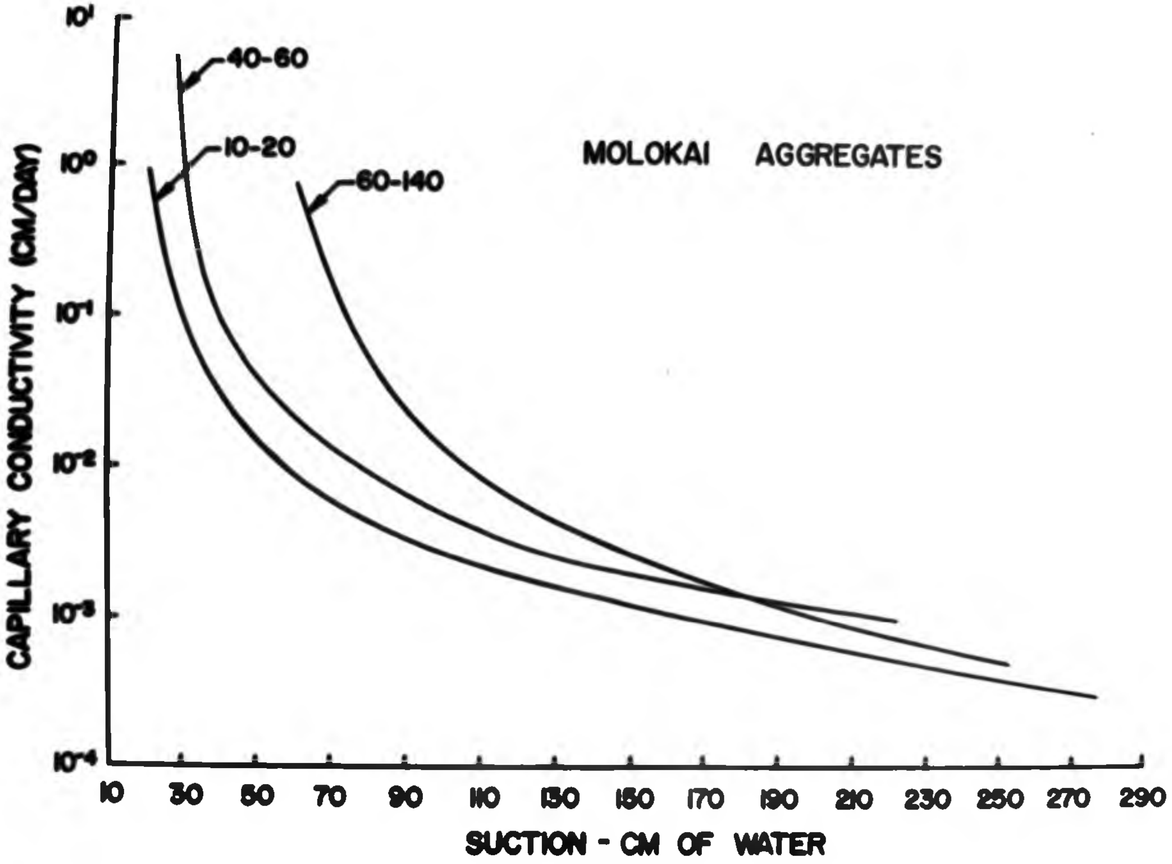

Fig. 18. Capillary Conductivity as a Function of Soil Moisture Suction for Water Stable Aggregates of the Molokai Soil Calculated from Figures 7 and 14

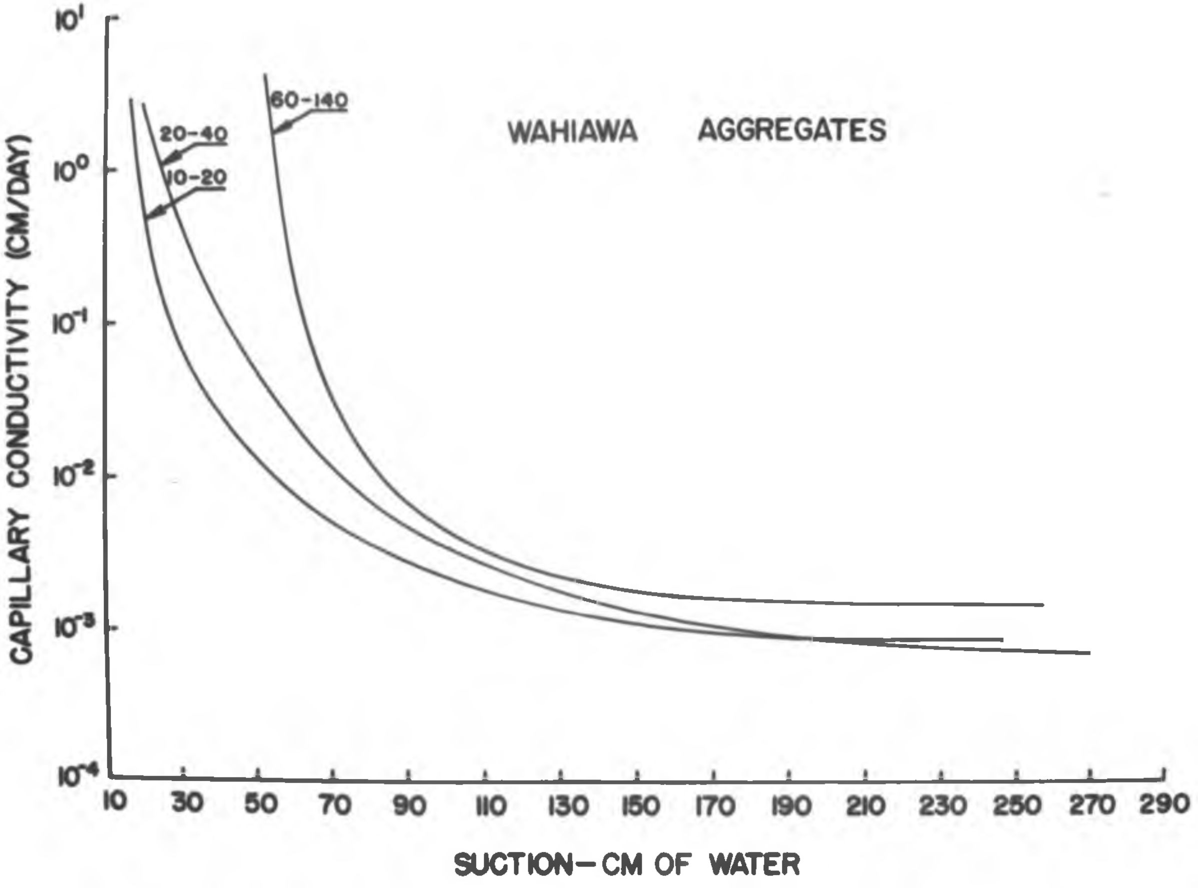

Fig. 19. Capillary Conductivity as a Function of Soil Moisture Suction for Water Stable Aggregates of the Wahiawa Soil Calculated from Figures 8 and 15

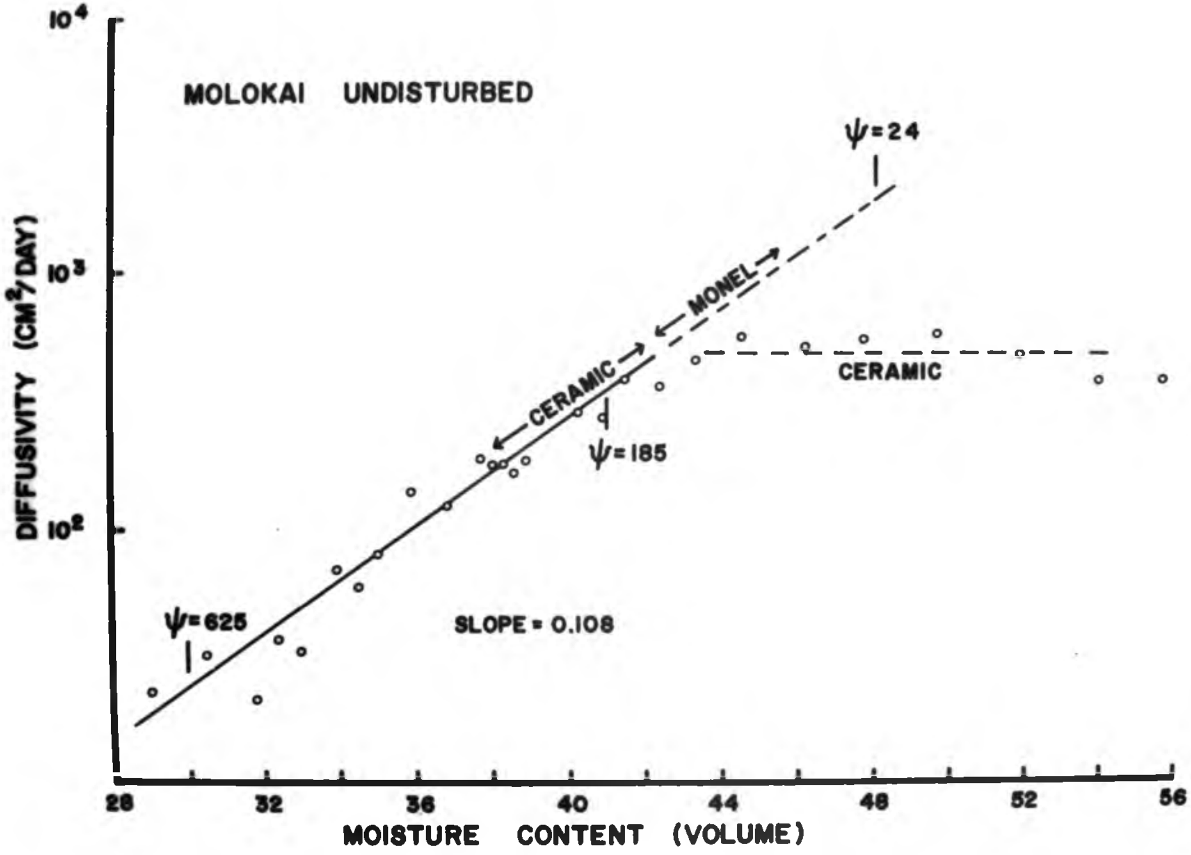

Fig. 20. Soil Water Diffusivity as a Function of Soil Moisture Content for Undisturbed Samples of the Molokai Soil

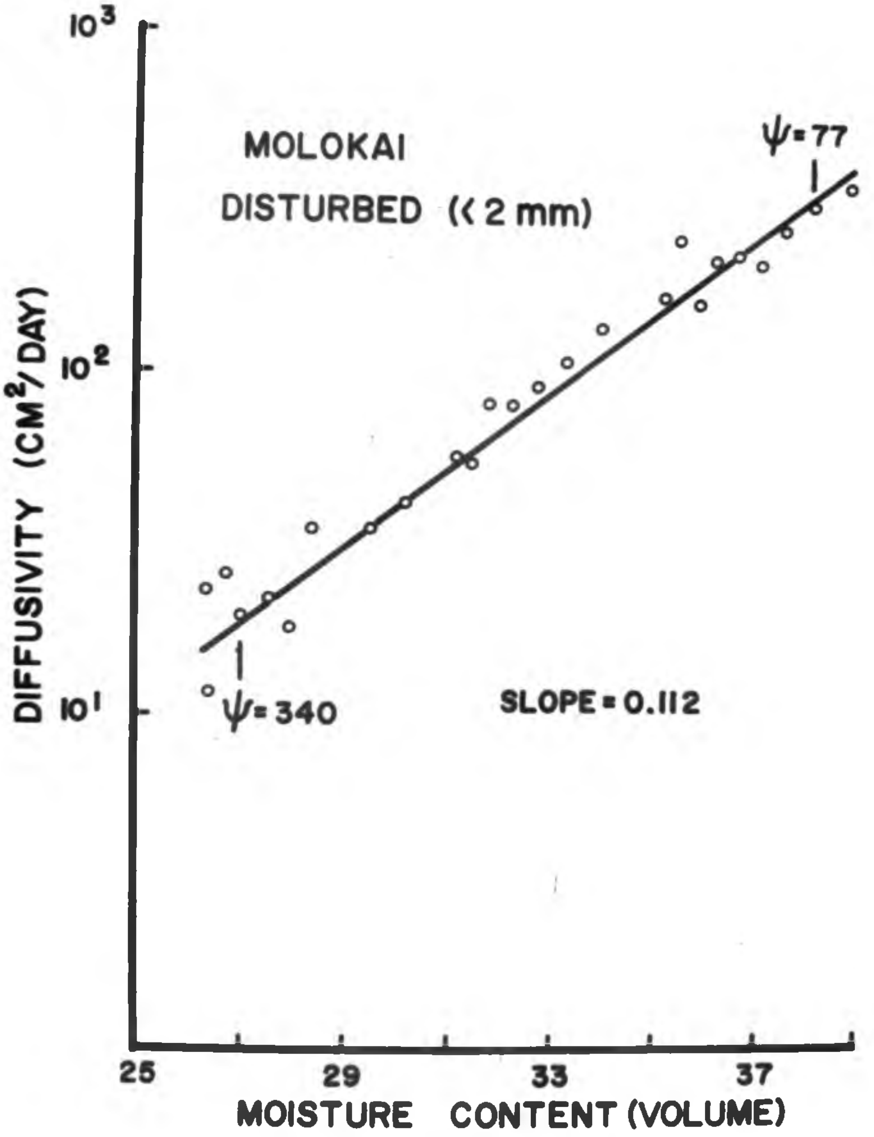

Fig. 21. Soil Water Diffusivity as a Function of Volumetric Water Content for Disturbed (<2 mm) Sample of the Molokai Soil

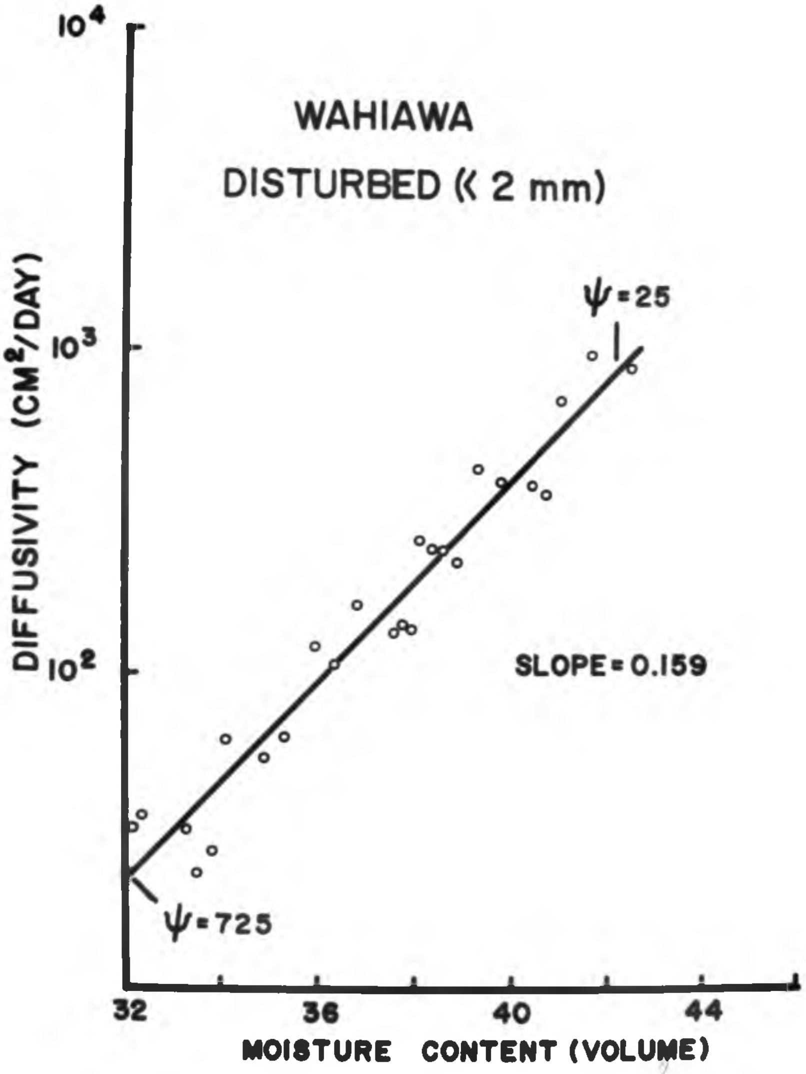

Fig. 22. Soil Water Diffusivity as a Function of Volumetric Water Content for Disturbed (<2 mm) Sample of the Wahiawa Soil

Capillary conductivity results were calculated, as previously described, and are represented in Figures 23 and 24. Inter-ped and intra-ped fabric again play significant roles in determining the rates at which $K(\theta)$ decreases with ψ. Note that $K(\theta)$ of the Wahiawa (Fig. 23) decreases at much lower suctions than $K(\theta)$ for the Molokai. However, conduction ceases at about 400 cm suction while the Molokai continues to conduct water at suctions up to 900 cm. Also, comparison of Figures 23 and 24 shows that $K(\theta)$ of the Molokai < 2 mm decreases much more rapidly than the undisturbed core. By extrapolating the curve of the undisturbed core in Fig. 24 to 900 cm suction, the $K(\theta)$ for both the < 2 mm and undisturbed core compare very favorably.

For comparison with the multi-step method, $K(\theta)$ was plotted as a function of θ in Fig. 25. Differences between the curves can be attributed to variation in bulk density. The curves tend to converge at low θ and high ψ as can be predicted from the moisture retention data.

2. Multi-step Method. To compare water transfer coefficients determined by the one-step method with independent measurements by another transient outflow technique, the so-called multi-step method developed by Gardner (1956) and simplified by Kunze and Kirkham (1962) was employed. This method utilizes successive increments of pressure with as many equilibration steps. The locally fabricated pressure plate extractor (Fig. 2) was used for these determinations. The critical point of the multi-step method is matching of the experimental outflow data with the theoretical transients. After

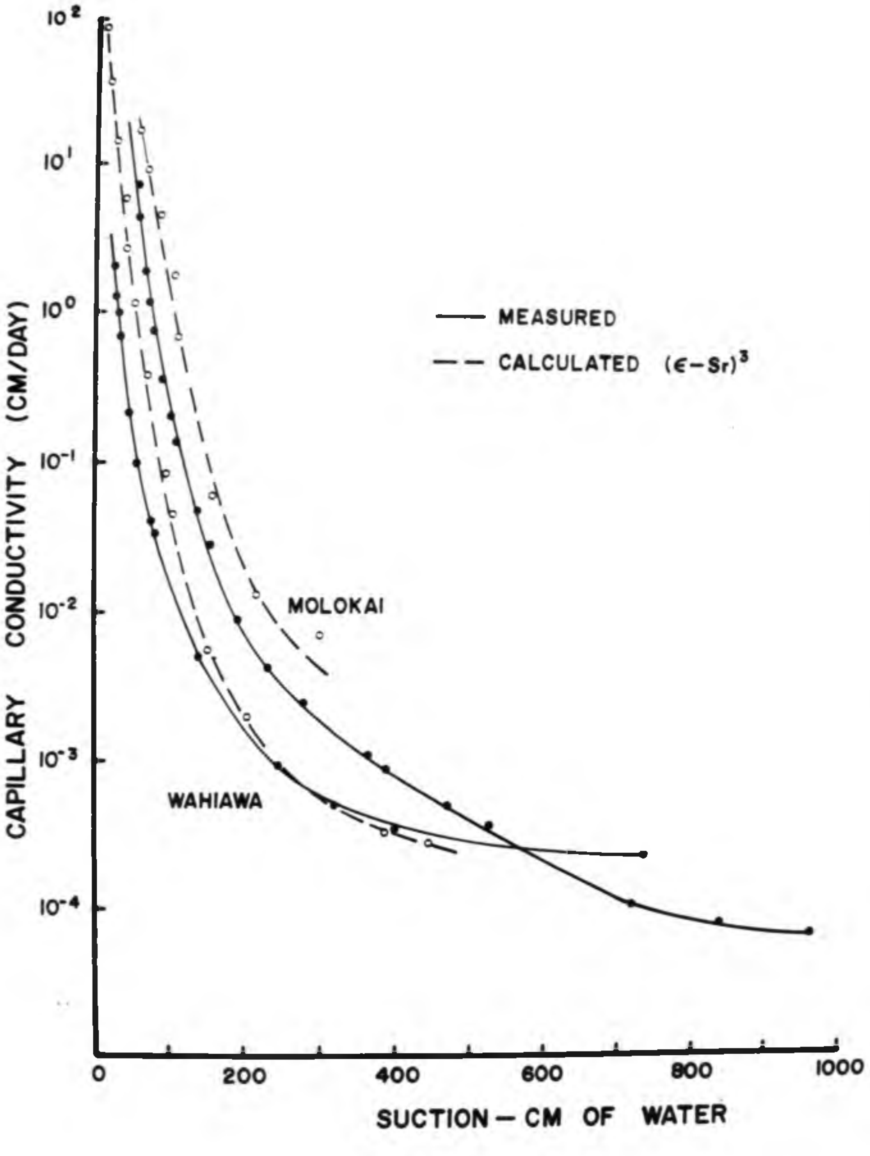

Fig. 23. Comparison of Measured and Calculated Capillary Conductivity of Disturbed (<2 mm) Samples of the Molokai and Wahiawa Soils

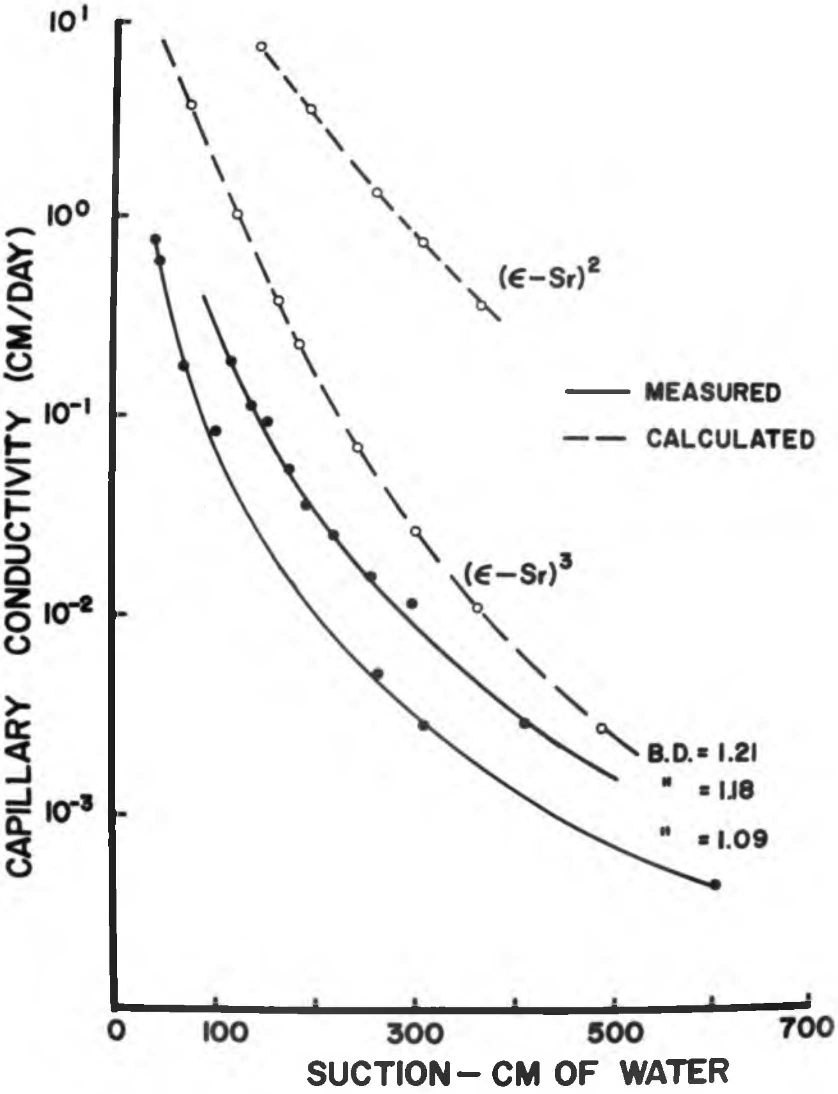

Fig. 24. Comparison of Measured and Calculated Capillary Conductivity of Undisturbed Samples of the Molokai Soil

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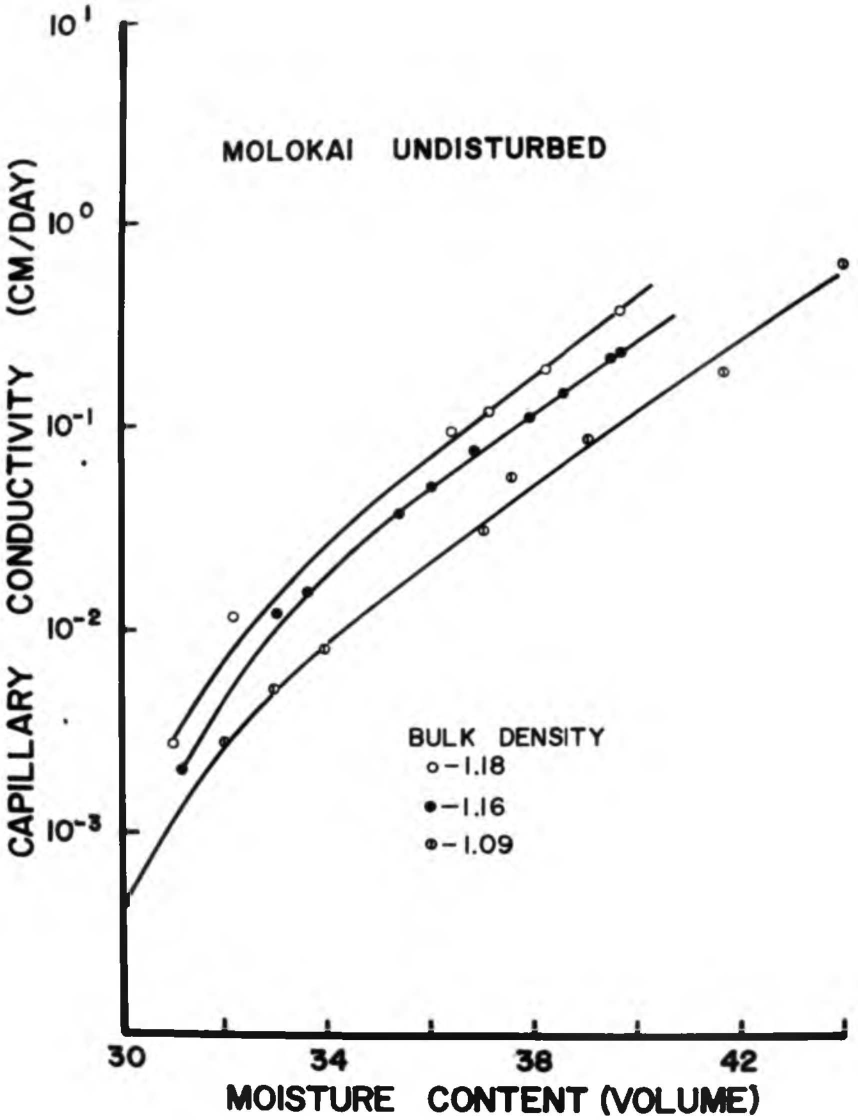

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Fig. 25. Capillary Conductivity as a Function of Volumetric Water Content for Undisturbed Samples of the Molokai Soil Calculated from Figures 11 and 20

coincidence of both experimental and theoretical horizontal axes of the plot of Q/Q_0 as a function of α_1^2Dt/L^2 was established, comparison of theoretical and experimental transients was obtained by matching the experimental points on to the theoretical transients. In most instances, the experimental points did not coincide with the theoretical over the entire curve, as shown in Fig. 26. If the initial stages of the experimental outflow were fitted to the theoretical, the later stages of outflow deviated considerably from the expected. Similarly, if the later stages were fitted, the early stages deviated from the theoretical. To be consistent, matching of initial stages of outflow to the theoretical was desirable since it was measured with greater accuracy. In matching the experimental with the theoretical, Kunze and Kirkham's (1962) method did not compensate for experimental outflow data equilibrated at suctions greater than 450 cm of water. As shown in Fig. 26, the theoretical transient, associated with $a=0$, is considerably lower than the experimental. Kunze and Kirkham (1962) fitted their experimental outflow data to the theoretical transients and obtained reasonably good estimates of $K(\theta)$ and $D(\theta)$. Later, Green, et al. (1964) compared the calculated $K(\theta)$ and $D(\theta)$ from actual field measurements of infiltration with those calculated from the theoretical method of Kunze and Kirkham (1962) and obtained good correlative results.

The parameters required to calculate the transmission coefficients are t_{RP} and α_1^2. Parameter t_{RP}, reference-point time, was determined by taking the actual time value that coincided with reduced time, α_1^2Dt/L^2, at 1.0. Parameter α_1^2 was determined by the

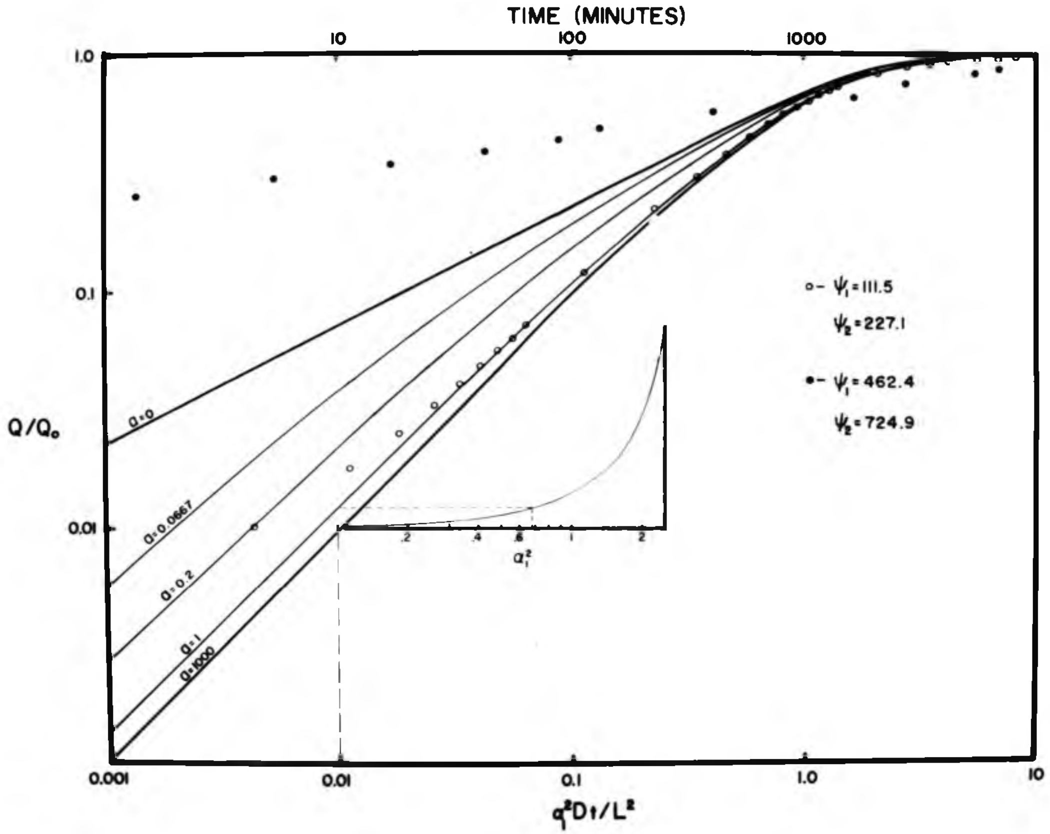

Fig. 26. Dimensionless Experimental Outflow Transients of Disturbed (<2 mm) Samples Fitted to Theoretical Outflow Transients

small inset graph in Fig. 26.

The calculated values of K(θ) and D(θ) by the multi-step method resulted in an apparently inconsistent relationship of both transmission coefficients with suction. A wide range of variability, as much as ten-fold differences, (Fig. 26), of K(θ) and D(θ) resulted from a single experimental outflow transient when the early or late stage of outflow was matched with the theoretical.

By taking the apparent best estimate of K(θ) for each transient match, K(θ) was plotted as a function of θ. Fig. 27 shows the decrease of K(θ) with a decrease in θ for both soils. The K(θ) of the Wahiawa was found to be less than that of the Molokai at equivalent moisture contents. The curve of K(θ) as a function of θ could not be established at lower values of θ, corresponding to suctions greater than 450 cm, because the experimental outflow points could not be fitted to the theoretical. Conceivably, this is the result of a small, total equilibrium outflow volume at higher suctions, where the initial outflow volume would constitute a large fraction of the total, Q_o. Hence, the ratio, Q/Q_o, would be too large and impossible to match with theoretical transients. The results of plotting K(θ) as functions of θ and as functions of ψ obtained by the one-step method (Figures 16, 17, and 25) compare favorably with a plot of K(θ) with by the multi-step method (Fig. 27).

Because a consistent D(θ) versus θ relationship could not be established by the multi-step method in the range of suctions utilized, conclusive comparison of D(θ) by the two methods was not possible.

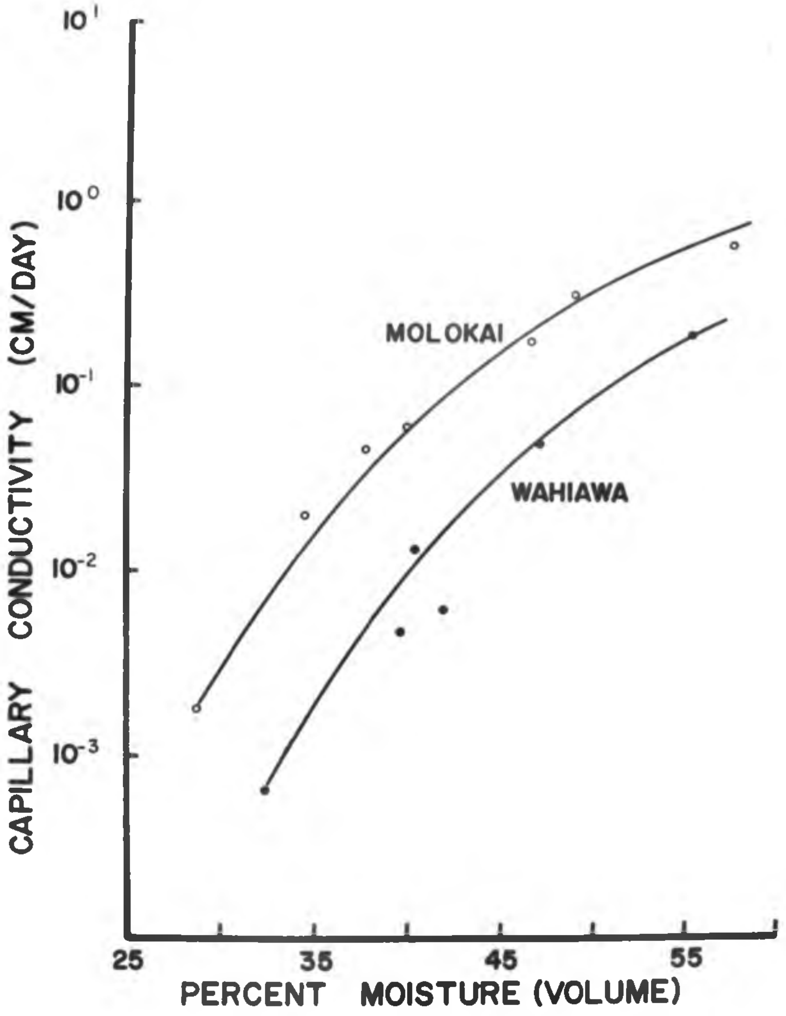

Fig. 27. Capillary Conductivity as a Function of Volumetric Water Content for Disturbed (<2 mm) Samples of the Molokai and Wahiawa Soils

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3. Calculation Method. Since previous investigations by Sharma (1966) on Hawaiian latosols revealed that the CCG, M, MQ, and BC methods of calculating K(θ) yielded over-estimates of the K(θ) evaluated experimentally, the modified equation developed by Sharma

$$K = 2.79 \times 10^{2} (\varepsilon - S_{r})^{2} n^{-2} [h_{1}^{-2} + 3h_{2}^{-2} + \ldots + (2n - 1) h_{n}^{-2}]$$

was used in calculating K(θ) in this study. Calculated K(θ) values were then compared with the experimentally obtained K(θ) of the < 2 mm samples of the Molokai and Wahiawa and the undisturbed sample of the Molokai by the one-step method, (Figures 23 and 24).

The modified equation of Sharma, however, also proved to be inadequate since the calculated K(θ) values were greater than the experimental. After much trial and error, the best fit of experimental data to calculated data was obtained by increasing the exponent of $(\varepsilon - S_r)$ from 2 to 3. Millington and Quirk (1961) established by probability theory that the exponent of ε should not be greater than 2. The failure of tropical soils to agree with the assumptions set forth by the MQ method suggests that ε or $(\varepsilon - S_r)$ be used as a mere matching factor. Furthermore, the MQ method does not involve the use of the S_r term.

The MQ method, as well as the CCG and M methods, were originally applied to sand and slate dust where the porous media could be considered to have an (1) isotropic distribution of total pore space and a (2) random distribution of pore size classes. Sharma's modified equation, although applied to soil material, describes

$K(\theta)$ adequately only for water stable aggregates where the assumptions of the MQ method could be fulfilled. In considering the whole soil of both the Molokai and Wahiawa, however, the assumptions are hardly expected to apply. Both the disturbed and undisturbed soil samples employed in this study are neither completely water stable nor inert; hence, it follows that the existence of an isotropic distribution of total pore space is untenable.

Comparison of the calculated and the experimental $K(\theta)$ curves, using $(\varepsilon - S_r)^3$, are presented in Figures 23 and 24. The slope of the calculated curves tend to coincide with the experimental data.

The S_r values were obtained by the method described by Brooks and Corey (1964). The S_r values for the Molokai < 2 mm, the Molokai undisturbed, and the Wahiawa < 2 mm were found to be 0.561, 0.573, and 0.515, respectively. These S_r values were used to calculate P_b and λ in Figures 12 and 13. The S_r values reveal the fraction of the total water held in the soil that is unavailable (intra-aggregate pore water) to flow.

4. Infiltration Column. In an attempt to study $K(\theta)$ at low suctions, the steady-state experimental set-up shown in Fig. 4 was used. Approximately 1.6 kg of disturbed (<2 mm) Molokai soil were required to fill the column to a bulk density of 1.2 g/cm^3. Darcy's equation was used to calculate $K(\theta)$ from the measured ψ and x values. An inconsistent relationship between $K(\theta)$ and ψ resulted from the calculation, possibly due to the absence of a unit hydraulic gradient within the column. Calculated $K(\theta)$ values, however, compared favorably to those obtained by transient methods. Errors

in this method arise from the poor contact of soil water with the tensiometer membrane interface. The unique structure of Hawaiian latosols makes the error resulting from poor interface contact very real and hence limits the use of this method. Since time was also a limiting factor, a comprehensive examination of this method was not possible.

D. Evaluation of Methods.

In determining the transmission coefficients of Hawaiian latosols, the one-step method has been found to yield apparently reasonable results in the suction range of zero to one bar. Favorable comparisons of both the multi-step and the calculation methods substantiate the reliability of the one-step method.

The major advantages of using the one-step method can be enumerated as follows: (1) a saving of considerable time since only 48 hours is usually required for equilibrium, and $D(\theta)$ can be determined directly from the dimensions of the outflow unit by equation (19); (2) $K(\theta)$ is readily calculated from moisture retention data and the previously determined $D(\theta)$; and (3) a small sample size is required when the one-step method is used in conjunction with the Tempe pressure cell.

In comparison to the multi-step method, the distinct advantage of the one-step method is that it eliminates the tedious and questionable matching of experimental points with theoretical outflow transients. The non-coincidence of the experimental outflow transient with the theoretical over the whole range are further evidence, as previously shown by Elrick (1963), Jackson, *et al.* (1963), and Bazargani (1964), of the inadequacies of present outflow methods utilizing porous plates.

However, for Hawaiian latosols, because of their unique structural development, the one-step method possibly provides a rapid and reasonable means of obtaining data on $D(\theta)$ and $K(\theta)$ for qualitative prediction of water movement. The use of calculation methods, though not as desirable as the experimental, can be used to determine $K(\theta)$ and $D(\theta)$ from moisture retention data, thus saving considerable time. Calculation methods also eliminate the need for tedious outflow measurements in the laboratory.

SUMMARY AND CONCLUSIONS

The conclusions stemming from the results of this study can be enumerated as follows:

1. The one-step method is the best available method to evaluate transmission coefficients of soil water for Hawaiian latosols in the suction range from 0.015 to 1.0 bar. The one-step method is preferred to the multi-step method because the former requires less time and eliminates error introduced by matching experimental with theoretical transients.
2. The multi-step method, although capable of measuring small volumes of outflow accurately, requires long periods of time for complete equilibrium for each pressure increment. Also, matching of theoretical and experimental outflow transients was not only tedious and time consuming, but the poor fit over the entire range of reduced time left much to be desired.
3. Accurate determination of $K(\theta)$ and $D(\theta)$ in the saturated or near saturated region is best described by steady-state methods developed by Elrick (1963) and tested by Sharma (1966) for Hawaiian latosols.
4. Calculation methods, utilizing the modified equation as previously described herein, provides an accurate means of determining $K(\theta)$ and $D(\theta)$ for the < 2 mm samples of the Molokai and Wahiawa and undisturbed core samples of the Molokai from pore size distribution data (moisture retention data). Hence, for Hawaiian latosols, reasonable, qualitative estimation of soil water transmission

coefficients can be made from a pre-determined moisture retention curve.

5. Structure, particularly the intra-ped fabric, plays a definite role in moisture retention and hence, moisture movement. The Molokai, because its structure is weakly developed in comparison to the Wahiawa, retains about five per cent less moisture at nearly all suctions. The effect of intra-ped fabric is very evident at higher suctions while inter-ped fabric plays a significant role at suctions less than 100 cm of water.
6. A comparison of disturbed and undisturbed samples of the Molokai soil gave analogous results. This indicates that natural undisturbed core samples may not be required for determinations of transmission coefficients in some cases.
7. The infiltration column may provide an estimate of capillary conductivity under actual field conditions but further work must be done before any kind of conclusion can be deduced.

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