ESTIMATION OF EVAPOTRANSPIRATION FOR VARIABLE WATER TABLE SITUATIONS

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SUMMARY

Many investigators have studied rates of evaporation from soils where water table is at shallow depth. The rate of evaporation may be controlled by either the capacity of the atmospheric environment to evaporate water or the capacity of the soil to transmit water to the surface. A review note has been prepared which gives in detail the methods for estimation of evapotranspiration which considers both the soil properties and meteorological factors. The evaporation rate decreases with water table depth more steeply in coarse textured soil than in fine textured soils.

1.0 INTRODUCTION

Evapotranspiration(ET) is the conversion of water to vapour and transport of that vapour away from the watershed surface into the atmopshere. Water is available at plant surface, soil surface, streams and ponds. Evapotranspiration flux moves large quantities of water from the soil back to atmosphere. Solar radiation is the main energy source. The amount of liquid water and the energy to vaporize it will vary both in space and time over the watershed surface. In humid zones, about 750 to 950 mm/year of water may be vaporized. In subhumid areas, 550 to 700 mm/year commonly evaporate from vegetated surfaces. Leupold and Langbein (1960) estimated that 70 percent of the precipitation falling on the United States is returned to the atmosphere through ET.

When the surface is partly bare, evaporation can take place from the soil as well as from plants. In the absence of vegetation and when the soil surface is subject to radiation and wind effects, evaporation occurs directly and entirely from the soil. It is a process which if uncontrolled can involve very considerable loss of water in both irrigated and unirrigated agriculture. Upward flow from water tables subsequent to evaporation and transpiration from soils is a significoant phenomenon, particularly in irrigated areas. Evaporation of soil water involves not only loss of water but also the danger of soil salinization. The hydrologist, also, is concerned with upward flow from water tables since such flow can have a substantial effect on the fluctuation of ground water storage.

1.1 Physical Conditions necessary for Evaporation

The following three conditions are necessary if the evaporation process from a soil surface is to persist:

1. There must be a continual supply of heat to meet the latent heat requirement, which is about 590 Cal/gm of water evaporated at 15^{o}C.
2. The vapour pressure in the atmosphere over the evaporating body must remain lower than the vapour pressure at the surface of the body i.e.there must be a vapour pressure gradient between the body and the atmosphere and the vapour must be transported away by the process of diffusion or convection or both.
3. The third condition is that there should be a continual supply of water from or through the interior of the body to the site of evaporation. This condition depends upon the content and potential of water in the body as well as upon its conductive properties, which together determines the maximum rate at which the body can transmit to the evaporation site.

If the top layer of soil is initially wet, as it is at the end of a storm, the process of evaporation into the atmosphere will generally reduce soil wetness and thus increase the matric suction at the surface. This in turn, will generally cause soil water to be drawn upward from the layers below provided they are sufficiently moist.

1.2 Conditions under which evaporoation may occur

1. A shallow groundwater table may be present at a constant or variable depth. Where a ground water table occurs close to the surface, steady state flow may take place from the

saturated zone beneath, through the unsaturated layer to the surface. In the absence of shallow groundwater, the loss of water at the surface and the resualting upward flow of water in the profile will be a transient-state process causing the soil to dry,

2. The soil profile may be uniform(homogeneous and isotropic), or its properties may change gradually in various directions or the profile may consist of distinct layers.,
3. The profile may be shallow or deep,
4. The flow pattern may be one dimensional(vertical) or two or three dimensional(as in the presence of the vertical crakes which form secondary evaporation planes inside the profile, and
5. External environmental conditions may remain constant or fluctuate.

2.0 REVIEW

Evaporation and transpiration from irrigated soil is an important phenomenon which causes the water table to rise within a few feet of the root zone of crops(Anat et al.,1965). The upward flow from water tables can have a substantial effect on the fluctuation of ground water storage.

Many investigators have studied rates of evaporation from soils in which there is a water table. The rate of evaporation may be controlled by either the capacity of the atmospheric environment to evaporate water or the capacity of the soil to transmit water to the surface. The maximum rate of upward flow will be lesser of these two capacities.

Gardner(1958) suggested a convenient equation for describing hydraulic conductivity, the most relevant soil parameter and from it developed methods for evaluating soil-limited evaporation in cases of higher water table. Anat, Duke and Corey(1965) and Stallman(1967) employed Gardner's general approach but used different soil parametric equations. They demonstrated the usefulness of dimensionless curves in solving problems of the type under consideration. The above mentioned work stressed the cases in which soil properties were the determining factor of evaporation. Cases of evaporation in which the atmospheric conditions play the main role can be treated by means of several purely meteorological equations (Slayter and Mc Ilory, 1961). This review note gives in detail the work done in both the cases i.e,estimate of evaporation considering various soil properties and secondly the use of various meteorological equations for estimation of evapotranspiration.

2.1 General Theory

Water in soil moves from points where it has a high energy status to points where it has a lower one. The energy status of water is the water potential which consists of several components. In hydrology, potential is usually expressed as energy per unit weight of soil water which has the dimension of length (cm) and potential is denoted as 'head.' Considering the matric head, hm, arising from local interacting forces between soil and water and gravitational head, z, arising from the gravitational force, total (hydraulic) head, H, can be expressed as:

$$H = hm + z \qquad ...(1)$$

Where the vertical coordinate z is considered positive in upward direction.

For each soil there exists a relation between the pressure head, h(cm), and the soil water content, θ (cm^3, cm^{-3}), i.e.

$$\theta = f(h) \qquad ...(2)$$

The flow of water in soil systems can be described using Darcy law. For one dimensional vertical flow, the volumetric flux can be written as:

$$q = -K\frac{dH}{dz} \qquad ...(3)$$

Where K is the hydraulic conductivity.

For saturated(groundwater), flow, the total soil pore space is available for water flow and the hydraulic conductivity is constant. With unsaturated flow, however, part of the pores are filled with air. Therefore K is not a constant but depends on the soil moisture content θ or on the pressure head because [$\theta = f(h)$].

$$K = f(\theta) \text{ or } K = f(h) \qquad ...(4)$$

Substitution eq.(1) in eq.(3) yields

$$q = - K(h) \left(\frac{\delta h}{\delta z} + 1 \right) \qquad \ldots(5)$$

Complete mathematical description can be obtained by applying the continuity principle

$$\frac{\delta\theta}{\delta t} = \frac{\delta q}{\delta z} \qquad \ldots(6)$$

Substitution of eq.(5) in eq.(6) yields

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z} \left[K(h) \left[\frac{\delta h}{\delta z} + 1 \right) \right] \qquad \ldots(7)$$

Let $C = \frac{d\theta}{dh}$ = soil water capacity ...(8)

Writing $\frac{\delta\theta}{\delta t} = \frac{d\theta}{dh} \frac{\delta h}{\delta t}$ ...(9)

and substitution of eq. (9) in to eq.(7) yields the one dimensional equation for water flow in heterogeneous soils

$$\frac{\delta h}{\delta t} = \frac{1}{C(h)} \frac{\delta}{\delta z} \left[K(h) \cdot \left(\frac{\delta h}{\delta z} + 1 \right) \right] \qquad \ldots(10)$$

2.2 Capillary Rise from a Water Table

The rise of water in the soil from a free water surface i.e.water table has been termed as capillary rise. The equation relating the equilibrium height of capillary rise, hc, to the radius of the pores is

$$hc = \frac{2\gamma \cos\alpha}{r \rho_{\omega} g} \qquad \ldots(11)$$

where

γ = surface tension,

α = wetting angle,

r = capillary radius,

ω = water density, and

g = gravitional acceleration.

The rate of capillary rise generally decreases with time as the soil is wetted to greater height and as equilibrium is approached. The wetting of an initially dry soil by upward capillary flow as shown in Figure 1, is a rare occurrence in the field. In its initial stages, this process is similar to infiltration, except that it takes place in the opposite direction. At later stages of the process, the flux does not tend to a constant value, as in downward infiltration, but to zero. The reason is that the direction of the gravititional gradient is opposite to the direction of the matric suction gradient and when the matric suction approaches the magnitude of the gravitional gradient, the overall hydraulic gradient approaches zero. In general, the conditions of soil water is not static but dynamic, that is, soil water is constantly flowing. Where a water table is present, soil water generally does not attain equilibrium or rest even in the absence of vegetation, since the soil surface is subject to the evaporating action due to the atmospheric conditions. if the soil and external conditions are constant, that is, if the soil is of stable structure, the water table is at constant depth, and the atmospheric evaporativity also remains constant(at least approximately) then, in time, a steady state flow situation can develop from water to the atmosphere via the soil.

2.3 Steady Evaporation in the presence of a Water Table

The steady state upward flow of water from a water table through the soil profile to an evaporation zone at the soil surface was studied by Moore(1939). Moore was one of the first research workers to investigate upward flow from water tables as affected by the hydraulic properties of soils. He introduced water tables at the bottom of soil columns and allows the soil to imbibe water. The surface was subjected to eva-

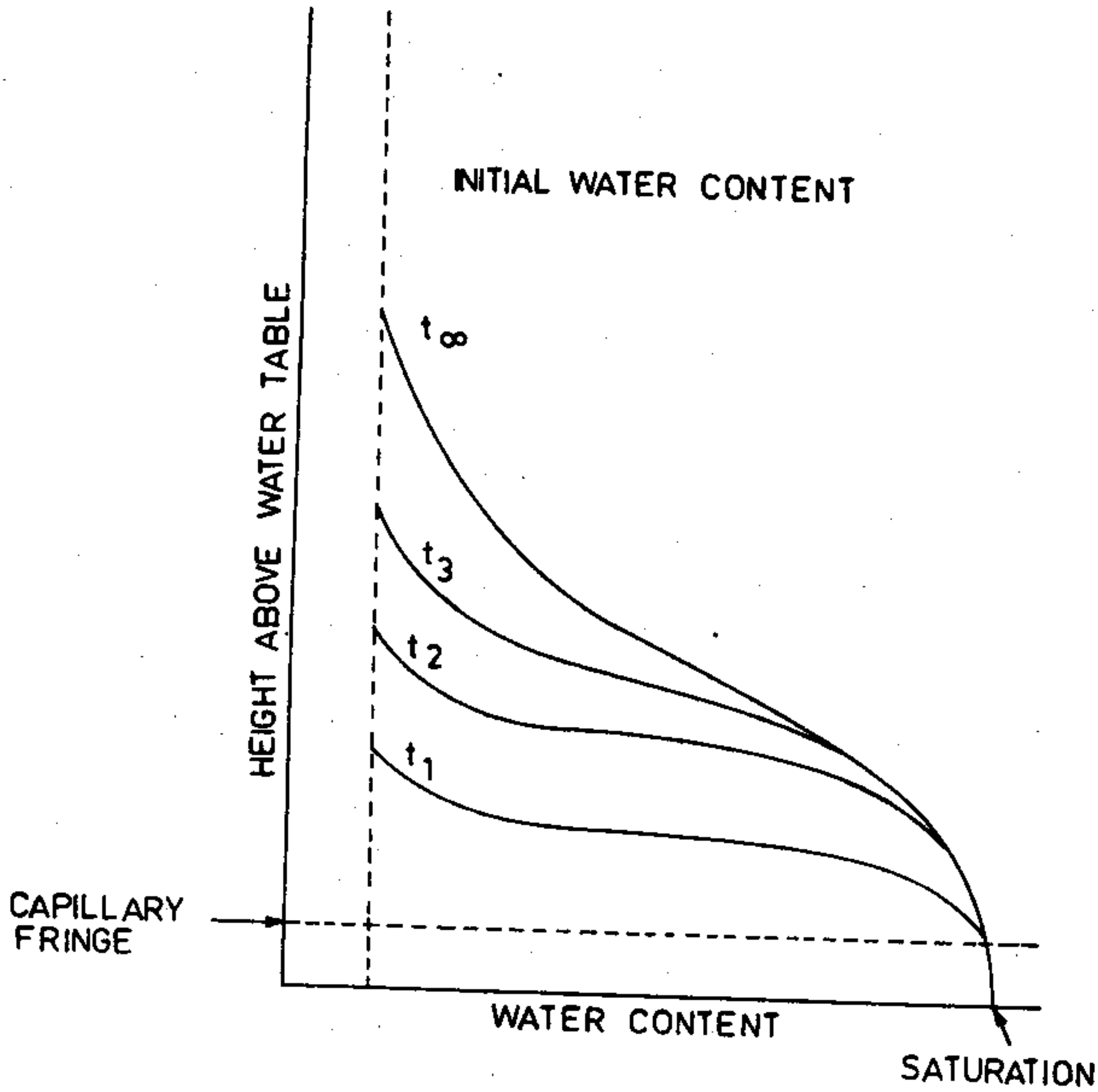

Figure 1- The upward infiltration of water from a water table into a dry soil

poration and, by employing tensiometer cups placed at intervals along the length of the column, he observed the relationship between water pressure, moisture content, and rate of water loss from the water table.

From his studies, with materials of widely varying texture, Moore concluded that at complete saturation,permeabilities of soils were arranged in the following order

Sand > fine sandy loam > light clay > clay

When the capillary potential was decreased to -100 ergs per gram, he found the order to be

Sand < fine sandy loam < light clay < clay

Moore(1939) concluded that when the water table is at greater depth the finer soil supported high evaporation rate. Theoretical solutions of the flow equation of this process has been given by several workers (Philip (1957 d), Gardner(1958), Anat et al.(1965) and Ripple et al.(1972).

The equation describing upward flow is

$$q = K(\Psi)(d\Psi | dz - 1) \quad \ldots(12)$$

or

$$q = D(\theta)\frac{d\theta}{dz} - K(\Psi) \quad \ldots(13)$$

where,

q = flux, equal to the evaporation rate under steady state conditions.

Ψ= suction head,

K = hydraulic conductivity,

D = hydraulic diffusivity,

θ= volumetric wetness, and

z = height above the water table.

The equation shows that flow stops (q=0) when $d\psi/dz=1$. The equation(12) can be written as

$$\frac{q}{K(\psi)} + 1 = d\psi | dz \quad \ldots(14)$$

The integration of equation(14) will give the relation between depth and suction or wetness i.e.

$$z = \int \frac{K(\Psi)}{K(\Psi) + q} \, d\Psi \qquad \ldots(15)$$

or

$$z = \int \frac{D(\theta)}{K(\theta) + q} \, d\theta \qquad \ldots(16)$$

In order to perform the integration in equation(15), the functional relationship between K and Ψ [i.e. $K(\Psi)$] must be known. An empirical equation for $K(\Psi)$ given by Gardner(1958) is

$$K(\Psi) = a(\Psi^n + b)^{-1} \qquad \ldots(17)$$

Where the parameter a,b, and n are constants which must be determined for each soil. Accordingly equation(12) becomes

$$e = q = \frac{a}{\Psi^n + b} \left(\frac{d\Psi}{dz} - 1\right) \qquad \ldots(18)$$

Where e is the evaporation rate.

Equations(15) and (17) can be integrated to obtain suction distributions with height for different fluxes as well as fluxes for different surface suction values. The theoretical solution is shown by Gardner(1958) in Figure 2 for a fine sandy loam soil with an n value of 3.

The curves show that the steady rate of capillary rise and evaporation depend on the depth of the water table and on the suction at the soil surface. The maximal transmitting ability of the profile depends on the hydraulic conductivity of the soil in relation to the suction.

Gardner (1958) obtained the following function for q_{max}

$$q_{max} = Aad^{-n} \qquad \ldots(19)$$

where

d= depth of water table below the soil surface,

a,n = constant from equation(17),

A = constant which depends on n,

q_{max} = maximum rate at which the soil can transmit the water from the water table to the evaporation zone at the surface.

Where the water table is near the surface, the suction at the soil surface is low and the evaporation rate is determined by external conditions. However, as the water table becomes deeper and the suction at the soil surface increases, the evaporation rate approaches a limiting value regardless of how external evaporativity may be. Equation(19) derived by Gardner suggests that the maximal evaporation rate decreases with water table depth, more steeply in coarse-textured soil than in fine-textured soils. Sandy loam soil can evaporate water even when the water table is as deep as 180 cm. Figure 3 shows the effect of texture on the limiting evaporation rate.

Anat et al.(1965) developed a modified set of equations employing dimensionless variables. Their theory also leads to a maximal evaporation rate e_{max} varying inversely with water table depth d to the power of n.

$$e_{max} = [\ 1 + 1.886/(n^2 + 1)\ ]\ d^{-n} \qquad \ldots(20)$$

A theoretical analysis of steady evaporation from a two-layered soil profile was carried out by Willis(1960). The following assumptions were taken into consideration:

(a) the steady flow through the layered profile is governed only by the transmission properties of the profile,

(b) matric suction is continuous at and through the interlayer boundary, though wetness and conductivity may be discontinuous i.e. change abruptly,

(c) the empirical function $K(\Psi)$ given by equation(17) holds for the both layers, but the values of parameters a,b, and n are

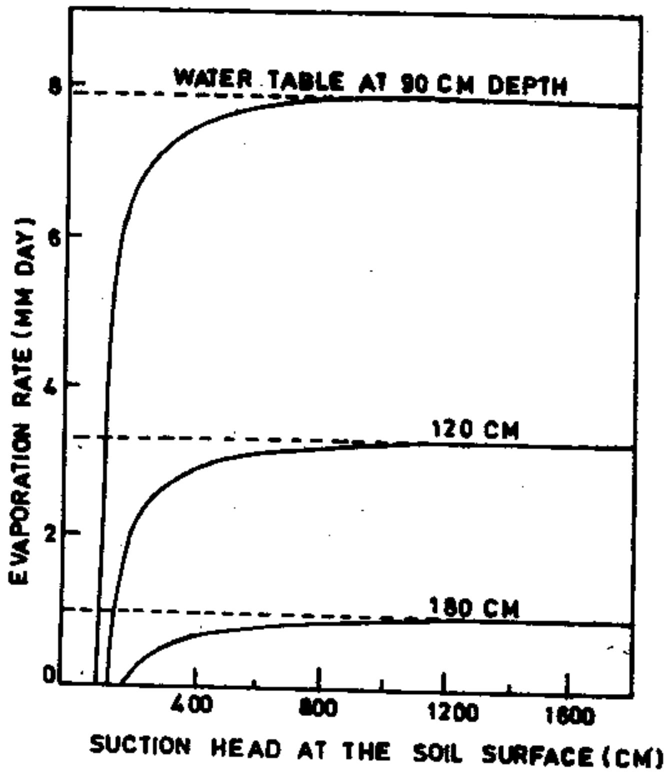

Fig. 2 Steady rate of upward flow and evaporation from a water table as function of the suction prevailing at the soil surface.

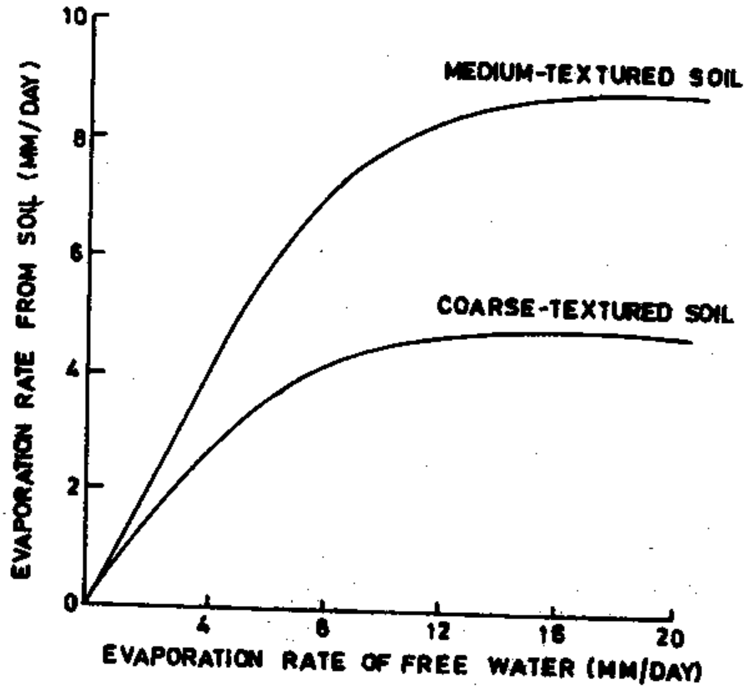

Fig.3 - Theoretical relation between the rate of evaporation from coarse and medium textured soil and the rate of evaporation from free water surface.

different, and

(d) each soil layer is internally homogenous.

Willis (1960) derived the following relationship, with the above assumptions, for a two layered soil profile.

$$\int_{o}^{L} dz + \int_{L}^{d+L} dz = \int_{\Psi_O}^{\Psi L} \frac{d\ \Psi}{1+e/K_1(\Psi)} + \int_{\Psi_L}^{\Psi L+d} \frac{d\ \Psi}{1+e/K_2(\Psi)} \quad ...(21)$$

Where L and d are the thickness of the bottom and top layers, respectively. The integral in this equation relates water-table depth L+ d to the suction at the soil surface for any given evaporation rate. By assuming that the suction at the soil surface is infinite, the maximal evaporation rate for any given water-table depth and profile layering sequence can be estimated. Willis(1960) developed a graphical method for obtaining the necessary solution.

All of the above treatments apply to cases in which soil properties are the sole factor determining the evaporoation rate. A more realistic approaoch should include cases in which meteorological conditions can also play a role. A more flexible treatment of steady-state evaporation from multilayer profiles should be based on numerical, rather than analytical or graphical, methods of solution.

2.4 Estimation of Evaporation Using Meteorological and Soil Parameters

Ripple et al.(1972) developed a procedure that combines meteorological and soil equations of water transfer for estimation of steady state evaporation from bare soils under conditions of high water table. This procedure takes into account the various atmospheric factors like air temperature, air humidity and wind velocity and the soil factors like soil water retention curves, water table depth which are responsible to convert water into liquid and vapour forms. The evaporation rate can be estimated for homogeneous soils as well as for layered soils. The results of their work are shown in Figure 4 and 5 respectively.

2.4.1 Theory

The Steady-State evaporative fluxes across the boundary between any given soil atmosphere system may be described by two functional relations. The first deals with the fluxes leaving the soil surface and entering the atmosphere. It can be represented by the meteorological equation.

$$S_u = F_m (E) \qquad ...(21)$$

The second describes the fluxes between the water table and the soil surface and may be expressed by the soil equation

$$L = F_g (S_u, E) \qquad ...(22)$$

where,

L = total distance between the water table and the soil surface,

S_u = water suction at the soil surface defined as the negative of the soil water pressure head,

E = rate of evaporation from the soil cm/day,

F_g = a function which relates E and S_u using soil parameters, and

F_m = a function which relates E and S_u using meteorological parameters.

2.4.2 Meteorological equation

The basic meteorological equation used for estimation of evaporation is of the type generally known as the bulk aerodynamic, or Dalton, (Slatyer and McIlory,1961).

$$E = G(V_a)[p(T_u)h_u - p(T_a)h_a] \qquad ...(23)$$

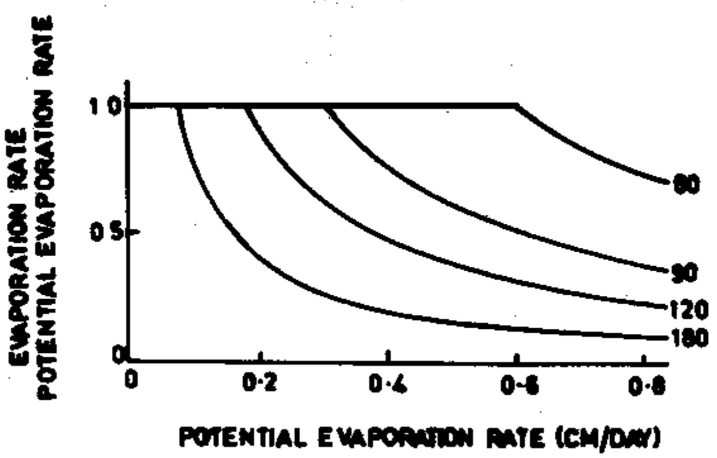

Fig.4- Dependence of relative evaporation rate upon the potential evaporation rate for a clay soil.

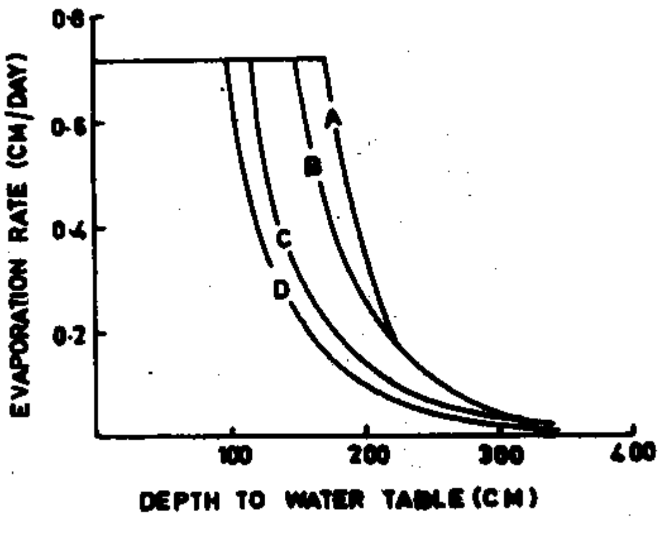

Fig.5 - Influence of layering on the relation between evaporation rate and depth to water table.

where,

$G(V_a)$ = a theoretically or empirically derived known function of wind speed, cm/day/mb,

V_a = wind speed at height H_a, cm/day,

h = relative humidity, dimensionless,

T = temperature K

p = saturation vapor pressure of water,mb,

p(T) = a known relation between the saturation water vapour pressure and temperature,

a = a subscript indicating a variable determined in the air, H_a cm above the soil surface and

u = a subscript indicating a variable determined at the soil surface.

The above equation has been used extensively for estiomating the loss of water by free water surfaces, plants and bare soils. The wind function $G(V_a)$ can be determined using the equation of Van Bavel(1966),

$$G(V_a) = \frac{\rho_a \varepsilon\ k^2}{\rho_\omega\ P} \frac{V_a}{(\log_e H_a/H_u)^2} \quad ...(24)$$

where,

ρ_a = air density at T_a, gm cm^{-3},

ρ_ω = Water density at T_a, gm cm^{-3},

ε = water air molecular ratio = 0.622, dimensionless,

k = von karman constant = 0.41, dimensionless,

P = ambient pressure, mb(P=1000 mb as suggested by Ripple et al.),

H_a = height of meteorological measurements, above the soil surface, cm, and

H_u = roughness parameter,cm(usually for bare soils, $0.01 < H_u < 0.03$)

The surface relative humidity, h_u, can be obtained from equation(23)

$$h_u = \frac{1}{p(T_u)} \left[\frac{E}{G(V_a)} + p\,(T_a)\,h_a \right] \qquad ...(25)$$

The surface relative humidity, h_u, specified by equation(25) can be substituted into the thermodynamic relation (Edlefson and Anderson, 1943)

$$S_u = -\frac{RT_u}{M.g}\log_e h_u \qquad ...(26)$$

where,

M = molecular weight of water (18gm $mole^{-1}$)

g = acceleration of gravity (981cm sec^{-2})

R = gas constant (8.32 x 10^7 ergs 0K^{-1} $mole^{-1}$

The above substitution would result in an equation expressing S_u in terms of atmospheric variables, the soil surface temperature T_u, and E.

To completely attain the form of equation(21), the variable on the right hand side of the equation should be, except for E, entirely meteorological. The term surface soil temperature, T_u, is present in the combination of equations(25) and (26). To replace T_u, with meteorological variables and parameters, an appropriate expression for T_u, may be developed. The T_u term is related to sensible heat transfer in the air by the following equation for turbulent transfer (Slayter and McIlory, 1961); Van Bavel, 1966)

$$A = -\lambda\gamma\, G(V_a)\,(T_u - T_a) \qquad ...(27)$$

where,

A = sensible heat transfer into the air,

λ = latent heat of vaporization of water at T_a,

γ = psychrometric constant.

Secondly, substitute A term into the following heat-balance equation (Slayter and McIlory, 1961; Van Bavel 1966)

$$Q_N = \lambda\rho_\omega E + \rho_\omega A + Q_g \qquad \ldots (28)$$

where,

Q_N = net radiation flux received by the soil surface, and

Q_g = soil heat flux into the ground.

The combined equations(27) and (28) after rearrangement yield the following for T_u:

$$T_u = T_a + \frac{Q_N - Q_g - \lambda\rho_\omega E}{\lambda\gamma\rho_\omega e \quad G(V_a)} \qquad \ldots(29)$$

Substituting the value of T_u into a combination of equations (25) and (26) will give the overall meteorological equation, equivalent to equation(21).

2.4.3 Soil Equations

The simplest system for a single layer and multilayers for estimation of evaporation using soil parameters is shown in Fig.6(a). Consider a homogeneous soil, underlain by a shallow water table, with the reference height Z measured positively upward from the piezometric surface. The soil is at Z=L.

Three cases have been considered for estimating evaporation under different soil layers and water table conditions,(Ripple et al,1972)

Case A - A homogeneous soil in which water is transferred exclusively in liquid form,

Case B - A layered soil in which water is transferred exclusively in liquid form, and

Case C - A homogeneous soil in which water is transferred in liquid and vapor forms.

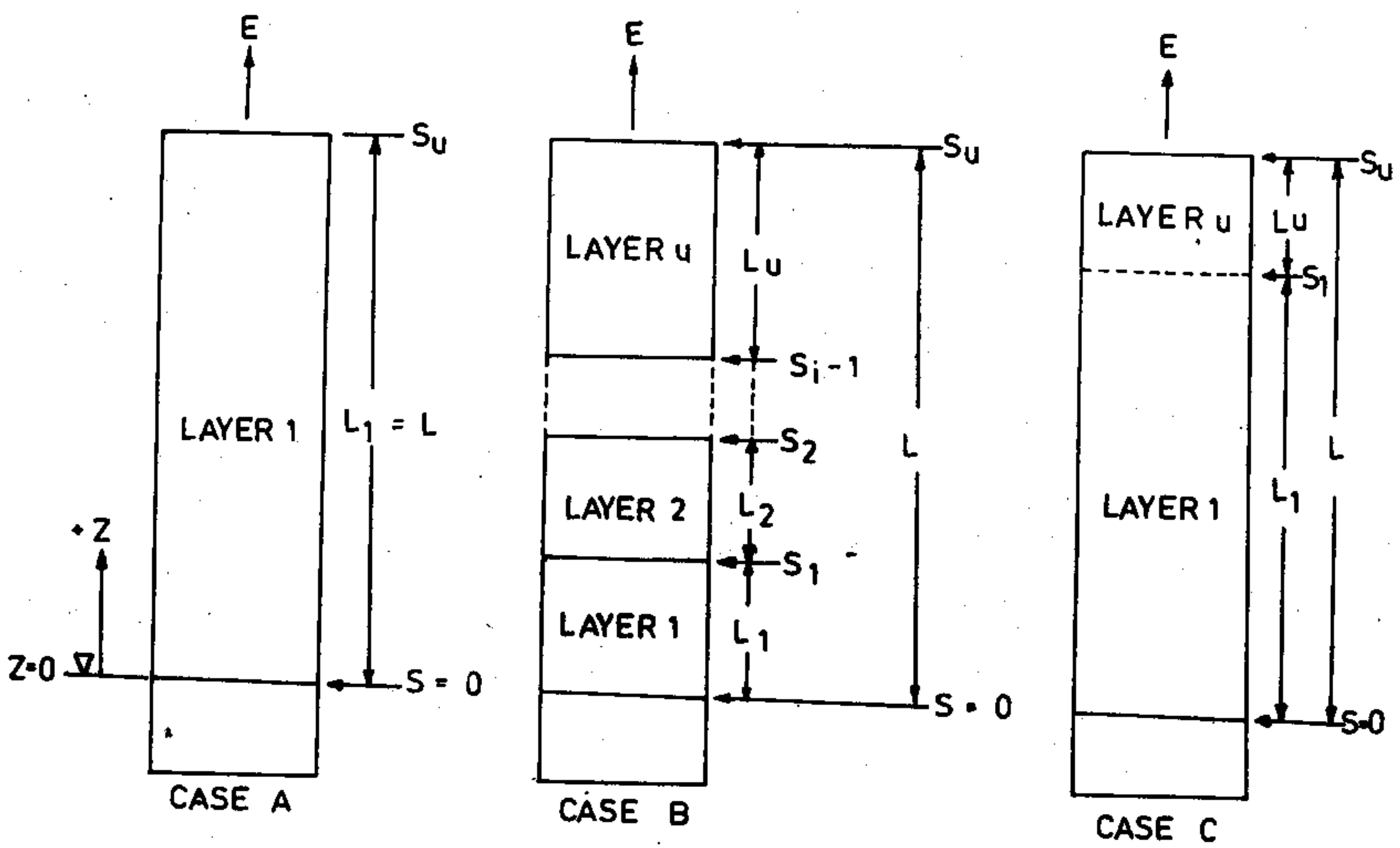

Fig.6 - Water table-soil-atmospheric system for a) homogeneous soil in which water is transferred in liquid form b) a layered soil in which water is transferrred exclusively in liquid form and c) a homogeneous soil in which water is transferred in liquid and vapour forms.

Gardner(1958) suggested an empiricial equation for estimating hydraulic conductivity for water transfer in liquid form. In 1964, Gardner presented a modified form which demonstrates more clearly the physical significance of the coefficients:

$$K = K(S) = \frac{K_{Sat}}{(\frac{S}{S_{1/2}})^{n} + 1} \qquad \ldots(30)$$

where,

K = hydraulic conductivity for liquid flow, cm day^{-1},

K_{sat} = hydraulic conductivity of water saturated soil,cm day^{-1},

S = soil water suction, defined as the negative of the soil water pressure head, cm of water,

$S_{1/2}$ = a constant coefficient representing S at $K = 1/2\ K_{sat}$ cm of water, and n= an integer soil coefficient.

Assuming that Darcy's equation holds for flow in both saturated and unsaturated soils, the flux q, which under steady state conditions must equal the evaporation rate E, may be described by

$$q = E = K\ (\frac{dS}{dZ} - 1) \qquad \ldots(31)$$

On rearranging and integrating, equation(31) becomes

$$Z' = \int_{o}^{Z'} dZ = \int_{o}^{S'} \frac{dS}{\frac{E}{K(S)} + 1} \qquad \ldots(32)$$

where,

$$S' = S \text{ at } Z = Z' < L$$

Substituting the value of K(S) from equation (30) into equation (32) will yield

$$Z' = \int_{o}^{S'} \frac{dS}{\frac{E}{K_{sat}}[(\frac{S}{S_{1/2}})^{n}+1] + 1} \qquad \ldots(33)$$

Equation (33) expresses explicity Z' as a function of S and E. It also defines implicitly the relation between E and S' for any given Z'. Both facts have been utilized in the past(Philip,1957a; Gardner 1958). Equation(33) can be converted into another form by using the following transformation, Assuming

$$e = E/K_{sat} \qquad ...(34)$$

$$Z' = \frac{1}{e} \frac{1}{(1+\frac{1}{e})} \int_o^{S'} \frac{dS}{\left[\frac{S/S_{1/2}}{(1+\frac{1}{e})^{\frac{1}{n}}} \right]^n + 1} \qquad ...(35)$$

with the substitution

$$Y = \frac{S}{S_{1/2}} \left(\frac{e}{1+e} \right)^{\frac{1}{n}} \qquad ...(36)$$

into equation (35) and integration lead to

$$(e+1) \left(\frac{e}{e+1}\right)^{\frac{1}{n}} \frac{Z'}{S_{1/2}} = \int_o^{y'} \frac{dy}{y^n + 1} \qquad ...(37)$$

where,

$$y' = \frac{S'}{S_{1/2}} \left(\frac{e}{e+1} \right)^{\frac{1}{n}}$$

At the soil surface, when $Z' = L$

$$(e+1) \left(\frac{e}{e+1} \right)^{\frac{1}{n}} \frac{L}{S_{1/2}} = \int_o^{y_u} \frac{dy}{y^n + 1} \qquad ...(38)$$

where,

$$y_u = \frac{S_u}{S_{1/2}} \left(\frac{e}{e+1} \right)^{\frac{1}{n}}$$

The integral on the right hand side of equations(37) and (38) is known in closed form for any positive n(Gradshteyn and Ryxhik, 1965). The form of equations(37) or (38) makes it possible to determine the relation between e and the suction(either S' or S_u) for any n by means of simple graph.

The use of equations(37) and (38) can be further simplified by adopting the dimensionless variables

$$s = \frac{S}{S_{1/2}} \qquad ...(39)$$

$$z = \frac{Z}{S_{1/2}} \qquad ...(40)$$

and

$$l = \frac{L}{S_{1/2}} \qquad ...(41)$$

in addition to the dimensionless $e = E/K_{Sat}$. With the exception of s, these dimensionless variables are similar to those employed by Staley (cited by Anat and others, 1965), whose hydraulic conductivity equation is also somewhat similar to equation(30).

The above dimensionless variables reduces the basic equation(38) to

$$(e+1)\left(\frac{e}{e+1}\right)^{\frac{1}{n}} 1 = \int_0^{y_u} \frac{dy}{y^n + 1} \qquad \ldots(42)$$

where,

$$y_u = S_u \left(\frac{e}{e+1}\right)^{\frac{1}{n}}$$

It is clear from physical considerations that an increase in the evaporative capacity of the atmosphere will produce an increased suction at the soil surface. This higher suction, in turn, must magnify the upward water flux through the soil. If equation(38) correctly describes reality, such a flux cannot increase without bound, because as S_u(and hence y_u) approaches infinite the integral on the right hand side of equation (38) approaches a finite limit,. $\pi/(n \sin \pi/n)$ as described by (Gradshteyn and Ryzhik,1965). It follows that a limiting soil-water-flux and hence a soil limited evaporation; e_∞ , exists.

For any particular soil system, a soil limited evaporation e_∞ is given by

$$(e_\infty + 1)\left(\frac{e_\infty}{e_\infty + 1}\right)^{\frac{1}{n}} \frac{L}{S_{1/2}} = \frac{\pi}{n \operatorname{Sin} \frac{\pi}{n}} \qquad \ldots(43)$$

or in completely dimensionless form

$$(e_\infty + 1)\left(\frac{e_\infty}{e_\infty + 1}\right)^{\frac{1}{n}} 1 = \frac{\pi}{n \operatorname{Sin} \frac{\pi}{n}} \qquad \ldots(44)$$

The equations (43) and (44) can be simplified considerably if $e_\infty << 1$ (that is if $E << K_{Sat}$). In such a case, $e_\infty + 1 \doteq 1$, and above two equations lead to

$$E_\infty = K_{Sat} \left[\frac{S_{1/2}}{L} \right]^n \left[\frac{\pi}{n \sin\frac{\pi}{n}} \right]^n \qquad \ldots(45)$$

and

$$e_\infty = \frac{1}{1^n} \left[\frac{\pi}{n \sin\frac{\pi}{n}} \right]^n \qquad \ldots(46)$$

Equation(45) is similar to the formulas for E_{Lim} given without derivation by Gardner(1958) for $n = \frac{3}{2}$, 2,3,4 and yields identical numerical coefficients.

2.5 Investigations at Colorado State University

Staley(1957) conducted an experiment to determine(1) how the evaporation rate from a fine sand is affected by independently varying the wind velocity and the depth of water table, and (2) how the functional relationship between evaporation rates under specified ambient conditions and the depth of the water table can be related to easily measurable properties of the sand.

To determine this, Staley(1957) conducted an experiment on wind tunnel. Two columns of a fine sand were placed side by side in a test secton of a wind tunnel. Each column was 46 inches in length and 12 inches in a diameter. One column, in which the water table was maintained at the surface, was used to evaluate the evaporativity while the water table depth in the second column was varied.

Both columns were initially fully saturated before the water table was lowered in one of the columns. The columns were supplied by water from a constant head source such that the rate of intake could be measured. Runs were made with the wind velocity varying from 0 to 50 feet per second while the water table was varied independently between the surface and a depth of 30 inches. The runs were conducted so that the column with the variable water table was always on the drainage cycle. Each run was continued until steady state flow was approached.

A critical depth of water table for the fine sand was found at about 24 inches. With the water table above this depth, evaporation rate were of the same order of magnitude as from the free water surface. When the water table was below 24 inches, the flow rate decreased rapidly with depth and at 30 inches the rate was too small to be measured accurately with the apparatus used for this study.

Staley(1957) also related the critical water table depth with capillary pressure-desaturation curve. He found that the critical depth corresponded to the value of $P_c/\rho g$ at which the initially saturated sand began to desaturate rapidly with further increase in capillary pressure, P_c. This value of P_c has been called bubbling pressure P_b by Brooks and Corey(1964).

In analysing the problem of upward flow from water tables, Staley (1957) expressed the flow rate, the capillary pressure and the elevation above the water table in terms of scaled variables as follows:

- the ratio of the flow rate to the hydraulic conductivity, q/c
- the ratio of capillary pressure to the bubbling pressure P_c/P_b
- the ratio of elevation above the water table to the bubbling pressure head $\frac{z}{Pb/\rho g}$

Staley (1957) also derived an equation for the steady upward flow rate q but used an approach somewhat different from Gardner(1958). Staley wrote Darcy's law for the case of one dimensional flow in the vertical direction and re-arranged this equation to give an expression describing the rate of change of capillary pressure, Pc, with z as a function of volume flux q, and the effective conductivity Ce, i.e.

$$\frac{d\,[\frac{Pc}{\rho g}]}{dz} = 1 + \frac{q}{C_e} \qquad ...(47)$$

Solution of above equation for z, yields.

$$z = \int_o^{P_c/\rho g} \frac{d(Pc/\rho g)}{1 + q/Ce} \qquad ...(48)$$

He assumed that C_e is a constant for $P_c < P_b$, and $P_c > P_b$, C_e is given by

$$C_e = C\,[\,P_b/P_c]^n \qquad ...(49)$$

Where C is the value of C_e when media are fully saturated and n is constant for particular type of soil.

Staley(1957) computed the value of n for sand using a method proposed by Burdine(1953) and arrived at a value of 16. Gardner's(1958) observation was that n would not be larger than about 4. After reviewing the literature, Staley concluded that a value of n equal to 8 should be about average for sands and that the larger value of n for his sand was due to its extremely uniform pore size.

Using n=8, Staley solved equatioN(48) by a numerical method for a particular water-table depth to obtained a relationship between P_c/P_b. (at the surface) and q/C. The water table depth assumed was such as to give a value of $z/P_b/\rho g)$ at the surface of 2.5.

Because the experiments of Staley were run in a wind tunnel which was not equipped to supply radient energy at the column surfaces, it was impossible for him to produce evaporativities sufficiently high to cause the evaporation rates to reach limiting values for most of the water table depths studied.

Schleusener(1958) conducted experiment in a chamber with controlled temperature, humidity and radiation such that evaporativity could be raised to much higher levels. He found that when evaporativity was increased beyond a critical level, the flow rates decreased.

Schleusener and Corey(1959) tried to discover the failure of Schleusener(1958) theory. They found that it may be due to severe temperature gradient at the surface was heated by incident radiation. They tried, therefore, to produce a decrease in upward flow from wet soil to drier soil by artificially producing an enormous temperature gradient. They finally concluded that the reduction in flow rates at high evaporativity was due to a reversal in the direction of pressure change at a point below the evaporating surface. They reasoned that a high evaporativity initially moves water out of the surface layer faster than it can be supplied from below. They also presented a capillary model to describe in detail how this hysteresis takes place.

King and Schleusener(1961) showed that cyclic conditional of radiation, temperature and humidity(simulating the daily fluctuations of these variables in the field) resulted in the same trends as were observed with steady ambient conditions. In particular, the reduction of flow rates at high levels of evaporativity(depending on the water table depth) were still observed.

Staley(1957), Schleusener(1958) and King and Schleusener(1961) measured the temperature gradients existing in soil profiles during evaporation. They observed, however, that large temperature gradients were confined to the upper surface layer which usually becomes dry. They concluded that the temperature gradient could have little effect on the liquid flow rates upto the dried layers and that the rate of liquid flow upto the dried layer usually controlled the evaporation rate.

Duke(1965) attempted to determine how the steady flow rates from water tables can be predicted for any soil with any water table depth. He analysed his data in terms of dimensionless variables introduced by Staley i.e. q/c, Pc/Pb, z/(Pb/ρg), and n He designated these as q,P,z and n respectively.

Duke (1965) adopted the empirical equation for the functional relationship $C_e(P)$ introduced (1956). He presented equation in the form

$$d = \frac{1}{1+q_m} + \int_1^{\infty} \frac{dp}{1+ q_m p^n} \qquad \text{...(50)}$$

Where d is the elevation from the water table to the bottom of layer of dried surface soil, which is the particular value of z at elevation of the bottom of the dried soil layer. The subscript m implies that q_m is the maximum value of q occuring when $P \to \infty$ at the scaled elevation d.

Duke(1965) devised a graphical solution for equation(50) involving the determination of the area under the curves as shown in Figure 7. He also developed a computer program which determines the integration in equation(50) for integral values of n from 2 through 20. This range of n includes all values observed for porous media to date. The solutions were compiled into a nomograph relating q_m to d for each value of n The complete nomograph is presented in Figure 8.

Duke also devised a method of measuring flow rates for columns in which the capillary pressure was maintained at a zero value at a fixed elevation near the bottom of the column. Rather than depending on evaporation from the surface to produce upward flow, the liquid was removed from the surface through a capillary barrier attached to a siphon Flow rates were determined by measuring the outflow from the siphon after

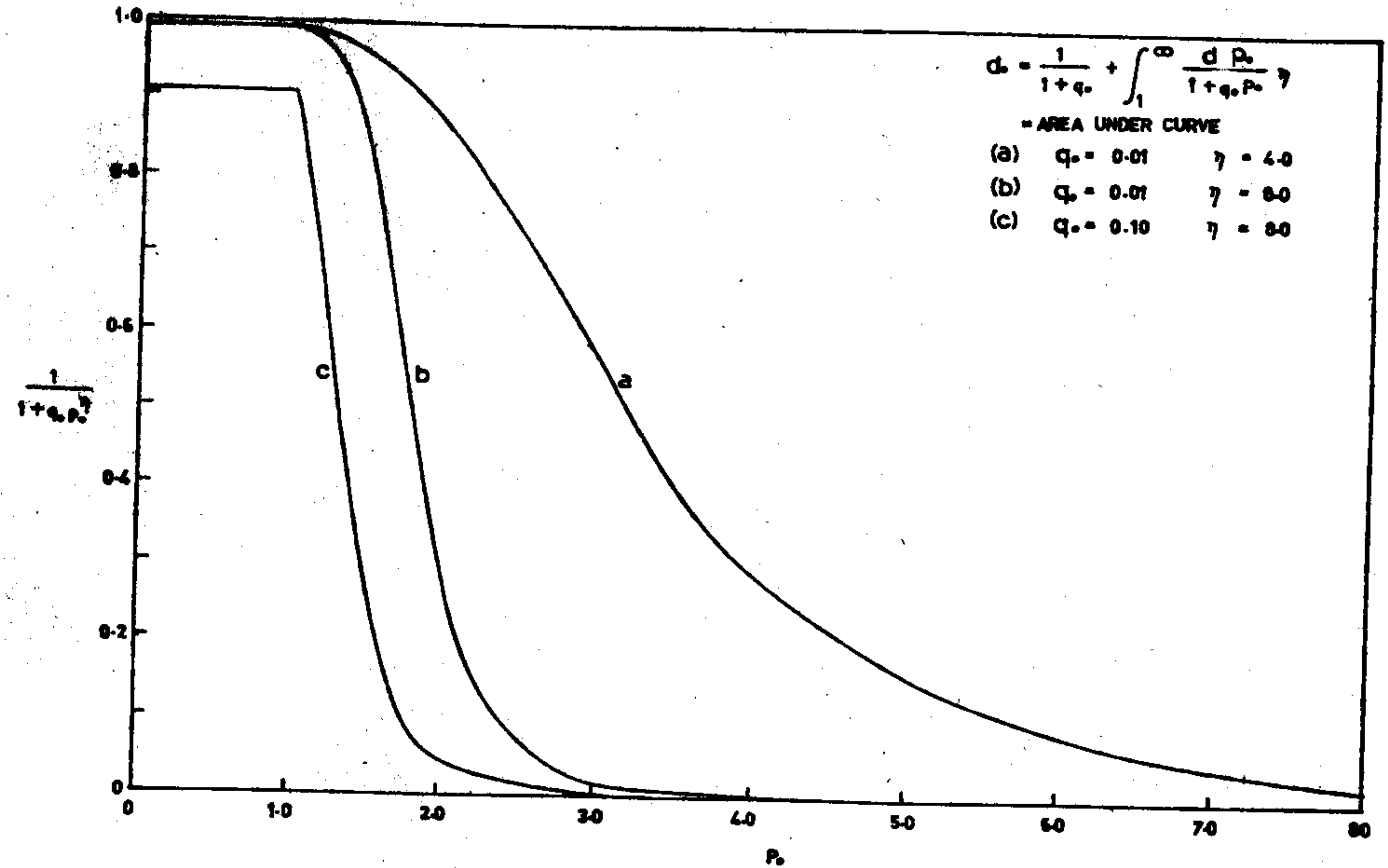

Fig.7 - Plot for the graphical solution of equation (50).

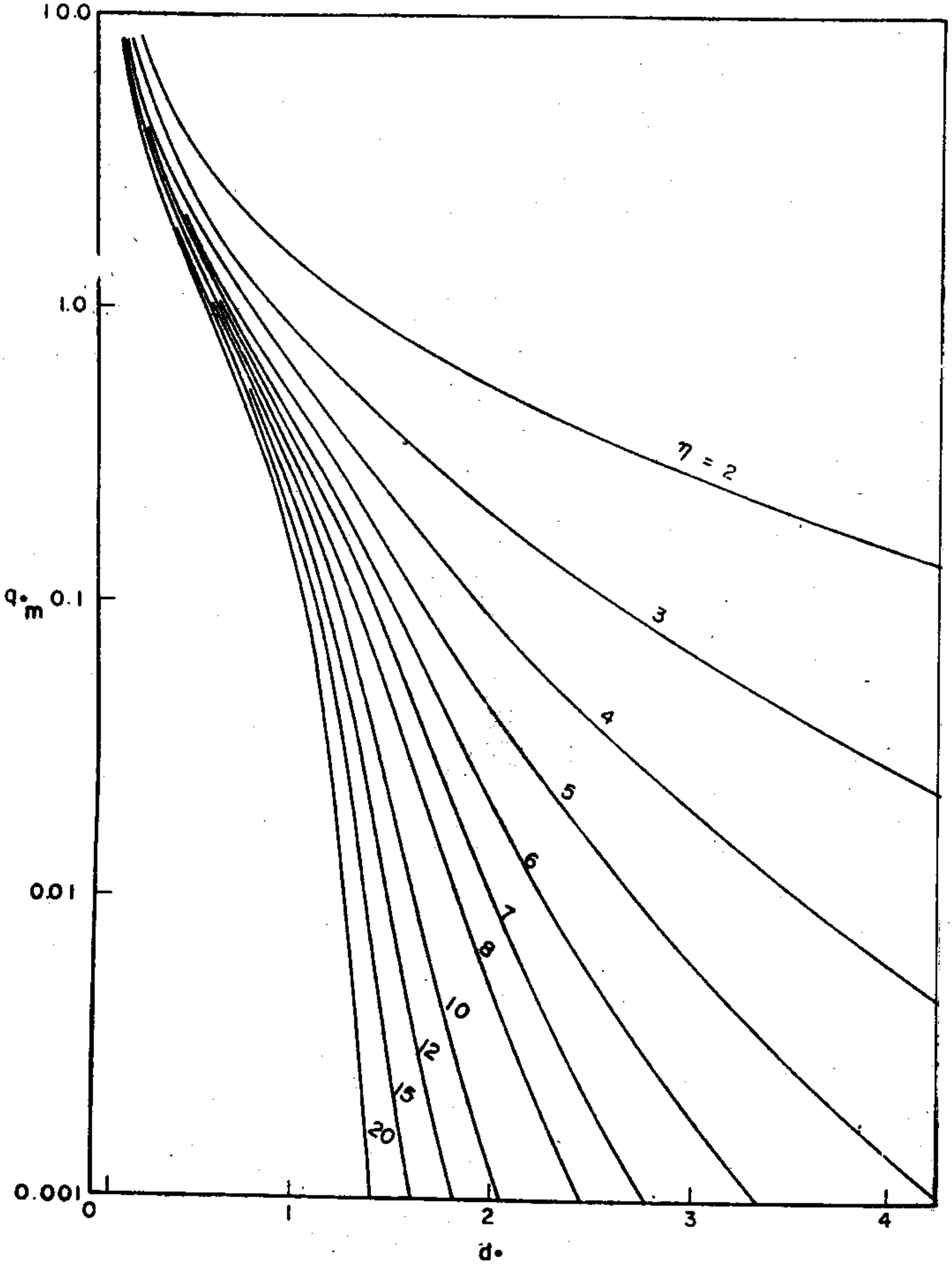

Fig.8- Nomograph of relation among d, q_m, and n.

the outflow rates became steady with a particular capillary pressure on the top barrier. A temperature control system held the temperature within reasonable limits.

Duke (1965) obtained the maximum flow rates by lowering the outflow siphon in increments until the flow rates appeared to approach a limiting value. The water table depth was varied by cutting the column after the maximum rate had been established for one particular length of column. A wide variety of soil types were used ranging in texture from heavy clay to sands.

Anat(1965) conducted an experiment at Colorado State University of determine more precisely the effect of hysteresis on upward flow from water tables. He has used the same experimental setup used by Duke but his procedures, in most cases, differed significantly from those followed by Duke.

Anat(1965) packed the columns by first filling them with soil and then vibrating them with an electric powered vibrating device until thev reached a relatively stable degree of compaction. Curves of relative conductivity as a function of capillary pressure were obtained which could be represented very well by equation(50). Both n and P_b were determined with precision as shown in Figures 9 and 10.

Anat(1965) observed that in some cases, curves of relative conductivity as a function of capillary pressure were also obtained on the imbibition cycle. This was accomplished by first draining the columns to a very low saturation(high P_c) and then raising the inflow reservoir and outflow siphon in increments. The conductivity was determined after steady state was obtained for each increments.

From these measurements, he observed that the conductivity at particular values of capillary pressure on the imbibition cycle is one

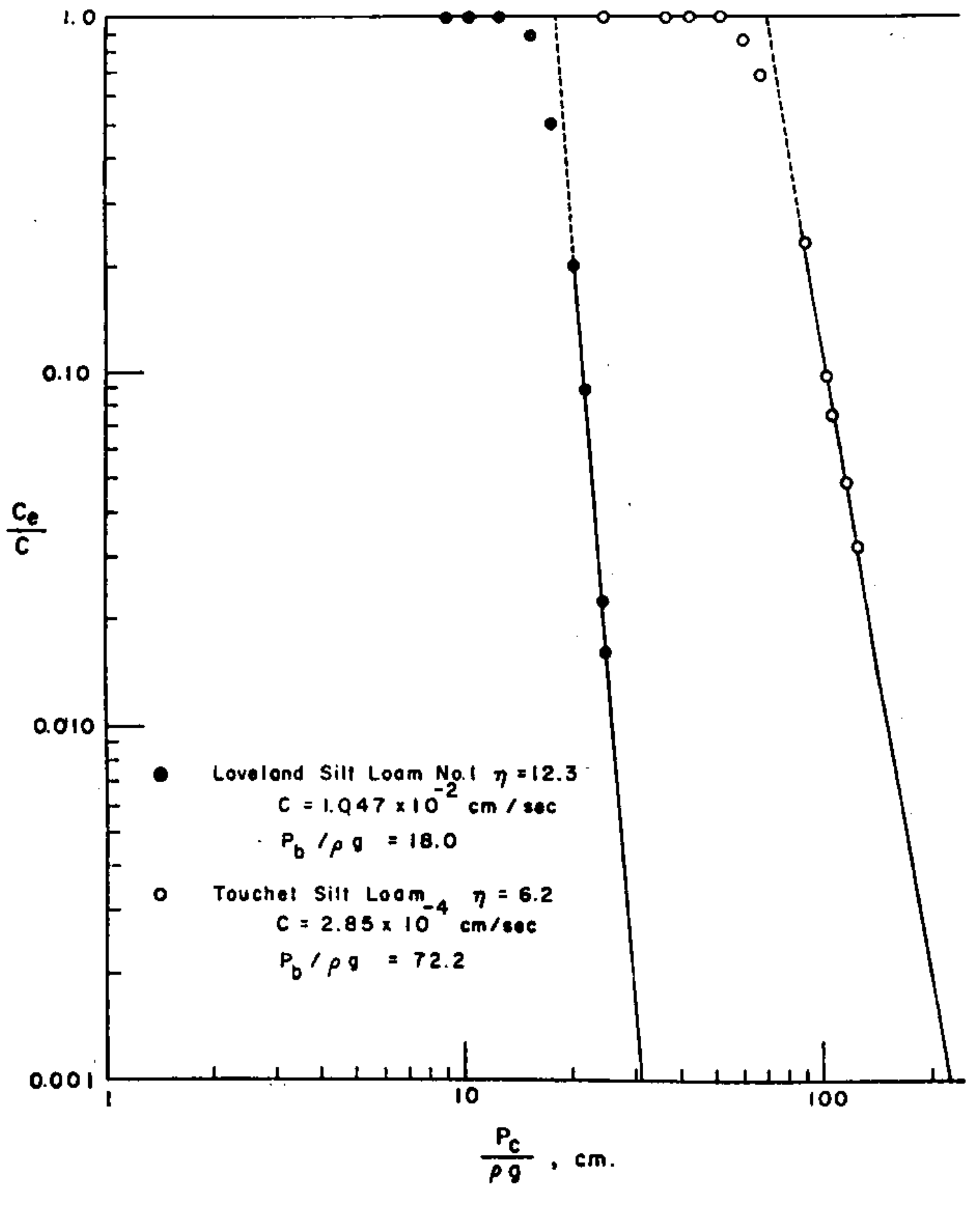

Fig.9 - Relative Conductivity-capillary pressure curves for drainage cycle

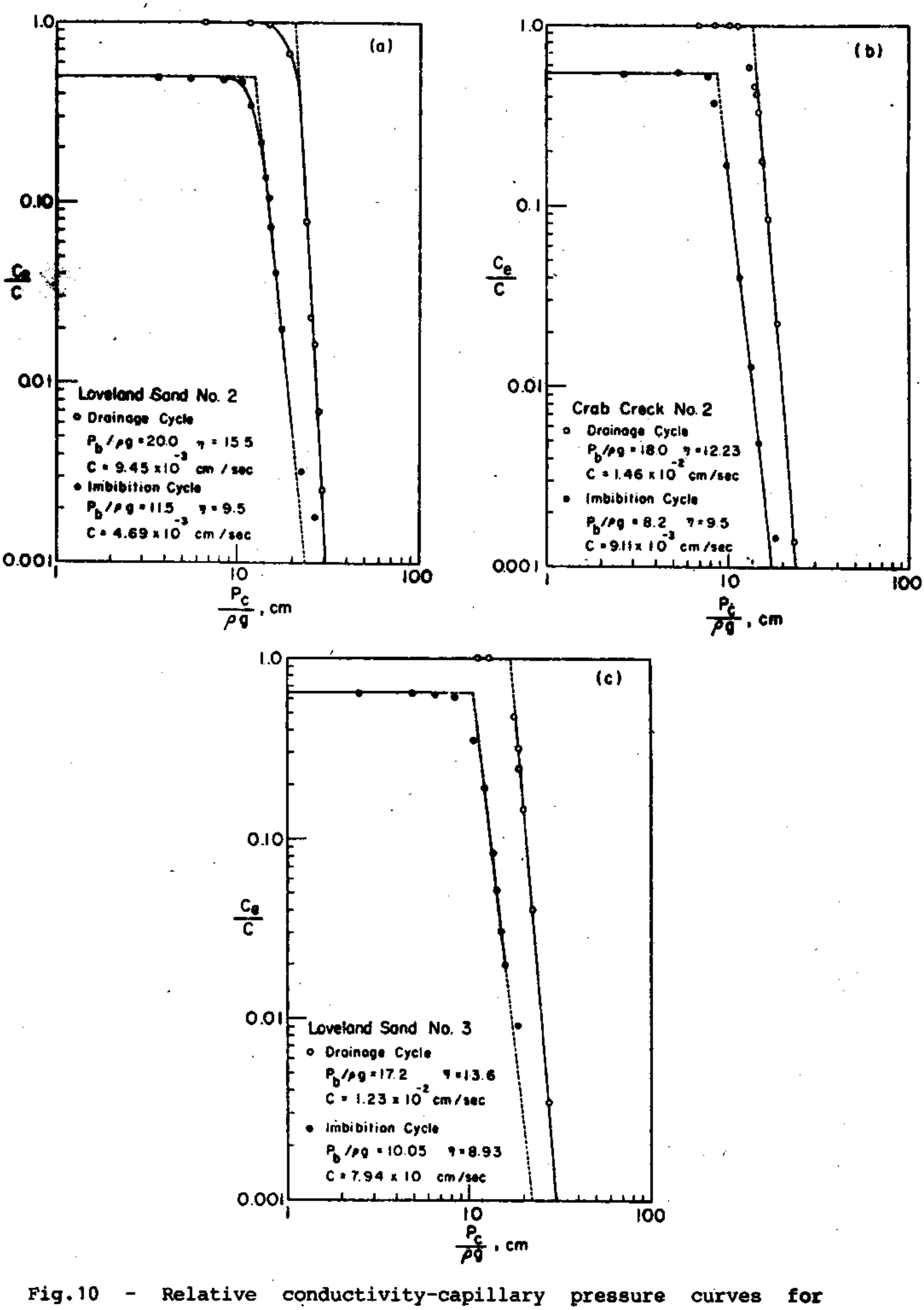

Fig. 10 - Relative conductivity-capillary pressure curves for drainage and imbibition cycles.

or two orders of magnitude less than on the drainage cycle. The exception is for capillary pressures less than the bubbling pressure, in which case, the conductivity on the imbibition cycle is about half that on the drainage cycle, and the values of P_b were about 0.6 those obtained on the drainage cycle.

Bloomsburg and Corey(1964) pointed out that for capillary pressures less than the bubbling pressure on the imbibition cycle, the conductivity is a function of time since the entrapped air will eventually diffuse from the system and the medium will become fully saturated.

2.5.1 Measurement of upward flow rates

The procedure used by Anat for determining maximum upward flow rates differed from that employed by Duke in several ways. The columns used were usually slightly longer than necessary for determining conductivity as a function of capillary pressure. Instead of shortening the column after a series of runs, Anat simulated greater depths to a water table by lowering the inflow siphon. When the system reached a steady state, the lower tensiometer was read and the outflow rate was measured. Knowing the fully saturated conductivity of the soil, it was possible to compute an equivalent depth to a zone of zero capillary pressure by using Darcy law. The first runs were made with the zone of zero capillary pressure at the elevation of the lower tensiometer. The capillary pressure at the surface was gradually increased in very small increments until the outflow appeared to approach a maximum, but the outflow siphon was not lowered more than this. After this, the inflow siphon was lowered to simulate a greater depth to the water table, but the elevation of the outflow siphon was not changed. This procedure resulted in no pressure reversal at the upper tensiometer. Furthermore, the flow rates agreed very closely with those predicted from equation(25). Results are shown in Figure 11. Anat (1965) plotted q_m as a function of d on log log paper. His reason for plotting was that, theoretically, this curve should approach a slope of n at low values of q_m, thus indicating a relationship to the relative conductivity curves shown in Figure 9 and 10. This result was experimentally confirmed by Anat(Figure 11).

2.5.2 Upward flow on the imbibition cycle

One series of runs was also made by starting with completely dry soil and allowing the soil to imbibe liquid from the lower inflow barrier. The initial imbibition was produced with the inflow siphon at a low elevation. The imbibition took place at an extremely slow rate and a considerable time passed before any flow from the outflow siphon was observed. When the inflow siphon was raised, the system reached steady state at increasingly shorter times, and it is probable that the data obtained represented maximum upward flow rates. At any rate, the data for the higher elevation of the inflow siphon agreed very closely with theoretical rates computed using conductivity data obtained on an imbibition cycle as shown in Figure 12.

Another series of runs was made in which the imbibition cycle was started after draining the soil to a high capillary pressure. The inflow siphon was then gradually raised in increments. The flow rates agreed very closely with the theoretical curves obtained by using the soil parameters obtained in an analogous manner, i.e. on the imbibition cycle starting with the soil moist but at a high capillary pressure. The results obtained by Anat are shown in Figure 12A and 12B.

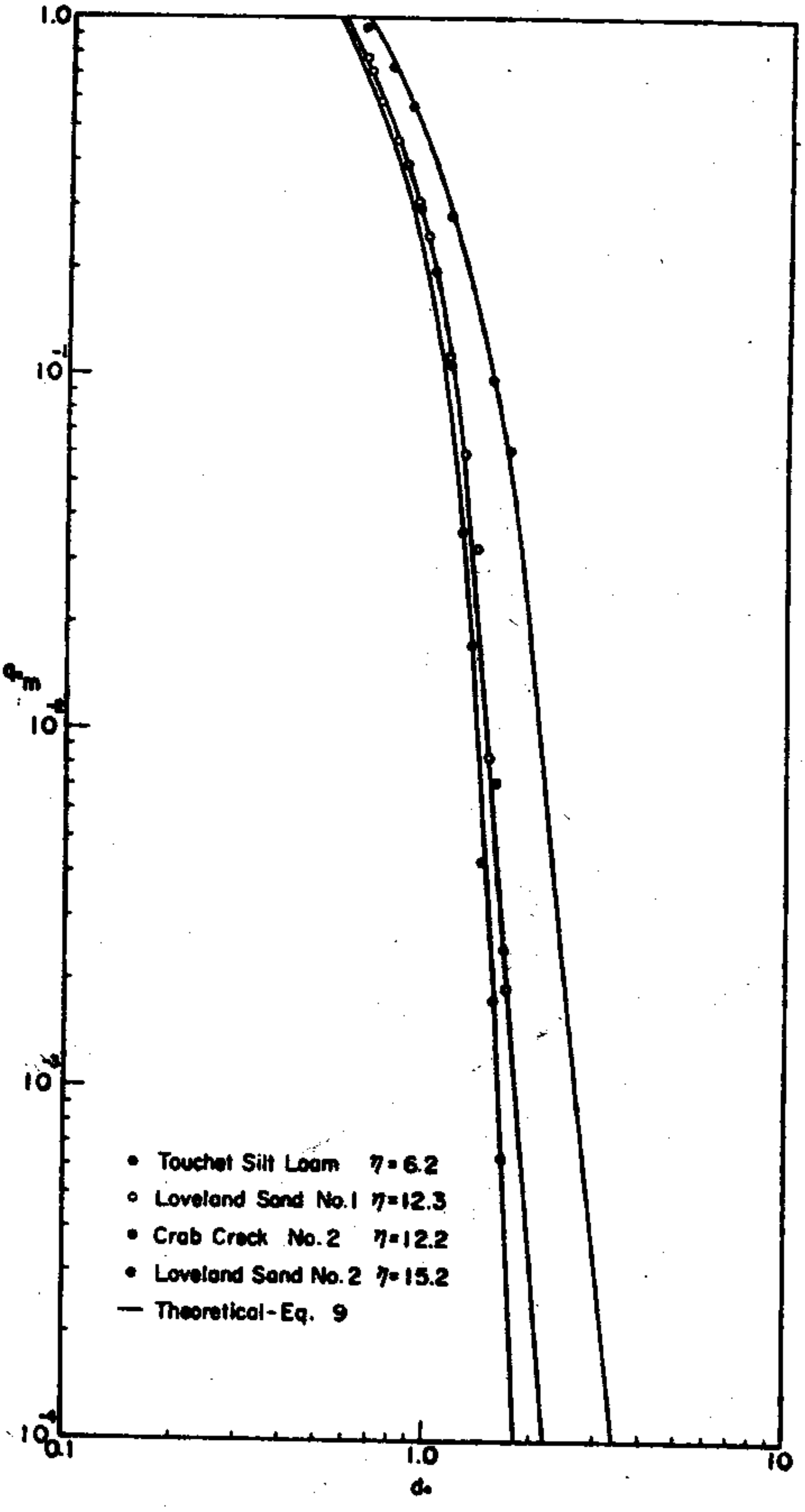

Fig. 11 - Comparison of experimental values of q_m as a function of d with computed values for drainage cycles.

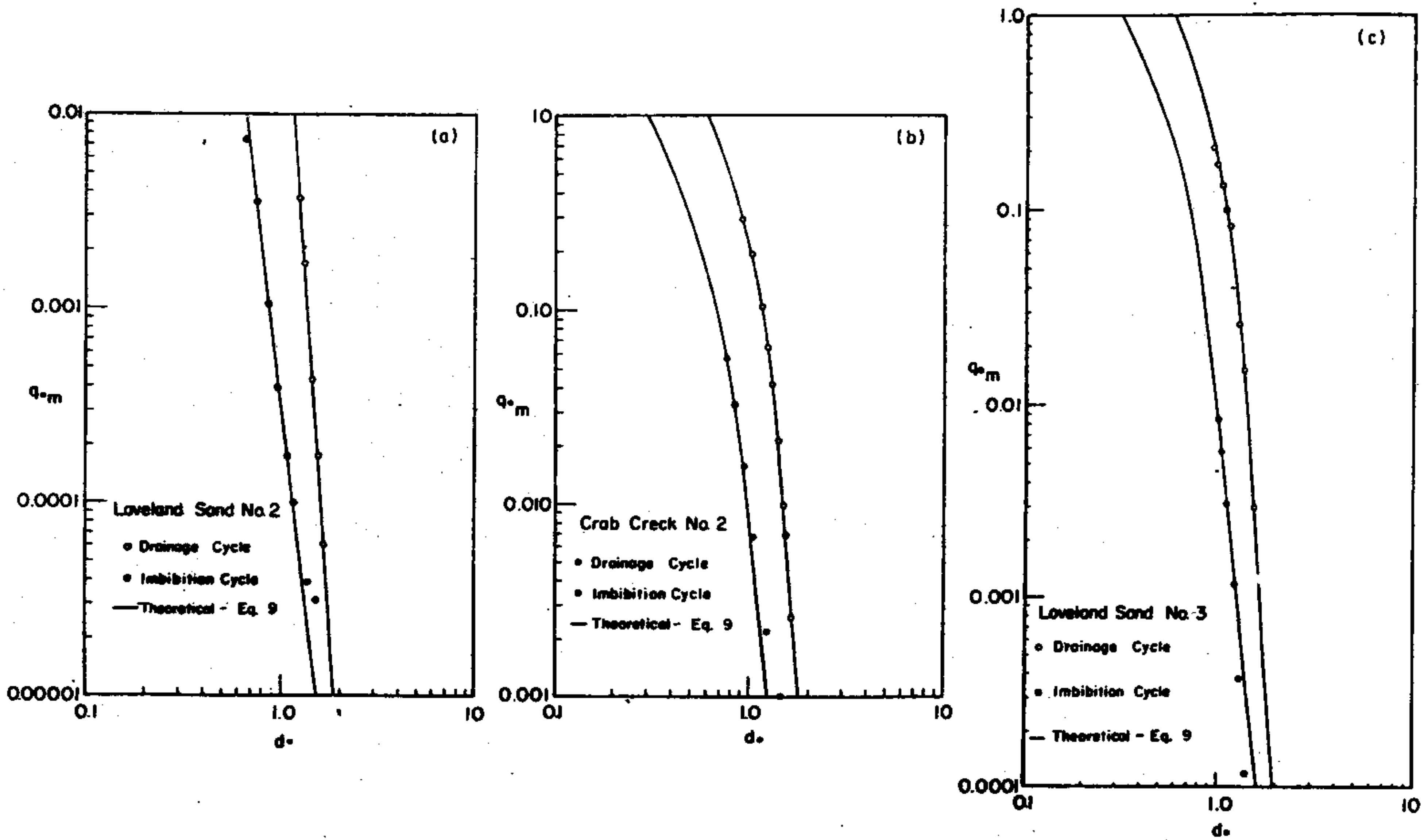

Fig. 12 - Comparison of experimental values of q_m as a function of d with computed values for drainage and imbibition cycles.

2.5.3 Effect of hysteresis

Figure 13 presents the results of one experiment in which the column was first saturated and the outflow siphon was lowered in increments while the zone of zero pressure was maintained at the lower tensiometer. Afterwards the outflow siphon was raised in increments while the zone of zero pressure was maintained at the same elevation. The broken line in Figure 14 represents the theoretical curve for the drainage cycle.

2.6 Lysimeter Study

Shih(1983) conducted an experiment in lysimeters for studying the evaporation from soil surface in relation to water table depth and standard pan evaporation. The experiment were conducted in lysimeters at the Agricultural Research and Education Center, Belle Glade, Florida,USA. The experimental set up is shown in Figure 15. The system consisted of four concrete lysimeters. Each lysimeter was 108 cm long, 76cm wide and 64 cm deep. The lysimeters were designed to maintain a constant water table at given depth. The water table depths studies were 8 cm and 38 cm with one replication. The soil was subjected to three wetting-drying cycles to permit any settling of the material. The STandard Class A National Weather Service Evaporation Pan located at the AERC was used to measure the standard pan evaporation. The distance between the standard class A pan location and lysimeter site was monitored by measuring change in water level at the water supply tank. The Standard Pan Evaporation from the Standard Class A pan was measured daily. A linear regression model was used to express the relationship between soil surface evaporation and standard pan evaporation. The form of the model was:

$$SSE = a_o + a_i \; SPE \qquad \ldots(51)$$

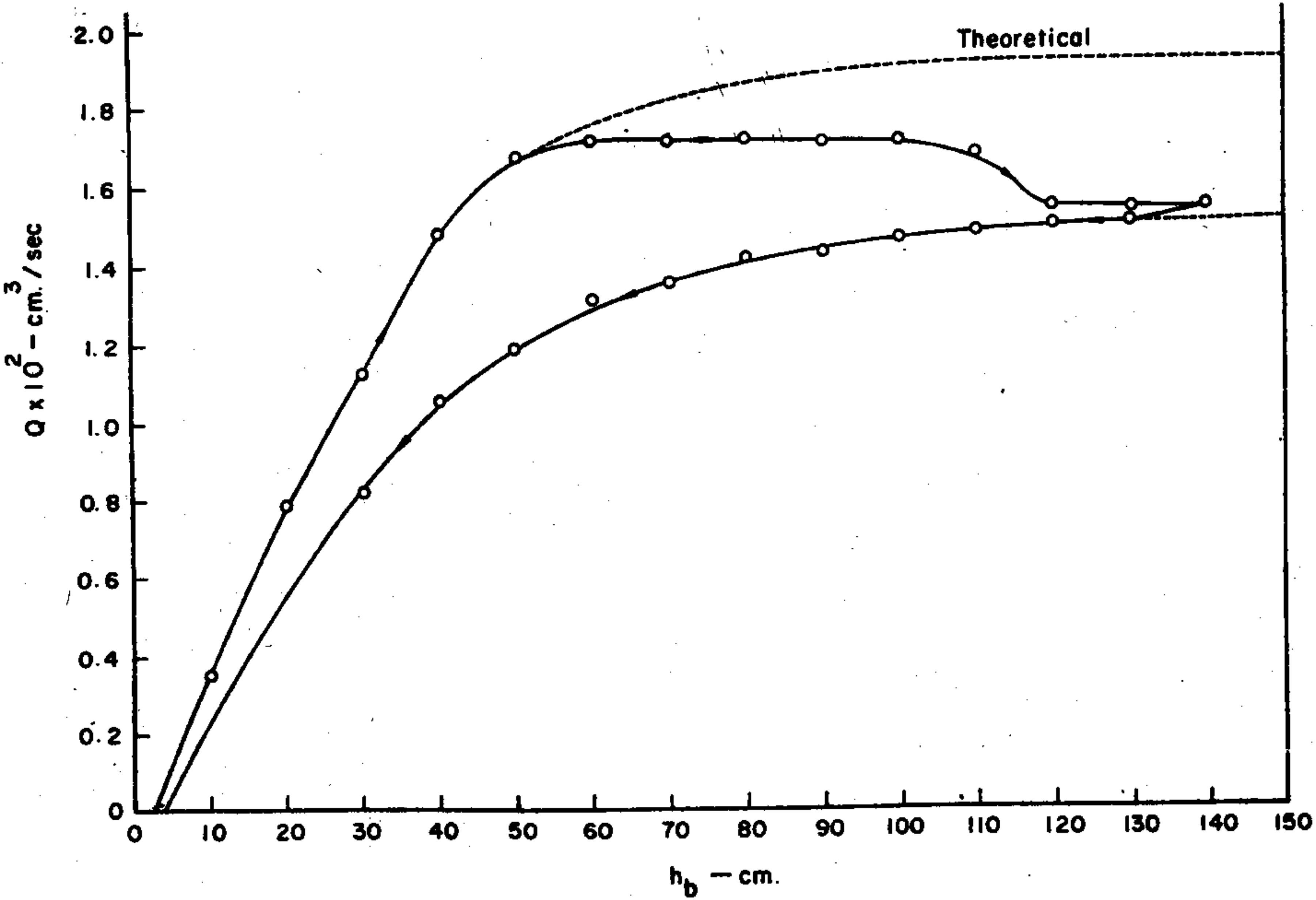

Fig. 13 - Effect of hysteresis on flow rate as a function of negative head at ourflow barrier

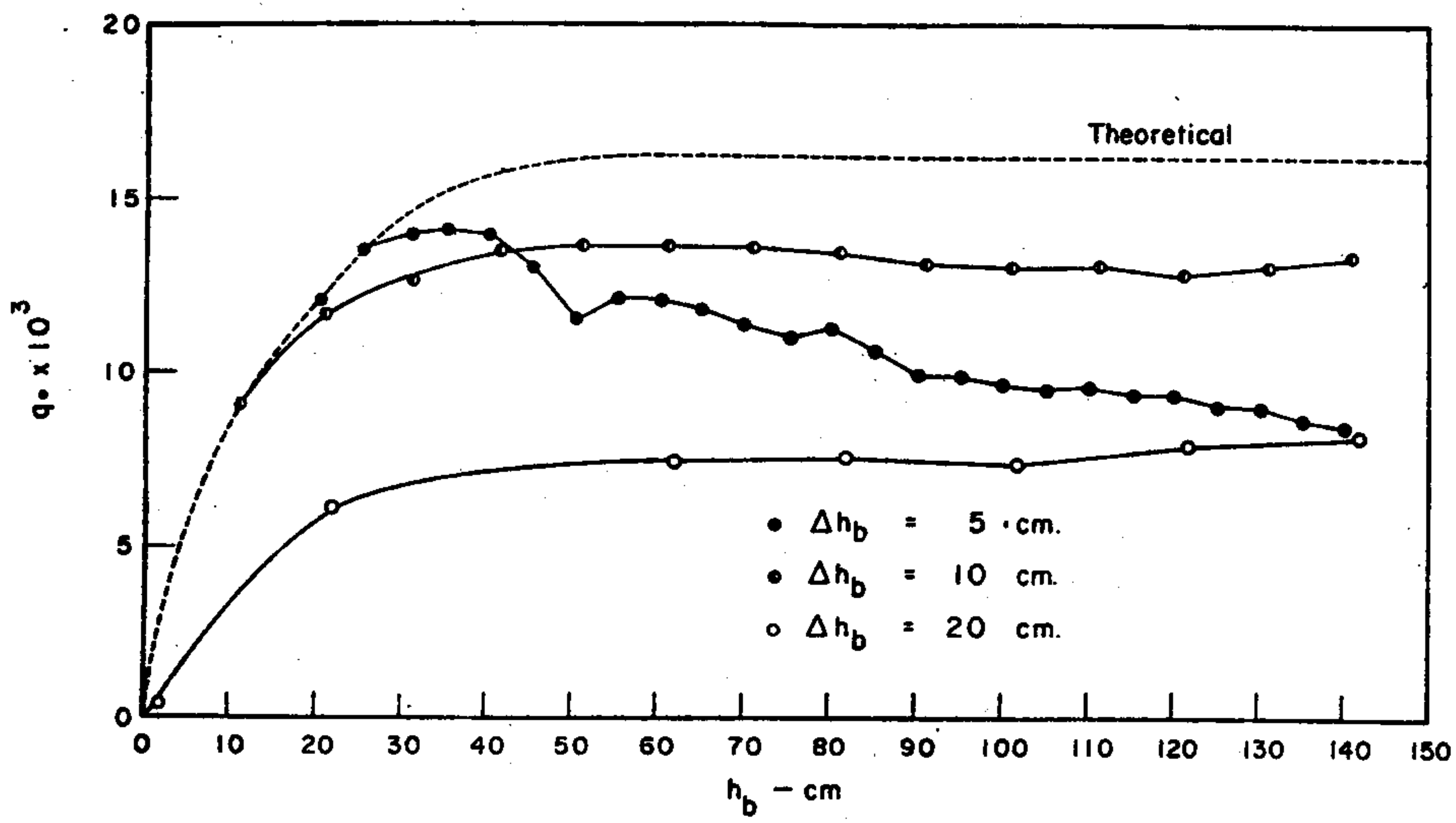

Fig.14 - Effect on flow rate of increment of increase of negative head at outflow barrier.

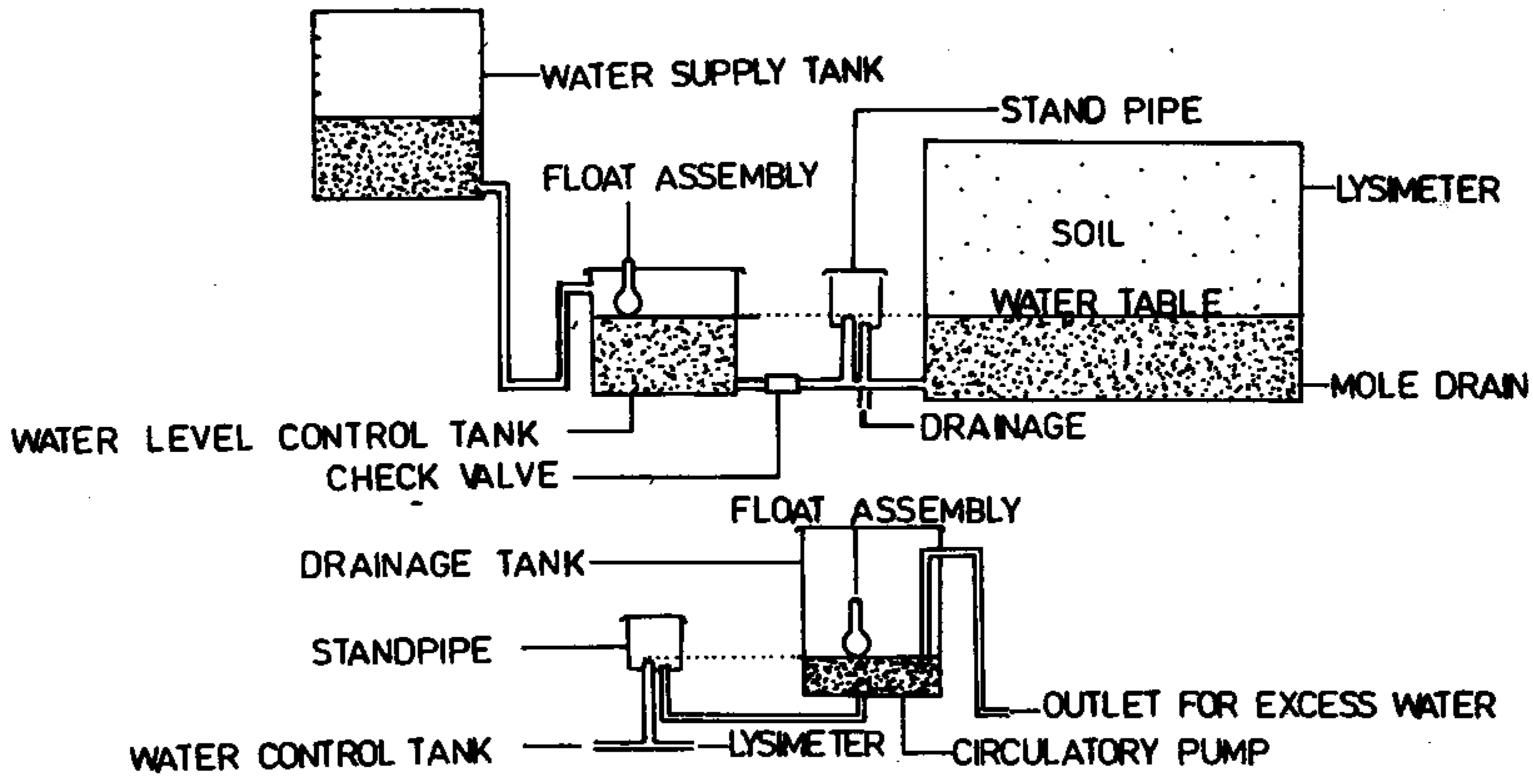

Fig.15 - Water Supply and Drainage System for a Concrete Lysimeter

in which a_o and a_i are coefficients which were estimated from the experimental data by using the regression analysis. A linear regression model was used to analyze the possible relationship between the ratio of soil surface evaporation to standard pan evaporation and the water table depth(WT). The form of the model was:

$$R = b_o + b_i \, WT \qquad \ldots (52)$$

in which b_o and b_i were coefficients which were estimated from the experimental data.

Shih(1983) concluded that the soil surface evaporation for both 8 and 38 cm water depths were directly proporotional to the standard pan evaporation. The soil surface evaporation for 38 cm water table depth was significantly less proportional to the standard pan evaporation than the soil surface evaporation for the 8 cm water table depth.

3.0 REMARKS

Evapotranspiration (ET) is the amount of water transpired by the plant and evaporated from the soil. The knowledge of the amount and rate of ET is useful not only for irrigation studies but also for the development of hydrological models. Evaporation of soil water involves not only loss of water but also the danger of soil salinization.

When the surface is at least partially bare, evaporation can take place from the soil as well as from plants. Since it is generally difficult to separate these two process, they are commonly lumped together and treated as a single process called evapotranspiration.

Many investigators have studied rates of evaporation from soils in which there is a water table. The steady state upward flow of water from a water table through the soil profile to an evaporation zone at the soil surface was first studied by Moore(1939). Theoretical solutions of the flow equation for this process were also given by Philip(1957 a,d) and Gardner(1958). The work done by Moore, Philip, Gardner etc. stressed the cases in which soil properties were the determining factors for evaporation. Some authors like Slayter and Mc Ilory(1961), Van Bavel (1966) etc., have developed some purely meteorological equations where atmospheric conditions play the main role in estimating magnitude of evaporation.

This review note gives in detail the work done in both the cases i.e.estimate of evaporation considering various soil properties and secondly the use of various meteorological equations for estimation of evapotranspiration.

Staley(1957) and Duke(1965) conducted an experiment on wind tunnel at Colorado State University to determine how evaporation rate is affected by depth of water table and wind velocity which has been reported in this review note. Shih's (1983) work on lysimeters for studying the evaporation from soil surface in relation to water table depth and standard pan evaporation has also been mentioned briefly.

Very less work has been done in India in the area of evapotranspiration considering ground water table fluctuation,soil profiles and external environmental conditions etc. The various set of conditions under which evaporation may occur are:

1. a shallow ground water table at a constant or variable depth,
2. soil profile may be shallow resting on bedrock, or it may be deep. Soil profile may be uniform or may consist of distinct layers differing in texture or structure,
3. the soil surface may or may not be covered by a layer of much differing from the soil in hydraulic and diffusive properties,and
4. external environmental conditions may remain constant or fluctuate.

To be more accurate and precise, systematic research work must be carried out for the various soil profiles condition, environmental and water table conditions.

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