MASTER**SOIL WATER REDISTRIBUTION DURING EVAPORATION**

by

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Daniel Dale Fritton

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An Abstract of
A Dissertation Submitted to the
Graduate Faculty in Partial Fulfillment of
The Requirements for the Degree of
DOCTOR OF PHILOSOPHY

Approved:

Don Kirkham
Don Kirkham

Robert W. Shaw
Robert W. Shaw
 In Charge of Major Work

John L. ...
John L. ...
 Head of Major Department

J. B. ...
J. B. ...
 Dean of Graduate College

Iowa State University
 Ames, Iowa

1968

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SOIL WATER REDISTRIBUTION DURING EVAPORATION

Daniel Dale Fritton

Under the supervision of Don Kirkham and R. H. Shaw
From the Department of Agronomy
Iowa State University of Science and Technology

Temperature and moisture distribution data were taken for a 9 by 11 by 19.5 cm. soil column where wind was the evaporation agent and for a similar soil column where radiation was the evaporation agent. A Webster silty clay loam soil characterized by a soil-water retention curve and an inflow diffusivity versus moisture content curve was used in the experiments. The experimental results were compared with results calculated from a solution of the diffusion equation normally used to describe isothermal movement of soil water and a heat and mass transfer solution which accounted for temperature effects. It was found that the isothermal diffusion equation would describe the cumulative evaporation for the wind treatment where the diffusivity had to be divided by a resistance factor and for the radiation treatment where the same diffusivity used for the wind treatment was corrected for an increased temperature. The isothermal diffusion equation, when an adjusted diffusivity coefficient was used with it, would also describe the moisture distributions in the wind case for all times less than 200 hrs. The isothermal diffusion equation, even when an adjusted diffusion coefficient was used with it, would

not predict the formation of a surface layer of dry soil and, thus, did not describe the moisture distributions for the radiation treatment for times greater than 20 hr. or for the wind treatment for times greater than 200 hrs. It was found that the heat and mass transfer equation would predict the development of a dry surface layer and with additional experimental knowledge of the soil-water retention curve as a function of temperature would describe the moisture distribution where temperature gradients were important.

SOIL WATER REDISTRIBUTION DURING EVAPORATION

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Approved:

Don Kirkham

Robert Hobbs
In Charge of Major Work

John Reese
Head of Major Department

JP Paulsen
Dean of Graduate College

Iowa State University
Ames, Iowa

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INTRODUCTION

Evaporation of water from a bare soil surface occurs in nature under a variety of environments. In a shaded environment such as that under a plant canopy, bare soil is subject mainly to evaporation due to unsaturated air movement. Evaporation rates in this case depend on the velocity of wind movement, and evaporation occurs as a nearly isothermal process. In a semiarid environment, on the other hand, bare soil can be subject to evaporation due to widely ranging conditions of incoming radiation, temperature, and wind. Evaporation rates can be very high and evaporation is definitely not an isothermal process. In a semiarid environment, extreme temperature gradients may exist and an analysis of the evaporation process may require more than just an isothermal model even for practical applications. A practical application which seems to have been neglected in the many studies of evaporation using either an isothermal model or a non-isothermal model is the prediction of soil moisture content with depth. It is one purpose of this study to examine the validity of the isothermal diffusion equation of unsaturated soil moisture flow when the equation is applied to prediction of soil moisture content with depth under near isothermal and non-isothermal evaporation conditions. Since it was anticipated in planning this study that the isothermal model would be inadequate for non-isothermal

evaporation, data was taken so that the results could be compared to a theory involving not only transfer of moisture, but also transfer of heat. A second objective of this study is to check the validity of a heat and moisture transfer equation when the equation is applied to prediction of soil moisture content with depth for non-isothermal evaporation conditions.

Sections on gamma-ray attenuation and liquid moisture diffusivity are included in this dissertation and are related to the problem of soil-water redistribution during evaporation through their use in the experimental and theoretical parts of the study. Gamma-ray attenuation techniques were used throughout this study to measure the moisture content of the soil. During the course of this study, we questioned whether the gamma-ray attenuation technique of measuring soil moisture content was as accurate as it might be if a correction was introduced for the resolving time of the apparatus. This question was answered with several experiments which were designed to measure the resolving time of our apparatus and to check the gamma-ray measured water content with the standard gravimetrically measured water content.

REVIEW OF LITERATURE

In the sections which follow, four areas of literature will be reviewed. These are: gamma-ray attenuation, liquid moisture diffusivity, isothermal diffusion, and heat and mass transfer. The areas will be reviewed in the above order because gamma-ray procedures were used in all experiments to determine moisture content, and thus some knowledge of gamma-ray techniques is needed in several chapters before the other topics can be covered. In a like manner, the liquid moisture diffusivity area is covered before the last two areas where use is made of the diffusivity function. The area on isothermal diffusion is covered before the area on heat and mass transfer because isothermal diffusion has been used more frequently to explain soil water movement and is the earlier theory. All of the areas will be reviewed as they relate to the experiments that will be described later and to the general problem of soil-water redistribution during evaporation.

Gamma-ray Attenuation

Resolving time and mass absorption coefficient

To understand the literature which shows the need for a resolving time correction for the gamma-ray method of soil moisture measurement, certain concepts are needed; one of which is the concept of resolving time. Resolving time of a

gamma-ray counting system is the minimum time that may separate two consecutive gamma-ray photons. A photon that arrives at the scintillation crystal before the minimum time has elapsed will not be recorded. This loss of counts is called coincidence loss. An observed counting rate can be corrected by using the experimentally determined resolving time and the equation (Overman and Clark, 1960)

$$I = R/(1 - TR) \quad (1)$$

where

I = true counting rate (counts/min.)

R = observed counting rate (counts/min.)

T = resolving time (min./count)

A percentage correction for the observed counting rate can be calculated from the equation (Overman and Clark, 1960)

$$\text{percentage correction} = [TR/(1-TR)](100) \quad (2)$$

When $T = 5 \text{ } \mu\text{sec./count}$ or $0.0833 \times 10^{-6} \text{ min./count}$, the percentage correction is 0.8% for $R = 10^5 \text{ counts/min.}$ and 9.1% for $R = 10^6 \text{ counts/min.}$ This correction varies with the counting rate and thus will cause the gamma-ray attenuation equation (Davidson, Biggar, and Nielsen, 1963)

$$\ln R/R_0 = -\mu_R \rho x \quad (3)$$

where

R_0 = counting rate with zero thickness of material

(counts/min.)

u_a = apparent mass absorption coefficient of a material ($\text{cm}^2/\text{g}.$)

ρ = density of material (g./cm^3)

x = thickness of a material normal to the gamma-ray beam ($\text{cm}.$)

to depend on the resolving time as well as on all factors in Equation 3. If the mass absorption coefficient is found by plotting $\ln R/R_0$ versus $-px$, then the mass absorption coefficient depends on both the true mass absorption coefficient and the resolving time.

Data in several publications indicate that a resolving time correction can change the values used for the experimentally determined mass absorption coefficients of water and soil. The theoretical value for the mass absorption coefficient of water is $0.0857 \text{ cm}^2/\text{g}.$ (Grodstein, 1957). Gurr (1962) experimentally determined the mass absorption coefficient of water to be $0.0856 \text{ cm}^2/\text{g}.$ after correcting for resolving time. Davidson, Biggar, and Nielsen (1963), Reginato and Van Bavel (1964), and Kirkham, Rolston, and Fritton (1967) did not use a correction for resolving time and found low values of 0.0815, 0.0748, and $0.0802 \text{ cm}^2/\text{g}.$, respectively.

Theoretical and experimental mass absorption coefficients for dry soil also show discrepancies. The theoretical values average $0.0775 \text{ cm}^2/\text{g}.$ (Reginato and Van Bavel, 1964) whereas,

all values for the mass absorption coefficient of soils determined without a resolving time correction, except Davidson, Biggar, and Nielsen (1963) who found $0.0769 \text{ cm.}^2/\text{g.}$, are smaller than $0.0710 \text{ cm.}^2/\text{g.}$ (Shalhevet and Yaron, 1967; Van Bavel, Underwood, and Ragor, 1957; Kirkham, Rolston, and Fritton, 1967; Reginato, and Van Bavel, 1964; Band Van Bavel, 1959). Gurr. (1962), who corrected for resolving time, found a value of $0.07785 \text{ cm.}^2/\text{g.}$

Bulk density

Kirkham, Rolston, and Fritton (1967) derived the expressions used to calculate bulk density and moisture content from gamma-ray attenuation data. In their derivation, the gamma intensities were measured after the gamma rays passed through an oven dry soil. When gamma intensities are measured after passing through an air dry soil of known gravimetric water content, similar equations can be derived.

Essential steps of the derivation which follows are from W. H. Gardner (1965) and from Kirkham, Rolston, and Fritton (1967). Assume that gamma intensity measurements have been taken through an empty container and that the container has a known inside dimension normal to the gamma beam of y units. The intensity of gamma radiation through the empty container is given by an equation similar to Equation 3, where the intensities have been corrected for resolving time and the true mass absorption coefficient is used. The equation is

$$I_c = I_a \exp (-\mu x)_c \quad (4)$$

where

I = radiation intensity (counts/min.)

a = subscript denoting air

c = subscript denoting the container wall

u = mass absorption coefficient ($\text{cm}^2/\text{g}.$)

ρ = density (g./cm^3)

x = thickness perpendicular to the gamma beam (cm.)

Similarly for the container filled with air dry soil, the intensity of gamma radiation through the air dry soil is given by

$$I_{ads} = I_a \exp [(-u\rho x)_c - (u\rho x)_s - (u\rho x)_w] \quad (5)$$

where

ads = subscript denoting air dry soil

s = subscript denoting soil

w = subscript denoting water

The thickness of bulk soil is given in Equation 5 by x_s and will be equivalent to the thickness of the container in which it is enclosed. Thus

$$x_s = y \quad (6)$$

Also, the thickness of water contained in the air dry soil can be related to the thickness of the soil and the volumetric water content by the relation

$$x_w = x_s \theta \quad (7)$$

where

Θ = volumetric water content (cm^3/cm^3)

The air dry volumetric water content can then be converted to the air dry water content on a weight basis using the equation

$$\Theta_{\text{ads}} = W \rho_s / \rho_w \quad (8)$$

where

W = air dry gravimetric water content (g./g.)

and therefore

$$x_w = y W \rho_s / \rho_w \quad (9)$$

When Equation 9 is substituted into Equation 5, the result is

$$I_{\text{ads}} = I_a \exp [-(u\rho x)_c - (u\rho x)_s - (u_w y \rho_s W)] \quad (10)$$

Then Equation 10 is divided by Equation 4 and the result is

$$I_{\text{ads}}/I_c = \exp [-(u\rho x)_s - (u_w y \rho_s W)] \quad (11)$$

After the natural logarithm of each side is taken and Equation 11 is simplified, the result is

$$\ln (I_{\text{ads}}/I_c) = -\rho_s y (u_s + u_w W) \quad (12)$$

And the bulk density is given by

$$\rho_s = [-\ln (I_{\text{ads}}/I_c)]/[y(u_s + u_w W)] \quad (13)$$

Gurr (1962) gives a similar equation.

Volumetric moisture content

When the soil is wet to some unknown moisture content and the bulk density of the soil is known, the volumetric moisture content can be calculated from the intensity of gamma radiation after it passes through the wet soil which is in a container. The intensity of gamma radiation after it has passed through the container and the wet soil is given by

$$I_{ws} = I_a \exp [-(\mu x)_c - (\mu x)_s - (\mu x)_w] \quad (14)$$

where

ws = subscript denoting wet soil

When Equation 7 is substituted into Equation 14, the result divided by Equation 10, and the natural logarithm of both sides of the new result is taken, the result is

$$\ln (I_{ws}/I_{ads}) = -(\mu\rho)_w x_s \Theta + u_x y \rho_s W \quad (15)$$

And the volumetric water content is given by

$$\Theta = [\ln(I_{ws}/I_{ads}) - u_w y \rho_s W] / [-(\mu\rho)_w y] \quad (16)$$

Liquid Moisture Diffusivity

Philip (1955) gives the historical development leading to the derivation of the isothermal diffusion equation for unsaturated soil water flow where the diffusion coefficient called the diffusivity is dependent on the concentration of water. Since that time, many experimental techniques have been developed to measure the diffusivity as a function

of moisture content. Klute (1965) lists several methods which can be used to measure the diffusivity and describes in detail a pressure plate outflow method. In addition to the methods described by Klute, there are those described by Doering (1965) and Whisler, Klute, and Peters (1968). The technique of Whisler, Klute, and Peters (1968) was used in this study because the equipment was readily available and there are fewer assumptions contained in the theory than contained in the theory which is used for outflow techniques. At the time the choice of a method was made, the possibility that measurements from inflow and outflow techniques might give different results was not fully appreciated. After the study was completed, a search was made for data which could be used to compare inflow and outflow diffusivity techniques, and results from several workers follow.

W. R. Gardner (1959) calculated the absorption and desorption relative weighted mean diffusivities and found that they both increased with increasing moisture content. The absorption diffusivity was greater than the desorption diffusivity by a factor of approximately two at low water contents and a factor of approximately five at higher water contents.

Klute, Whisler, and Scott (1964) used radial flow pressure cells to measure both inflow and outflow diffusivities. They found that diffusivities increased with moisture content for outflow measurements and decreased with moisture content for

inflow measurements. Differences between inflow and outflow measurements were as great as a factor of 100. Their measurements for the most part were taken between 20 and 365 cm. of water tension except one measurement which was taken between 365 and 1550 cm. of water tension. Two sandy loam soils were used in the study.

Inflow data which was obtained by Bruce and Klute (1956) indicated that inflow diffusivities may increase with increasing water content to some maximum and then decrease. Bruce and Klute's data were also taken for sandy soils.

Davidson, Biggar, and Nielsen (1963) used the method of Bruce and Klute (1956) for a silt loam soil and found an increasing diffusivity for increasing water content for the range, 10 to 40 percent water content, that was investigated.

Jackson (1964) measured absorption and desorption vapor diffusivities. He found minor differences between the sorption and desorption diffusivities with each showing a peak between 0 and 10 percent water content for most cases.

The above results seem to indicate that the shape of the diffusivity curve depends on the range in tension for which measurements are made for both inflow and outflow techniques. When the entire range of moisture contents is considered, it would seem that both inflow and outflow diffusivity curves would tend to increase with increasing moisture content for a silty clay loam soil but may decrease for narrow ranges of

moisture content at very low and very high moisture contents.

Wiegand and Taylor (1961) summarize some general conclusions about systems in which the diffusivity increases with moisture content. These conclusions are: there will be a linear relation between the amount of evaporation and the square root of time for a semi-infinite soil column for small times; when the above proportionality ceases to become linear, the curves become concave toward the square root of time axis and approach a final value asymptotically; desorption curves will be linear for a shorter period of time than the absorption curves; desorption curves will be more sensitive to the form of the diffusivity than will absorption curves; and desorption will be slower than absorption.

Since the technique of Whisler, Klute, and Peters (1968) was used in this study, the theory for this technique will be described starting with the isothermal diffusion equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left(D(\theta) \frac{\partial \theta}{\partial x} \right) \quad (17)$$

where

- θ = volumetric moisture content (cm^3/cm^3)
- $D(\theta)$ = diffusivity ($\text{cm}^2/\text{min.}$)
- t = time (min.)
- x = distance (cm.)

When water is allowed to flow into a horizontal soil column, the boundary conditions are

$$\theta(0, t) = \theta_c \quad t > 0 \quad (18)$$

where

θ_c = constant moisture content near saturation

and

$$\lim_{x \rightarrow \infty} \theta(x, t) = \theta_d \quad t \geq 0 \quad (19)$$

where

θ_d = initial dry moisture content

The initial condition is

$$\theta(x, 0) = \theta_d \quad x \geq 0 \quad (20)$$

The Boltzmann transformation

$$\lambda(\theta) = x/t^{1/2} \quad (21)$$

can be applied to the diffusion equation to give the equation for the diffusivity

$$D(\theta) = -1/2 \frac{d\lambda}{d\theta} \int_{\theta_d}^{\theta} \lambda d\theta \quad (22)$$

Now if x is fixed at some point x_1 , we find

$$D(\theta) = 1/4 \frac{(x_1)^2}{t^{3/2}} \left(\frac{dt}{d\theta} \right) \int_{\theta_d}^{\theta} 1/t^{1/2} d\theta \quad (23)$$

In Equation 23, experimental data are used to evaluate the right side so that the diffusivity can be calculated as a function of water content. Judging from the experimental

results given by Whisler, Klute, and Peters, the diffusivity that is calculated in Equation 23 is the liquid moisture diffusivity which we will denote by $D_{\theta, \text{liq}}$ for purposes which will become apparent. The experimental technique used to evaluate the right side of Equation 23 is not sensitive enough to pick up movement of water in the vapor phase.

Isothermal Diffusion

W. R. Gardner (1959) analyzed the problem of evaporation from a bare soil using the isothermal diffusion equation (Equation 17). He used an evaporation potential, 4 cm./day, produced by a combination of air movement and heat lamps. He compared isothermal diffusion theory for the finite case with experimental results by fitting cumulative evaporation curves for four lengths of column and four initial water contents. In all cases, he was able to fit the cumulative evaporation curves using a diffusivity function nearly identical with one measured with the pressure plate outflow method.

W. R. Gardner also gives results for the solution of the diffusion equation for the semi-infinite case. In this case, the isothermal diffusion equation predicts that the cumulative evaporation is a function of the square root of time as long as the use of the Boltzmann transformation is valid, i.e., as long as the differential equation and the boundary condition are valid. Gardner showed that this was true experimentally for a long soil column for which he used a clay soil and an evaporation potential of 1 cm./day.

For the finite case, W. R. Gardner presents a theoretical graph of cumulative evaporation versus time. He also presents theoretical graphs for the moisture content versus depth for both the finite and semi-infinite cases. Although Gardner compares theoretical and calculated cumulative evaporation data, he shows no experimental data to compare moisture content at different depths with his theoretical results from the diffusion equation.

Jackson (1964) calculated desorption diffusivities from experimental moisture content data for horizontal distances in a soil column using an equation similar to Equation 23 except that time was held constant. The experimental curves of moisture content which he shows are characteristic of the type found when the isothermal diffusion equation is solved.

Wiegand (1960) found experimental curves of moisture content versus depth of what he calls parabolic form. These curves are similar to the theoretical curves shown by W. R. Gardner (1959). He indicates that this shape of curve has been predicted by a number of theories from analysis of surface tarnishing reactions of metal and drying of wood and clay. The theory for drying of wood and clay used a diffusion equation which considered the diffusion coefficient to be a constant.

Heller (1967) used a form of Darcy's Law to solve for the pressure distribution that would exist in a drying vertical

porous column. The solution requires the knowledge of the furthest penetration of air into the column for the constant rate of evaporation stage. When the evaporation rate decreases, he introduces the thickness of the dry layer at the surface as a region of vapor diffusion and solves for the rate at which this thickness increases. The thickness of the dry layer at the surface was predicted to increase with the square root of time. The same statement about the increase in dry layer thickness with the square root of time was made by W. R. Gardner (1959) although he showed no mathematical derivation. The analysis of Heller assumed that the column was maintained at a constant temperature with an evaporation rate low enough that significant temperature gradients did not occur. Heller shows some experimental data for columns of fluorite saturated with chloroethene which shows that the thickness of the dry zone is a function of the square root of time for the first 4 cm. thickness of the dry zone in one case, and for 7 cm. thickness of the dry zone in a second case. His columns were 16 cm. long.

W. R. Gardner and Hillel (1962) compared cumulative evaporation curves under different evaporation potentials with the isothermal diffusion model, Equation 17. Their evaporation potentials varied from 4 cm./day to 0.3 cm/day. The potential evaporation was obtained using both wind and radiation for the high rate, and wind alone for the lower rates

of evaporation potential. They found that the length of the period of constant evaporation rate could be predicted using results from the diffusion equation. They also found that the cumulative evaporation curve for infinite evaporation potential could be translated along the time axis to a point where the constant rate of evaporation period ended and that this curve would fit relatively well for all of the evaporation potentials. They took some temperature measurements and found that the worst fit of the cumulative evaporation curves was found for the one case where there was a temperature gradient downward into the column. There were no data presented for the variation of moisture content with depth. These data indicated that temperature gradients need to be considered in moisture transfer.

Hanks and H. R. Gardner (1965) used the diffusion equation to investigate by calculation only the influence of the variation of the diffusivity function on the evaporation of soil water. They calculated both cumulative evaporation and moisture content versus depth for a homogeneous and a two layered soil. They found that changes in the diffusivity at low moisture contents, less than 10 percent, made no significant difference in the cumulative evaporation or the moisture content versus depth profiles. However, they did not use the diffusion equation for moisture contents in this range, less than 10

percent, except at the very surface. Thus this conclusion may not hold for soil columns that have a considerable portion of their depth below 10 percent moisture content. Variation of the diffusivity at the moisture contents greater than 10 percent gave large differences in cumulative evaporation and in the moisture content versus depth profiles. The change in shape of the moisture profiles due to variations in diffusivity was not significant. H. R. Gardner and Hanks used the numerical solution of the diffusion equation given by Hanks and Bowers (1962). This numerical solution is similar to the one used by W. R. Gardner (1959) for his finite case. Since the same numerical techniques will be used to analyze the experimental data in this study, this numerical solution will be described in detail. Some comments on the Hanks and Bowers solution are interpolated.

The starting point of the numerical solution is the isothermal diffusion equation (Equation 17) written in a slightly different form

$$\partial \theta / \partial t = \partial [K \partial H / \partial x] / \partial x \quad (24)$$

where

θ = volumetric water content (cm^3/cm^3)

H = hydraulic head (cm. of water)

x = distance (cm.)

K = hydraulic conductivity (cm./hr.)

The origin of the coordinates was taken so that the distance into the soil was negative (i.e. vertical upward is positive).

When the diffusion equation is used to describe soil water movement, the described soil will conduct water at a rate restricted only by the diffusivity or conductivity relationships. Most soils have diffusivities and conductivities for moisture contents near the saturation moisture content that are large enough that the soil is theoretically capable of conducting water to the surface faster than the water can be removed. The boundary conditions which will now be given will impose a further restriction on the diffusion equation so that water will never move to the soil surface faster than it can be removed by the atmosphere. The boundary condition which will restrict the water movement during the early stages of evaporation is

$$\theta - \theta_1(0) \quad E \geq EP \quad x = 0 \quad t \geq 0 \quad (25)$$

where

θ_1 = initial moisture content (cm.³/cm.³)

E = evaporation rate from the column (cm.)

EP = potential evaporation rate (cm.)

The boundary conditions which controlled the evaporation after the soil was no longer able to conduct water to the surface at a rate greater than the evaporation potential were

$$\theta = \theta_0(0) \quad E < EP \quad x = 0 \quad t \geq 0 \quad (26)$$

where

θ_0 = moisture content at equilibrium with the atmospheric conditions

and

$$K \frac{\partial H}{\partial x} = 0 \quad x = -L \quad t \geq 0 \quad (27)$$

The initial conditions were

$$\theta = \theta(x) \quad -L < x < 0 \quad t = 0 \quad (28)$$

The numerical form of Equation 24 is

$$\frac{\theta_i^j - \theta_i^{j-1}}{\Delta t} = \frac{(h_{i-1}^{j-1} + h_{i-1}^j + 2G - h_i^{j-1} - h_i^j) K_{i-1/2}^{j-1/2}}{2(\Delta x)^2} - \frac{(h_i^{j-1} + h_i^j + 2G - h_{i+1}^{j-1} - h_{i+1}^j) K_{i+1/2}^{j-1/2}}{2(\Delta x)^2} \quad (29)$$

where

h = matric potential (cm. of water)

$G = \Delta x$ = gravitational term (cm.)

i = distance subscript

j = time subscript

In order to change the moisture content in Equation 29 to the matric potential, we must assume a unique relationship

between matric potential and moisture content. Then the difference on the left side of Equation 29 can be written as

$$\frac{\theta_1^j - \theta_1^{j-1}}{\Delta t} = \frac{(h_1^j - h_1^{j-1}) C_1^{j-1/2}}{\Delta t} \quad (30)$$

where

$$C_1^{j-1/2} = \left(\frac{\partial \theta}{\partial h} \right)_1^{j-1/2}$$

When this quantity is substituted into Equation 29, the numerical form of Equation 24 becomes

$$\begin{aligned} \frac{h_1^j - h_1^{j-1}}{\Delta t} = & \frac{(h_{i-1}^{j-1} + h_{i-1}^j + 2G - h_i^{j-1} - h_i^j) K_{i-1/2}^{j-1/2}}{2(\Delta x)^2 C_1^{j-1/2}} \\ & - \frac{(h_i^{j-1} + h_i^j + 2G - h_{i+1}^{j-1} - h_{i+1}^j) K_{i+1/2}^{j-1/2}}{2(\Delta x)^2 C_1^{j-1/2}} \end{aligned} \quad (31)$$

In order to use this equation, the conductivity was expressed in terms of liquid water diffusivity. Hanks and Bowers state that the use of the diffusivity gives better results for numerical calculations because the diffusivity does not vary as widely as the conductivity. The expression used to convert the conductivity to the diffusivity was

$$K_{i-1/2}^{j-1/2} = D_{i-1/2}^{j-1} \frac{(\theta_{i-1}^{j-1} - \theta_i^{j-1})}{(h_{i-1}^{j-1} - h_i^{j-1})} \quad (32)$$

The diffusivity used in Equation 32 was an average diffusivity determined from the experimental data by the relation

$$D_{i-1/2}^{j-1} = \frac{\sum_{\theta=\theta_L}^{\theta_{i-1}^{j-1}} D_{\theta, \text{liq}} \Delta\theta - \sum_{\theta=\theta_L}^{\theta_i^{j-1}} D_{\theta, \text{liq}} \Delta\theta}{(\theta_{i-1}^{j-1} - \theta_i^{j-1})} \quad (33)$$

where

θ_L = lowest value of θ for which $D_{\theta, \text{liq}}$ has been determined.

$D_{\theta, \text{liq}}$ = liquid moisture diffusivity (cm^2/hr)

Hanks and Bowers (1962) then used the technique of Richtmyer (1957, pp. 101-104) to solve the tridiagonal matrix resulting from Equation 31.

Heat and Mass Transfer

Philip and De Vries (1957) derived flow equations for liquid and vapor water transfer under a simultaneous temperature and moisture gradients. The equation for vapor flow is

$$q_{\text{vap}}/\rho_w = -D_{T, \text{vap}} \left(\frac{dT}{dz} \right) - D_{\theta, \text{vap}} \left(\frac{d\theta}{dz} \right) \quad (34)$$

where

q_{vap} = vapor flux density ($\text{g./cm}^2\text{sec.}$)

ρ_w = density of liquid water (g./cm^3)

$D_{T, \text{vap}} = f D_{\text{atm}} v \beta h \frac{a}{dT}$ = thermal vapor moisture diffusivity ($\text{cm}^2/\text{sec.}^\circ\text{C}$)

(35)

$f = a + \theta \quad \theta < \theta_c \quad (\text{cm}^3/\text{cm}^3)$

- ρ_o = saturated water vapor density (g./cm.³)
 g = acceleration due to gravity (cm./sec.²)
 $\rho = \rho_o h = \rho_o \exp(g\psi/RT)$ = water vapor density (g./cm.³)
 R = gas constant of water vapor (erg./g. °C)
 ψ = matric potential (cm. of water)

The equation for liquid flow is

$$q_{liq}/\rho_w = - D_{T,liq} dT/dz - D_{\theta,liq} d\theta/dz - K \quad (39)$$

where

- q_{liq} = liquid moisture flux (g./cm.² sec.)
 $D_{T,liq} = (30.0) (K\gamma\psi)$ = thermal liquid diffusivity (cm.²/sec. °C) (40)
 K = hydraulic conductivity (cm./sec.)
 $\gamma = (1/\sigma) d\sigma/dT$ = temperature coefficient of surface tension of water (°C⁻¹)
 $D_{\theta,liq} = K(\partial\psi/\partial\theta)$ = liquid moisture diffusivity (cm.²/sec.) (41)

When Equations 34 and 39 are combined, the equation for total moisture flow is

$$q/\rho_w = - D_T dT/dz - D_\theta d\theta/dz - K \quad (42)$$

where

- q = moisture flux (g./cm.² sec.)
 $D_T = D_{T,vap} + D_{T,liq}$ = thermal moisture diffusivity (cm.²/sec. °C) (43)

$$D_{\theta} = D_{\theta, \text{vap}} + D_{\theta, \text{liq}} = \text{moisture diffusivity (cm.}^2/\text{sec.)} \quad (44)$$

An equation for heat transfer which is similar to Equation 42 is given by Philip (1957) and is rewritten here as

$$Q = - \lambda dT/dz - \rho_w L D_{\theta, \text{vap}} d\theta/dz \quad (45)$$

where

Q = soil heat flux density, positive upward
(cal./cm.²sec.)

λ = thermal conductivity which includes that portion
due to vapor movement (cal./cm.sec. °C)

L = latent heat of evaporation of water (cal./g.)

Philip (1957) eliminates the dT/dz term from Equations 42 and 45 to obtain the expression

$$d\theta/dz = -[\lambda(K+E) - D_T Q] / [\lambda D_{\theta} + \rho_w D_T L D_{\theta, \text{vap}}] \quad (46)$$

When Equation 46 is inverted and the variables separated, we get

$$dz = -\{[\lambda D_{\theta} + \rho_w D_T L D_{\theta, \text{vap}}] / [\lambda(K+E) - D_T Q]\} d\theta \quad (47)$$

which, when integrated on both sides, may be written as

$$\int_0^z dz = - \int_{\theta(0)}^{\theta(z)} \{[\lambda D_{\theta} + \rho_w D_T L D_{\theta, \text{vap}}] / [\lambda(K+E) - D_T Q]\} d\theta \quad (48)$$

and when the integrations are carried out, as

$$z = - \int_{\theta(0)}^{\theta(z)} \{[\lambda D_{\theta} + \rho_w D_T L D_{\theta, \text{vap}}] / [\lambda(K+E) - D_T Q]\} d\theta \quad (49)$$

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From Equation 49, we can calculate the moisture content as a function of depth with the only knowledge required being the surface moisture content and the functions under the integral. Equation 49 is used in this study to describe moisture content versus depth for the case of simultaneous heat and mass transfer.

De Vries (1958) extended the treatment of these previous papers to include the heat of wetting, the transfer of sensible heat by the vapor, and the changes of moisture content in the vapor phase. These effects are all assumed to be negligible in the present study.

Taylor and Cary (1964) derived equations for the simultaneous flow of moisture and heat in a moist soil from the theory of irreversible thermodynamics. Their equations are of the same general form as Equations 42 and 45.

Several researchers have used either the analysis of Philip and De Vries (1957) or Taylor and Cary (1964) for various purposes. Both theories have found some experimental verification in several studies in the literature which follow.

Hanks; H. R. Gardner, and Fairbourn (1967) determined experimental moisture and temperature profiles for a silt loam, a loamy sand, and a sand. They used both wind and radiation induced evaporation potentials. From the analysis of Philip and De Vries (1957), they calculated the relative contribution of vapor and liquid flow to the total water loss due to evaporation. From this analysis, they showed that the

vapor component is insignificant at the lower depths for a wind treatment, but became significant in the upper 5 cm. for the same wind treatment. For the radiation treatment, the vapor component was important for all depths except for early times. There was no attempt to fit their moisture content versus depth profiles although they must have had the necessary data to do so.

Ekern (1966) calculated the thermal vapor and liquid diffusivities from the analysis of Philip and De Vries (1957). He assumed that the theory was valid and used these calculations to get some idea of the moisture movement due to thermal gradients for typical temperature data in Hawaii.

Gee (1966) used both the analysis of Philip and De Vries (1957) and Taylor and Cary (1964) to analyze water flow in a closed soil sample with an imposed temperature gradient. He found that the analysis of Philip and De Vries modified to calculate the vapor moisture diffusivity according to a technique of Jackson (1964) underestimated the flow by an average factor of 2.7. The analysis of Taylor and Cary overestimated the flow by an average factor of 0.77.

Dirksen and Miller (1966) used the analysis of Philip and De Vries to calculate the movement of water due to thermal gradients for the case of freezing soils. They concluded that the analysis gave results in the right order of magnitude for the movement of water in soil subject to a thermal

gradient. They decided that the equations of Taylor and Cary were inadequate for analysis of transient flow because they assumed that steady state conditions were met.

Cary (1967) studied evaporation from a loam soil in the laboratory. He found that even under low wind velocities the vapor transfer in the dry layer at the surface of a soil is in part due to viscous air flow. The mass flow factor which is used in the analysis of Philip and De Vries (1957) is an attempt to bring this factor into the analysis. The equation that Philip and De Vries use for the mass flow factor was taken from steady state experiments, and they state that it may not be valid for transient flow.

The studies (Hanks, H. R. Gardner, and Fairbourn, 1967; Gee, 1966; Dirksen and Miller, 1966; and Cary, 1967) which were discussed briefly in the preceding paragraphs use the analysis of Philip and De Vries (1957). The different workers used various independent measurements to calculate the diffusivity functions which are needed to use the analysis of Philip and De Vries. Gee (1966), Dirksen and Miller (1966), and Cary (1967) calculated one experimental flow independent of the experimental flow data. Hanks, H. R. Gardner, and Fairbourn (1967), on the other hand, used all of their experimental data to calculate the flow due to vapor and liquid moisture. In our study, we use independent measurements to calculate the moisture content distribution and then use the

experimental moisture distribution to provide a check on the validity of the theory when it is applied to evaporation of soil water from columns of soil which are not at constant temperature.

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PROCEDURE

Soil Properties

Soil preparation

Webster silty clay loam soil, used for all experiments reported, was taken from the surface 30 cm. of a soil profile located in a grassed area at the Iowa State University Agronomy Farm located between Ames and Boone, Iowa. Surface organic residue was removed before the soil was collected. The collected soil was passed, while moist, through a 5 mm. screen so that large roots and clods could be removed. The soil was spread to dry, ground to pass through a 2 mm. sieve, mixed thoroughly, and stored at room temperature and humidity until needed for experiments.

Particle density

Particle density was determined using a pycnometer technique described by Blake (1965) except that the boiling of the soil suspension in the pycnometer to remove air was replaced by use of a 73 cm. of Hg vacuum and a vacuum chamber in which the pycnometers were placed.

Particle size distribution

Particle size distribution was determined using a sonic vibration dispersion technique described by Edwards and Bremner (1967) except that a Sonifier Cell Disruptor (Branson Sonic Power, Melville, Long Island, New York) was used as the sonic

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vibrator. A 10 g. sample of soil in 25 ml. of water was dispersed with sonic vibration for 15 min. instead of 30 min. as shown in the reference. This change was due to the greater power generated by the different vibrator.

Organic matter

Organic matter content of the soil was determined using a wet combustion technique described by Mobius (1960). Since this technique is used to measure organic carbon in the soil, the organic matter content was calculated by multiplying the organic carbon content by 1.724 as given by Allison (1965).

Soil-water retention curve

Soil water retention was determined for three ranges of matric potential. Each range was determined on a different piece of equipment. The high matric potential range (low tension range) was determined using Tempe Cells (Soil Moisture Equipment Company, Santa Barbara, California). A Tempe Cell is described by Reginato and Van Bavel (1962). The soil is retained by a 3 cm. tall and 5.3 cm. diameter brass cylinder. The porous plates used in the cells were tested and held at least 300 cm. of water tension. The lowest saturated conductivity of the plates used was 0.004 cm./hr. Soil was packed into three cells to a bulk density of 1.2 g./cm.³ and allowed to wet from the bottom. After wetting, the samples were allowed to stand in water so that a thin layer of water was on the surface of each sample for one week. Since Backer and

klute (1967) have shown that tension and pressure methods are not equivalent if the initial sample used for tension measurements is not water saturated, special care was taken to insure saturation of the sample. After wetting was completed, the samples were brought to equilibrium with the water level at the center of the brass cylinder and this level was used as the zero tension reference level. After approximately four days, the tension was increased to 10 cm. of water. Water outflow from the column was measured in a 1 ml. horizontal pipette which was calibrated in 0.01 ml. increments. The soil water was allowed to come to equilibrium with the applied tension range for approximately 24 hours before the tension was increased. In all cases, the reading of water outflow volume reached a maximum and then decreased. The volume used for the calculation of the water retention curve was the difference between the maximum volume reading and the reading at the time that the water tension had been increased for that tension step. Water retention at tensions of 10, 20, 30, 40, 50, 60, 75, 100, 125, 150, and 175 cm. of water was determined using this technique.

Soil-water retention in the intermediate range of matric potential was determined using a pressure cooker (Soil Moisture Equipment Company, Santa Barbara, California) with a ceramic porous plate. This equipment has been described by Richards (1965). Samples were run in duplicate and were packed in

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Gamma-ray Attenuation

Equipment

The gamma-ray moisture determination apparatus is the same as that used in a previous study (Kirkham, Rolston, and Fritton, 1967). It consists of a 251-millicurie cesium-137 source, a scintillation detector (Nuclear-Chicago Model 955) with a 2.5-cm. diameter and 2.5-cm.-thick thallium-activated NaI crystal with photomultiplier, and an analyzer-scaler (Nuclear-Chicago Model 8727). The analyzer was set to accept all pulses above 0.530 Mev. The pulse pair resolution (the matter of resolving time was described in the literature review) of the analyzer was 1.5 μ sec. The gamma beam is collimated at the source with a 5-mm. diameter hole in lead 12.5 cm. long. The beam must pass through a second 5-mm. diameter hole 12.5 cm. long before it reaches the detector. Scattered radiation is essentially eliminated using these collimators. The equipment meets requirements discussed by W. H. Gardner (1965) for the gamma-ray attenuation moisture measurement technique.

Resolving time and mass absorption coefficient

The counting rate was determined for a number of accurately measured thicknesses of material (lead, plastic, water, or soil) being used. Five 1-minute counts were averaged for the counting rate of lead and plastic. Four 1-minute counts were averaged for the counting rate of distilled water and soil.

The temperature of the distilled water was recorded and the value used to look up the density in standard tables. The soil used was oven dried at 109°C for 41 hours. The distilled water and the soil were placed in brass cylinders of 2.2-cm. radius 3.8-cm. length, and enclosed by plastic endplates. The various thicknesses of water and soil were obtained by placing several cylinders end to end. A maximum of five cylinders could be placed end to end and still fit between the source and detector. Volumes of the brass containers were calculated from the weight of water held by the containers and the density of the water at the measured temperature. Bulk density of soil was calculated by dividing the weight of oven-dry soil used to fill the container by the volume of the cylinder.

In another experiment, the mass absorption coefficient of water and soil was determined using the resolving time. Three replicates of four lengths of water and soil were used. Brass cylinders 4.6 cm. in diameter were used in this determination.

Water content using best mass absorption coefficient and resolving time

Brass cylinders of diameter 4.6 cm. and length 10.1 cm. were packed with soil previously mixed with water at five different water contents. Three replicate cylinders were measured. The wetted soil was packed into the cylinders with a circular piece of plastic on the end of a rod. The piece

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of plastic was the same diameter as the cylinders. Oven-dry contents (24 hrs. at 100°C) were compared with the water contents calculated from the gamma-ray intensities when corrected for resolving time.

Liquid Moisture Diffusivity

Liquid moisture diffusivity was determined using a technique described by Whisler, Klute, and Peters (1968). A soil container, constructed with Plexiglas, was 100 cm. long and 1.75 cm. thick. The width varied from 9.91 cm. to 9.97 cm. and was accurately measured at the locations where gamma-ray measurements were taken. A 0.8 cm. thick coarse porous glass bead plate was fastened to one end of the soil slab and water was introduced through the plate at time zero. The soil container was packed several times with oven dry soil. After each packing, the bulk density was measured with the gamma-ray attenuation technique. As uniform as possible a bulk density was obtained. At time equal to zero, water was pulled into the region behind the porous plate by a short application of vacuum and this water was allowed to start wetting the soil. The gamma beam was initially positioned at 10 cm. from the porous plate, and continuous 10 sec. readings were taken. After the wetting front moved past the 10 cm. location, the soil slab was moved to the next gamma beam location by turning a lathe-bed type screw. Data were taken at 10, 19, 25, and 35 cm. The experiment was performed in a constant temperature room, $27 \pm 2^\circ\text{C}$.

Moisture and Temperature Profiles During Evaporation

Soil columns

Two soil columns of rectangular cross section were constructed from 1 cm. thick Plexiglas. The column used for the radiation treatment was 10.75 cm. wide, 9.09 cm. long, and 19.55 cm. tall. The column used for the wind treatment was 10.86 cm. wide, 8.88 cm. long, and 19.65 cm. tall. A layer of aluminum foil was placed on the bottom and sides of the columns to reflect any radiation which tended to enter or leave the columns. Each column was insulated by placing a 2.5 cm. thickness of styrofoam on all sides outside of the aluminum foil. The column used for the radiation treatment was placed on a laboratory wooden bench on a 1 cm. thick asbestos board. The column used for the wind treatment was placed directly on a wooden table top. During the construction of the columns, holes were drilled in the side of the columns through which thermocouple wires could be placed. These holes were drilled so that four thermocouples could be located at the surface of the soil and at the 0.5 and 1.0 cm. depths; three thermocouples at the 2.0 and 3.0 cm. depths; and two thermocouples at one centimeter intervals from 4.0 to 10.0 cm. and 2 cm. intervals from 10.0 to 18.0 cm. The thermocouple wires were placed perpendicular to the direction that was used to take gamma-ray measurements from type 30-55-1 copper-constantan thermocouple no. 30 wire (Leeds

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and Northrup Co., Philadelphia, Pennsylvania). The thermocouple joints were insulated with Glyptal enamel (General Electric, Schenectady, New York). The 40 thermocouples used for each column were attached to a Amphenol type 57-30500 plug (Newark Electronics Corporation, Chicago, Illinois). This plug could be connected to a manual switching box when the temperature measurements were being taken. Another thermocouple wire ran from the plug to an ice bath which was used as a reference junction. During the gamma-ray measurements the plug could be disconnected so that the columns could be moved to the gamma-ray source. The manual switching box was constructed with the switches from a Decade Capacitor (General Radio Company, Concord, Massachusetts). Two lead wires ran from the switching box to a K-2 potentiometer (Leeds and Northrup Co., Philadelphia, Pennsylvania). A high sensitivity, 0.46 microvolts/mm., galvanometer (Catalogue No. 2430A, Leeds and Northrup Co., Philadelphia, Pennsylvania) was used with the potentiometer to detect current flow. The system for temperature measurement was capable of measuring 1.0 microvolt differences. This difference corresponds to a 0.025°C temperature difference. The temperatures were recorded to the nearest 0.025°C . This temperature sensitivity turned out to be more than adequate for our work.

The center of the gamma-ray beam, 5 mm. diameter, was located for sampling at depths of 0.55, 0.75, and 1.25 cm.;

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and at 1.0 cm. intervals from 1.55 to 18.55 cm. Three measurements were taken at each depth with the gamma-ray apparatus whenever data were taken. The measurements were taken 1.5, 4.5, and 7.5 cm. from the side of the column. Two JLC-Hatfield Heatflow Meter Disks (Joyce, Loebel, and Co., Ltd., Gateshead, England) were placed in the column. One was placed as near to the surface as possible and the other one was placed on the side of the column between the 1 and 2 cm. depths. The heat flow disks were 1 cm. in diameter and 2 mm. thick.

Preparatory procedures

Before holes were drilled in the columns, the volume of the columns was determined by filling the columns with distilled water. The volume of the columns was measured from the weight of water in the column and the density of the water at the recorded temperature. While the columns were filled with water, the thickness of the columns was determined in the direction of the gamma-ray measurements by determining the intensity of gamma radiation. The intensity with water, the intensity of the empty columns at the same location, the mass absorption coefficient of water, and the density of water were used to calculate the thickness of the columns.

Two 250 watt reflector heat lamps were placed approximately 60 cm. above the top of the column to be used in the radiation treatment. With the column removed and the asbestos

lying on the tabletop, a miniature net radiometer (Model MNR 601, C. W. Thornthwaite Associates, Centerton, New Jersey) was used at the height of the column top to check the uniformity of the radiation. The two heat lamps were adjusted in height and angle until a uniform area of radiation of approximately $1.5 \text{ cal./cm.}^2 \text{ min.}$ was found that covered an area larger than the surface area of the column. The heat lamps were not disturbed after a final check before which the heat lamps had been on for 16 hr.

The conditions for the wind treatment were produced by a table fan (Model 77648-TE, Emerson Electric, St. Louis, Missouri). The base of the fan was located approximately at the level of the column top and approximately 30 cm., horizontally from the column. The wind velocity immediately above the column was measured with an anemometer (Taylor Instrument Company, Rochester, New York). The anemometer has a 10 cm. diameter blade and was calibrated at the factory so that its dial read in feet of wind movement. Time needed to complete the wind velocity measurement was taken with a stopwatch.

The evaporation potential of the two treatments was determined by filling the columns with water and placing them in the positions which were to be used in the later experiment. Water loss was measured by the weight loss of the column after 15.33 hr. of evaporation. The results were converted to cm. of water loss per 24 hr.

Air dry soil was packed into the columns using a rod with a 3 cm. square piece of Plexiglas on the end. Thermocouple wires were positioned when the soil level reached the predetermined level of the thermocouple. The packing was continued above the top of the column for 10 cm. in an effort to insure uniform packing. This extra soil was then scraped off so that the surface of the column was level. Several samples were collected during the packing process on which the air dry moisture content was determined by drying in an oven at 110°C for 24 hr. After packing, the columns were placed in the gamma-ray beam and the intensity of gamma radiation was measured for air dry soil. Previous to these measurements, the intensity of the gamma radiation had been measured for the empty columns at each location that was to be used later for air dry intensity and wet intensity measurements. The columns were placed in exactly the same location on a stable platform for all gamma-ray measurements.

After the measurement of the intensity of gamma radiation through the air dry soil, a Plexiglas sheet with drilled holes for air escape and cut to the size of the column top was placed on each column. A small lead weight was placed on top of the Plexiglas sheet. Water was then introduced into the columns with a positive 50 cm. head with reference to the bottom of the columns. After about 1 hr., it became apparent that the small lead weight would not stop

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is swelling. The small lead weight was replaced with a 5 x 10 x 20 cm. lead brick. Wetting was continued until the soil surface was visually wet, approximately 6 hr. The columns were allowed to equilibrate overnight. The columns were again connected to the water source for 1 hr. the next day and for 6.5 hr. two days later. Some gamma-ray measurements were taken in between these wetting periods, and the whole process was taken in an attempt to get the columns as wet as possible. The last two wetting periods caused only small quantities of water to flow into the soil. The columns were again allowed to set overnight with the lead brick removed and the columns covered to prevent any evaporation. The surface of the columns appeared sopping wet the next morning when the initial measurements were taken for the experiment. Even with the precautions of keeping a weight on the soil surface, the soil swelled approximately 3 to 5 mm., with the greater swelling taking place where there were no thermocouple wires to hold the soil.

Periodic measurements

Measurements of column weight, net radiation, heat flow, and wind velocity were taken at various times during the experiment. The above measurements were taken each time the moisture and temperature distributions were determined except that the wind velocity was taken only four times during the experiment. The moisture and temperature distributions were

determined after approximately 1, 2, 3, 6, 9, 15, and 28 days of evaporation. The room temperature was $27 \pm 2^{\circ}\text{C}$ during the experiment. The relative humidity as measured by a sling psychrometer was 35 ± 7 percent during the experiment.

Final measurements

At the end of the experiment, the moisture distribution was determined gravimetrically by sectioning the column in 1 cm. intervals and drying the soil in an oven at 112°C for 24 hr. The sectioning was done immediately after the determination of moisture content with gamma-ray attenuation at the 28 day time. The soil obtained when the columns were sectioned was used after oven drying to determine the soluble salts contained in the soil. The quantity of soluble salts was determined so that a check of the magnitude of salt movement could be made and the effect of the salt movement on the gamma-ray attenuation readings and on the water movement could be examined. The quantity of soluble salts was determined with a technique described by Bower and Wilcox (1965). The extract was obtained by leaching approximately 60 g. of soil with 250 ml. of distilled water. The leaching was done in a Büchner funnel with a vacuum flask attached. Standard line vacuum was used to pull the water through the soil. The electrical conductivity in mmhos/cm. was multiplied by 640 as given by Bower and Wilcox to obtain mg. of salt per liter of solution and divided by the measured volume

of solution at the time of measurement to give mg. of salt. This value was then divided by the weight of oven dry soil to give mg. of soluble salt per g. of oven dry soil.

Numerical

Gamma-ray attenuation

Resolving time and mass absorption coefficient The

data obtained were used to calculate the slopes, u_p for lead and plastic, and u_a for water and soil, resulting from Equation 3 when $\ln R$ was plotted against $-x$ for lead and plastic and $-px$ for water and soil. When the data are plotted in this way, the intercept is $\ln R_0$. The slope was calculated by linear regression after a resolving time correction was made using Equation 1. Slopes were calculated for resolving times of 0.1 to 7.0 $\mu\text{sec.}$ in steps of 0.1 $\mu\text{sec.}$ The best value of the resolving time was determined by looking at the minimum weighted variance of the points from the line. A weighted variance was used as calculated from the equation

$$\text{weighted variance} = \frac{n}{\sum_{i=1}^n} (\text{deviations from regression})^2 / L$$

where n is the number of points used for the line. This equation weights the deviations from regression by the error ($\sqrt{I}$) expected in the counting rate.

The mass absorption coefficient was calculated from the slope of the line for the best resolving time (minimum weighted variance) and used as the true mass absorption coefficient. The computer program written for the IBM 360/65 computer is given in Appendix A.

Bulk density Gamma-ray intensity data taken in counts/min. for the intensity of the empty container and the container filled with air dry soil or soil at a known weight moisture content were corrected for resolving time using Equation 1. The corrected data were then used in Equation 13 to calculate the bulk density of the soil on an oven dry basis. A computer program for the calculation of soil bulk density is given in Appendix B.

Moisture content Gamma-ray intensity data taken in counts/min. for the intensity of the empty container, the container filled with soil at a known initial weight moisture content, and the container filled with soil at an unknown moisture content were corrected for resolving time using Equation 1. The corrected data were then used in Equation 16 to calculate the volumetric moisture content of the soil. A computer program for the calculation of volumetric moisture content is given in Appendix C.

Liquid moisture diffusivity

Data taken at a single location on the soil slab used for horizontal infiltration of water were plotted as intensity in counts per 10 sec. versus time in minutes. A

smooth curve was drawn through the points and points were read off of the smooth curve. These points were read into the computer where they were converted to counts per minute, corrected for resolving time using Equation 1 and were then used by the computer to calculate and read out moisture content. The values of moisture content were then used in Equation 23 to calculate the liquid water diffusivity in units of cm^2/min . where the differential and integral terms in Equation 23 were replaced by finite differences and numerical integration with the following substitutions:

$$\frac{dt}{d\theta} \approx \frac{t_{j+1} - t_j}{\theta_{j+1} - \theta_j}$$

and

$$\int_{\theta_d}^{\theta} 1/t^{1/2} d\theta \approx \sum_{n=1}^j \left[\frac{(1/t_{j+1}^{1/2} + 1/t_{j+1}^{1/2})}{2} (\theta_{j+1} - \theta_j) \right]_n$$

where j is the time index. The computer program which guided the above operations is shown in Appendix D.

Solution of the isothermal diffusion equation

Values of volumetric moisture content, pressure, summation of the diffusivity as given in the numerator of Equation 33, and initial moisture and pressure distributions were read into the computer which used Equation 31 to calculate

pressure distributions in the soil at various values of time. The pressure distributions were converted to moisture content distributions using the soil-water retention curve. The evaporation rate and cumulative evaporation were calculated from the decrease in moisture content summed over all depths. Each value of the moisture content in the moisture distribution was assumed true for 0.5 cm. above and below the depth for which the moisture content was calculated for the summation over all depths. The bottom boundary condition was met by setting the hydraulic conductivity to zero at the depth of 19.5 cm. Delta time in Equation 31 was given an initial value of 0.01 hr. At the start of the theoretical redistribution, sometimes the column would wet up for the first few values of delta time due to the value of the initial moisture content at the surface of the column which was used for the boundary condition in Equation 25. When the soil wetting occurred, delta time was given a value twice its previous value. Otherwise delta time was calculated as the time needed to give 0.1 cm. of evaporation based on the previous time's evaporation rate. If the delta time was greater than 24 hr., it was redefined as a constant of 22 hr. to match the theoretical interval as closely as possible with the experimental. When the top boundary condition was changed from Equation 25 to Equation 26, at the time that the evaporation rate from the soil column first dropped lower than the evaporation potential,

delta time was set to 0.02 hr. This change in delta time was needed because the evaporation rate increased after the change in boundary condition was performed. For a period after this change, the evaporation rate still exceeded the evaporation potential. When this occurred, the cumulative time was advanced to maintain the constant evaporation rate until the evaporation again dropped to a value lower than the evaporation potential. The program which was used is given in Appendix E.

Calculated moisture profiles from heat and mass transfer

The moisture profile for the 28 day time was calculated using Equation 49. The functions used in the equation were determined at values of the temperature which were measured at the depth at which the equation was being determined. The quantities needed in Equation 49 were determined as follows:

Liquid moisture diffusivity Values for the liquid moisture diffusivity were taken from the smoothed values divided by 35.0 which were determined experimentally at 27°C and multiplied by the ratio of the viscosity at the soil temperature to the viscosity at 27°C to correct for temperature.

Hydraulic conductivity Hydraulic conductivity was calculated by multiplying the liquid moisture diffusivity by the slope of the moisture retention curve (Equation 41).

Soil-water tension was corrected for temperature with the equation

$$\psi(T) = \psi(27) / [1 - (30.0) (\gamma) (T - 27)] \quad (50)$$

where the symbols were defined under Equations 34 and 39 and the temperature coefficient of matric potential is assumed equal to 30.0 times the temperature coefficient of surface tension. The factor of 30.0 was obtained by fitting data in Gardner (1955) and Taylor, Evans, and Kemper (1961).

Thermal liquid moisture diffusivity Thermal liquid moisture diffusivity was calculated with Equation 40 where each term was calculated at the temperature of the soil at the depth being considered.

Vapor moisture diffusivity Vapor moisture diffusivity was calculated with Equation 36 where each term was calculated at the temperature of the soil at the depth being considered.

Thermal conductivity Thermal conductivity which includes the effect of heat transfer due to vapor movement was calculated at the temperature of the soil at the depth being considered with the technique described by De Vries (1963) where the assumed decrease in the thermal conductivity of water vapor filled pores was calculated as suggested by Cochran, Boersma, and Youngberg (1967). The value of ϵ_a (defined in De Vries, 1963) used for these calculations was 0.125. A value for the thermal conductivity of the dry soil

was taken from Jackson and Kirkham (1958) for their O moisture content treatment. The Colo soil which they used is in the same soil association as the Webster and it was assumed that the basic soil minerals of the two soils would be the same. The value of D_{atm} was taken from De Vries and Kruger (1967).

Thermal vapor diffusivity Thermal vapor diffusivity was calculated with Equation 35 where each term was calculated at the temperature of the soil at the depth being considered.

Thermal moisture diffusivity The thermal moisture diffusivity was calculated with Equation 43.

Moisture diffusivity The moisture diffusivity was calculated with Equation 44.

Evaporation rate The evaporation rate was calculated from the slope of the cumulative evaporation curve for the column with the radiation treatment at the 28 day time.

Soil heat flux density The soil heat flux density was determined by calculating the heat needed to evaporate the amount of water used to give the evaporation rate given in the preceding step. The amount of heat used to heat the soil in the column was assumed to be negligible, and the amount of heat that flowed out of the column was assumed not to affect the vertical moisture profile.

The program which was used for the calculation of moisture profiles from heat and mass transfer is shown in Appendix F.

RESULTS

Soil Properties

Table 1 shows the magnitude of several quantities which were measured to help characterize the soil used in the experiments reported in this dissertation. Values are given for the particle density of the soil, percentage sand, silt, and clay, and percentage organic carbon. In addition the results of some simple calculations are given to make the measured values more meaningful. Total porosity (53.45 and 50.20 percent) is calculated for two values of bulk density (1.294 and 1.200 g./cm.³) which are used frequently. The 3.25 percent change in total porosity for a 0.094 g./cm.³ change in bulk density will be used in the discussion. The value for organic carbon is converted to the more familiar value for organic matter.

Figure 1 shows the soil-water retention curve which was determined experimentally for a Webster silty clay loam soil with a bulk density of 1.2 g./cm.³. The three ranges of tension used are identified by three different symbols. The total porosity (53.45 percent) is shown at the tension of 1 millibar. All values of moisture content which are plotted at 1 millibar are actual values for the moisture content when the tension was 0 at the reference point in the center of the brass Tompe Cell cylinder. Values of the third replication for the Tompe Cell have been ignored in drawing

Table 1. Physical properties for a selected Webster silty clay loam soil

Soil Property	Replication						Average
	1	2	3	4	5	6	
Particle density (g./cm. ³)	2.5803	2.5839	2.5800	2.5801	2.5684	2.5784	2.5779
Total porosity ³ (1.2 g./cm. ³ bulk density)	-	-	-	-	-	-	53.45%
Total porosity ³ (1.294 g./cm. ³ bulk density)	-	-	-	-	-	-	50.20%
Percentage sand	19.66	19.78	20.10	-	-	-	19.85
Percentage silt	47.67	48.04	48.28	-	-	-	48.00
Percentage clay	32.67	32.18	31.62	-	-	-	32.16
Percentage organic carbon	4.00	3.71	3.89	3.86	3.78	-	3.85
Percentage organic matter	-	-	-	-	-	-	6.64

Figure 1. Soil-water retention curve for a Webster silty clay loam soil

the smooth curve because they did not start as close to saturation as did the other two replications and the curve is considerably below the other two curves. The moisture tension curve, as determined experimentally, was then extrapolated to zero moisture content. The extrapolated curve is not shown below the 10 percent level of moisture content, but it is the extension of the lower end of the curve to 0 percent water content at 10,000,000 mb. tension.

Gamma-ray Attenuation

Table 2 shows a typical set of data used to determine the best resolving time and best mass absorption coefficient from the minimum-weighted variance. A minimum-weighted variance existed for all materials used in this study. Five thicknesses of soil and water and the corresponding five average gamma-ray intensities were used to calculate the data in this table. The thicknesses ranged from 3.83 cm. to 19.16 cm. The intensities ranged from 1,339,921-266,641 counts/minute for soil and 1,453,880-422,355 counts/minute for water. The change in the value of the mass absorption coefficient for both water and soil with a change in resolving time can be determined by using the data in this table. The mass absorption coefficient of water with no correction is $0.08105 \text{ cm.}^2/\text{g.}$ and for soil is $0.07309 \text{ cm.}^2/\text{g.}$

Table 3 shows the resolving times determined by finding the minimum-weighted variance for various materials and ranges

Table 2. An example of data used to determine the best resolving time and best mass absorption coefficient for soil and water by finding the minimum weighted variance

Resolving time μsec.	Mass absorption coefficient		Weighted Variance	
	Water	Soil	Water	Soil
0.1	0.08116	0.07381	157.7	221.2
0.5	0.08162	0.07350	137.8	191.1
1.0	0.08219	0.07391	113.6	155.9
2.0	0.08336	0.07474	70.3	93.6
3.0	0.08457	0.07559	35.3	44.8
3.8	0.08557	0.07629	15.1	18.3
3.9	0.08569	0.07639	13.2	16.1
4.0	0.08582	0.07647	11.4	13.8
4.1	0.08595	0.07656	9.8	11.8
4.2	0.08608	0.07665	8.3	10.0
4.8	0.08685	0.07719	2.9	4.5
4.9	0.08698	0.07728	2.6	4.4
5.0	0.08711	0.07738	2.5	4.6
5.1	0.08725	0.07747	2.6	5.1
5.2	0.08738	0.07756	2.9	5.9
6.0	0.08845	0.07830	13.1	23.1
7.0	0.08984	0.07926	48.8	75.7

55

Table 3. Values of the resolving time and slope when the weighted variance is a minimum for several materials

Line	Material	Resolving time (μ sec.)	Slope ($\mu_a \rho$) (1/cm.)	Minimum weighted variance	Range of Counts (counts/minute)	Number of thicknesses of material
1	lead	4.1	1.2031	8.68	932,979-143,340	16
2	"	1.0	1.1628	109.84	2,054,352-260,541	10
3	"	1.0	1.1616	113.68	2,053,847-260,982	10
4	"	4.0	1.2006	3.59	1,303,660-260,982	9
5	plastic	4.6	0.0983	1.74	855,162-449,455	11
6	"	6.6	0.1042	7.48	2,056,020-1,041,920	12
7	"	5.6	0.1009	14.86	2,055,000-259,442	10
8	"	5.2	0.1004	2.44	1,672,964-259,442	9
			Slope (μ) (cm. ² /g.)			
9	water	5.0	0.0871	2.50	1,453,880-422,355	5
10	soil	4.9	0.0773	4.42	1,339,921-266,641	5

of counting rates. Data used for each line of Table 3 were independent except that data for line 2 were a combination of selected data from line 1 plus independent data that extended the range of counts of line 1. Data for line 4 are the same as line 3 and data for line 8 are the same as line 7 except that the intensity through air has been removed in each case.

Table 4 shows the mass absorption coefficient of soil and water for three replicates each of four lengths of column corrected for a resolving time of 5.0 μ sec. The data in Table 4 are independent of data used for Table 3.

Table 5 shows the water content of a soil with and without a resolving time correction. For water contents determined with corrected gamma-ray attenuation, the mass absorption coefficients and data were corrected for a resolving time of 5.0 μ sec. For water contents determined with uncorrected gamma-ray attenuation, no resolving time correction was used for the counting rates or the mass absorption coefficients.

Liquid Moisture Diffusivity

Table 6 shows the bulk density of the soil slab used to measure liquid moisture diffusivity. The distance from the porous plate was a horizontal distance. The variable width of the soil slab was taken into account when the bulk density was calculated.

Figure 2 shows a plot of the volumetric moisture content versus the Boltzmann transformation function, $\lambda = x/t^{1/2}$.

Table 4. Values of the mass absorption coefficient corrected for a resolving time of 5.0 μ sec. for four column lengths replicated three times

Column length	Mass absorption coefficient ($\text{cm}^2/\text{g}.$)	
	Water	Soil
5.04 cm.	0.08656	0.07802
"	0.08667	0.08028
"	0.08641	0.07711
10.13	0.08671	0.07847
"	0.08665	0.07921
"	0.08681	0.07871
15.18	0.08662	0.07833
"	0.08680	0.07914
"	0.08690	0.07851
20.26	0.08673	0.07857
"	0.08662	0.07875
"	0.08672	0.07882
Average	0.08668	0.07886

Table 5. Comparison of gravimetric determinations of water content by weight with gamma-ray attenuation determinations of water content by weight when no resolving time correction has been made and when a 5.0 μ sec. resolving time correction was made

Percentage water content by weight	Gamma-ray		Water content difference (gravimetric-gamma-ray)	
	Corrected	Uncorrected	Corrected	Uncorrected
Gravimetric				
7.68	8.46	5.22	0.78	-2.46
7.80	9.24	5.99	1.44	-1.81
7.67	8.55	5.35	0.88	-2.32

Table 5. Continued

Grav- metric	Percentage water content by weight		Water content difference (gravimetric - gamma-ray)	
	Gamma-ray		Corrected	Uncorrected
	Corrected	Uncorrected		
13.10	12.46	9.20	-0.64	-3.90
12.92	12.38	9.15	-0.54	-3.77
13.04	12.59	9.34	-0.45	-3.70
18.25	17.92	14.61	-0.33	-3.64
18.35	17.92	14.65	-0.43	-3.70
18.40	17.66	14.36	-0.74	-4.04
30.48	30.20	28.15	-0.28	-2.33
30.40	30.28	28.26	-0.12	-2.14
30.18	29.31	27.36	-0.87	-2.82
35.74	35.49	33.13	-0.25	-2.61
37.06	37.01	34.53	-0.05	-2.53
36.04	34.86	32.47	-1.18	-3.57

In order to use the Boltzmann transformation, a plot of λ versus moisture content θ should be the same (the curves should superpose) for each horizontal distance from the porous plate (wetting face) as the moisture content changes and the time increases for that position. The various locations 10, 19, 25, and 35 cm. where measurements were taken are shown by different symbols as indicated on the graph. The points for the plot at the greater distances stop at water contents which are lower than the water contents of the 10 cm. distance because not as great a value of water content was obtained at the greater distances.

Table 6. Width and bulk density at various locations in a horizontal soil slab used in the determination of liquid water diffusivity

Distance from porous plate (cm.)	Bulk density g./cm. ³	Width cm.
0.5	1.2871	9.96
2.0	1.3105	9.97
5.0	1.3189	9.96
10.0	1.3177	9.95
15.0	1.2588	9.93
19.0	1.3096	9.93
25.0	1.2572	9.91
29.0	1.3156	9.91
35.0	1.2682	9.93
40.0	1.2927	9.94
44.0	1.2808	9.95
50.0	1.2490	9.94
55.0	1.2453	9.93
60.0	1.2915	9.94

Figure 3 shows the values of the liquid moisture diffusivity as determined from the experiment. The values determined from the measurements at the various locations 10, 19, 25, and 35 cm. are given with the symbols indicated on the graph. The smooth curve was extended at the right to higher moisture contents so that the liquid moisture diffusivity would have values up to saturation for use in the numerical analysis of the experimental results.

Figure 2. Data of volumetric moisture content versus the Boltzmann transformation function, $x/t^{1/2}$, for four locations which were used to determine liquid moisture diffusivity

Figure 3. Liquid moisture diffusivity plotted as a function of volumetric moisture content for four locations in a horizontal soil slab

Moisture and Temperature Profiles During Evaporation

Wind treatment

In Figure 4 the initial values of bulk density and water content plotted on a sensitive scale for three sampling locations at various depths in the soil columns used for this experiment are shown. The letters L, C, and R on the curves indicate whether the sampling location was on the left, center, or right side of the column. Bulk density measurements were taken immediately after packing and before any wetting of the soil. Initial water content measurements were taken immediately before the soil column was placed under the fan to start the experiment. After the wetting period, the bulk density values are only relatively correct because the soil swelled during wetting. These bulk density values were used to calculate the water contents after wetting, and thus the water contents determined with the technique of gamma-ray attenuation after soil wetting are subject to errors due to the inadequacy of the values of bulk density. Values of the average bulk density and average moisture content are given on the graph to indicate the overall average as calculated from the gamma-ray data and the value that was calculated from the weight and volume of the column. Swelling of the soil surface was taken into account in calculating the average moisture content.

Figure 5 shows the distribution of soil temperature at times of 0.00, 21.67, 66.25, 206.92, and 676.92 hrs. for curves

Figure 4. Initial values of bulk density and moisture content for the wind treated soil as a function of depth

Figure 5. Initial and subsequent average values of temperature for the wind treated soil plotted as a function of depth for several times

0, 1, 2, 3, and 4 respectively. The lowering of the temperature for curves 1 and 2 was due to the cooling effect of evaporation while the increase in temperature shown in curves 3 and 4 was due to changes in the room temperature. The temperature profiles shown correspond to the water content profiles which will be shown later.

Figure 6 shows the curve of cumulative evaporation for the wind treatment. The points plotted in the graph are experimental points calculated from the weight of the soil column at various times. The smooth curve is the curve drawn through points calculated using the numerical solution to the isothermal diffusion equation (Equation 24) using the technique of Hanks and H. R. Gardner (1965) where θ_0 was $0.0667 \text{ cm.}^3/\text{cm.}^3$, $h(\theta_0)$ was $-980,000 \text{ mb.}$, and EP was 0.1 cm./hr. The values of other quantities which can be used to interpret the cumulative evaporation curve are given on the graph. The measurement of these values is explained in the procedure. In order to obtain the fit of the experimental points, the values of the liquid moisture diffusivity, $D_{\theta, \text{liq}}$ shown in Figure 3 had to be divided by a factor of 35.0. With this factor, a very good fit of calculated versus experimental data for the full month's data was obtained. W. R. Gardner (1959) obtained a similar well fitting curve for a similar type experiment for a different soil. Gardner did not have to use a factor. The factor, 35.0, will be considered in the discussion.

Figure 6. Experimental and theoretical cumulative evaporation of water from the wind treated soil

Figure 7 shows the fit of the experimental moisture content distributions with those calculated from the diffusivity equation using the adjusted diffusivity function, $D_{e,liq}$, just described for various times (0.00, 21.67, 66.25, 206.92, and 676.92 hrs.) of evaporation as shown. The curves in Figure 7 are labeled 0, 1, 2, 3, and 4. Each number corresponds to a gamma-ray determined curve indicated by the line connecting the points and to a theoretical curve indicated by the heavy smooth line except curve 0 which is the same as the experimental curve. Also each curve number corresponds to the time (21.67 hrs. for curve 1, 66.25 hrs. for curve 2, etc.) at which data were taken.

Radiation treatment

Figure 8 shows the initial distribution of bulk density and moisture content at three sampling locations for the radiation treated soil column. The explanations given for Figure 4 apply also to Figure 8.

Figure 9 shows the distribution of temperature with depth for the radiation treated soil column for times 0.00, 20.58, 64.25, 206.10, and 657.32 hrs. for curves numbered 0, 1, 2, 3, and 4 respectively. The temperatures are averages of all thermocouple temperatures taken at the depth indicated. The variation in temperature at any one depth was frequently 5°C for the 0.0 and 0.5 cm. depths and usually less than 1°C for all other depths. During the two hour sampling period for

Figure 7. Experimental and theoretical curves of moisture content versus depth for several times for the wind treated soil

Figure 8. Initial values of bulk density and moisture content for the radiation treated soil

Figure 9. Initial and subsequent average values of temperature for the radiation treated soil plotted as a function of depth for several times

gamma-ray measurements when the column was removed from the heat lamps, the temperatures at the surface dropped from about 80°C to approximately 40°C, the temperatures in the lower part of the soil column dropped from about 50°C to approximately 43°C, and the temperatures in the middle layers of the column dropped from about 55°C to approximately 49°C.

Figure 10 shows the experimental points of cumulative evaporation for the radiation treated soil column, and the curve of cumulative evaporation calculated from the numerical solution of the diffusion equation where θ_0 was 0.02 cm.³/cm.³, $h(\theta_0)$ was -5,000,000 mb. and EP was 0.154 cm./hr. The liquid moisture diffusivity which was determined experimentally had to be divided by a factor of 10.0 to fit this cumulative evaporation curve. (For the different evaporation treatments, wind and radiation, different linear factors had to be used to change the liquid moisture diffusivity so that the theory would fit the experimental cumulative evaporation curves. The difference in the factors used will be covered in the discussion, but is mainly due to the difference in the temperature of the two soil columns.) The other values given on the graph are determined according to the techniques listed in the procedure.

Figure 11 shows experimental and theoretical profiles for times of 0.00, 20.58, 64.25, 206.10, and 657.32 hrs. of evaporation for curves 0, 1, 2, 3, and 4 respectively for the radiation treated soil column. Each number corresponds

Figure 10. Experimental and theoretical cumulative evaporation of water from the radiation treated soil

Figure 11. Experimental and theoretical curves of moisture content versus depth for several times for the radiation treated soil where the numbers 0, 1, 2, 3, or 4 indicate comparable curves

to an experimental and a theoretical curve. Each point on the experimental curve determined with gamma-ray attenuation is the average of three sampling locations. The points on the gravimetrically determined experimental curve are for a 1 cm. thick section of the column and were determined only at the end of the experiment. The smooth curves are calculated using the isothermal diffusion equation with the liquid moisture diffusivity divided by the factor 10.0. The initial moisture distribution was the same for the experimental and numerical procedures.

Table 7 shows the values of the salt content of several depths in the two soil columns which were used for wind and radiation evaporation treatments. These values were taken at the end of the experiment. The values have been calculated from values of electrical conductivity as was described in the procedure.

Table 7. Soluble salt content of soil at several depths after approximately 28 days evaporation under wind and radiation treatments

Treatment	Depth (cm.)	Salt content (mg. of salt/g. of dry soil)
Wind	0-1	1.36
"	16-17	1.03
Radiation	0-1	1.47
"	1-2	1.43
"	3-4	1.12
"	10-11	1.00
"	17-18	1.06

Figure 12 shows the values of heat flow measured in the soil column for the radiation treatment. Most quantities are negative because heat flowing out of the column was taken as positive for the De Vries and Philip analysis. The values of net radiation were adjusted to the values which would have been measured at the surface of the soil column. This was done by measuring beforehand the variation of net radiation with height above the asbestos surface after 16 hours of radiation. A value of the thermal conductivity of the insulating plastic wall of the column plus the styrofoam was calculated from the value of heat flow at the last point given for heat flow through the wall of the column. This value was $0.0059 \text{ cal./cm.}^2 \text{ min./}^\circ\text{C/cm.}$ This value seems high for an insulating material. The temperature gradient could only be estimated and thus there is probably considerable error in this calculation. Heat flow into the column was calculated for the last time shown on the figure and was 16.62 cal./min. At the same time heat flow out of the sides of the column was 42.6 cal./min. When the value of net radiation is used to calculate the heat flow into the column, heat flow into the column is 118.6 cal./min. The three measurements are not consistent and cannot be used for any purpose other than comparison of each quantity as it varies with time.

Figure 12. Values of heat flow and net radiation for the radiation treated soil for several times

Moisture profile from heat and mass transfer

Figure 13 shows the liquid and vapor moisture diffusivities as a function of volumetric water content for temperatures of 81.34, 51.23, and 20.00°C. The functions were calculated as described in the numerical procedure. The total moisture diffusivity is the sum of the two functions shown here. It is nearly equal to the diffusivity function that is greatest at the moisture content in question.

Figure 14 shows the hydraulic conductivity as a function of volumetric moisture content for temperatures of 81.34, 51.23, and 20.00°C. This function was calculated from the liquid moisture diffusivity shown in the previous graph and the slope of the temperature corrected moisture retention curve shown in Figure 1.

Figure 15 shows the thermal moisture diffusivity for liquid and vapor movement for temperatures of 81.34, 51.23, and 20.00°C. In both cases the curves were calculated as described in the numerical procedure.

Figure 16 shows the results of the calculation of thermal conductivity described in the numerical procedure for temperatures of 81.34, 51.23, and 20.00°C.

Figure 17 shows the experimental and theoretical moisture profiles at the 657.32 hr. time for the radiation treated soil column. The experimental curves are identical to the experimental curves in Figure 11 for the 657.32 hr. time. The

Figure 13. Liquid and vapor moisture diffusivity calculated for three temperatures as functions of moisture content

Figure 14. Hydraulic conductivity calculated for three temperatures as a function of moisture content

Figure 15. Thermal liquid and vapor moisture diffusivity calculated for three temperatures as functions of moisture content

Figure 16. Thermal conductivity calculated for three temperatures as a function of moisture content

Figure 17. Experimental and theoretical curves of moisture content versus depth for the radiation treated soil for the 657.32 hr. time where the experimental curves are the same as those numbered 4 in Figure 11

curve for the theory of heat and mass transfer was calculated using the integral (Equation 49) of Philip (1957) and the functions as described in Figures 13-16. The values of the evaporation and heat flux calculated as described in the procedure were: $E = 0.8072 \times 10^{-6}$ cm./sec. and $Q = 0.453 \times 10^{-3}$ cal./cm.² sec. The surface moisture content was given a value of 2 percent.

DISCUSSION

Soil Properties

In the Philip and De Vries (1957) theory, temperature was assumed to be between 20 and 30°C. In the extension of the theory to higher temperatures, the quantities which were used to calculate the functions in Equation 49 were varied with temperature. The effect of temperature and several other factors on the soil-water retention curve can be shown by reference to studies in the literature.

R. Gardner (1955) and Taylor, Evans, and Kemper (1961) give results of studies where soil at a constant water content was sealed into a beaker and brought to equilibrium with several temperatures. The soil-water tension was recorded at equilibrium in these experiments and the data were presented as a graph of soil-water tension versus temperature for several values of moisture content. Equation 50 which is the equation of a straight line for soil-water tension plotted against temperature with a slope of $30.0 \gamma\psi$, was found to roughly fit the data in R. Gardner (1955) and Taylor, Evans, and Kemper (1961) and was used to adjust the soil-water retention curve to the temperature of the soil at the depth in question. The slope needed to fit the data required a factor of 30.0 to be multiplied by the slope which was given in Philip and De Vries

(1957). It is assumed that this same slope can be used for moisture tensions outside of the range of the tensiometer. Taylor, Evans, and Kemper (1961) give other data which suggest that the factor may be considerably greater for lower moisture contents.

Box and Taylor (1962) showed that at any given temperature and for tension greater than 0.3 bars there was little difference in the value of volumetric moisture content as the bulk density was varied. At 25°C and 0.3 bars of tension, a change of bulk density from 1.10 to 1.65 g./cm.³ resulted in only a 1 percent change in water content. For tensions near 0, for which Box and Taylor show no data, a change in bulk density would give a change in moisture content equal to the change that simultaneously results in the total porosity. Thus for a 0.1 g./cm.³ change in bulk density, the moisture content would change 3.25 percent, the value mentioned in the results. A change in bulk density would have little effect on the soil-water retention curve for the tensions encountered in this study.

Swartzendruber (1968) discussed the use of the static moisture tension curve to analyze a dynamic flow system. The moisture tension curve shown in Figure 1 was determined under static conditions and is being applied to a dynamic flow system. Swartzendruber cited several references where it had been shown that the moisture tension relationship is different

for static conditions and for transient flow conditions. The differences between static and dynamic soil-water retention will be ignored for this study.

King and Schleusener (1961) hypothesized that hysteresis was the cause of an effect which they observed for water evaporation from sand with a water table at a certain depth. The effect was the decrease in evaporation rate from the sand when the evaporation potential was increased. Since their experiments were for the steady state condition, it is possible that hysteresis could cause the stated effect. In our case, the experimental conditions cause the soil to dry at a fast rate and it is improbable that the soil ever wetted enough in our experiments to cause hysteresis.

Gamma-ray Attenuation

From Table 2, we see that the weighted variance is a minimum at a certain resolving time. This minimum weighted variance indicates that a correction for the best resolving time results in a better fit of the $\ln I/I_0$ versus ρx or x data (see Equations 1 and 3) with a straight line than data with no correction. The procedure for calculating the data in Table 2 involved the assumption that Equation 1 is a valid correction equation for the resolving time. Several other equations used to correct for resolving time are given in Kohman (1949). Equation 1 seems the most acceptable and

is the only one cited in Overman and Clark (1960). Kohman also gives three methods of measuring resolving time: paired sources, multiple paired sources, and proportional sources. These methods are not directly applicable to the case of one source, such as the gamma-ray attenuation apparatus.

From Table 2, the mass absorption coefficient of water with a 0.1- μ sec. resolving time correction is 0.08116 cm^2/g . The value of the mass absorption coefficient increases as the resolving time increases. At a resolving time of 3.9 μ sec., the mass absorption coefficient is identical to the theoretical mass absorption coefficient for water of 0.0857 cm^2/g . (Grodstein, 1957). The value of the mass absorption coefficient increases above 0.0857 cm^2/g . (0.08711 cm^2/g .) for the best resolving time of 5.0 μ sec.. The mass absorption coefficient for the best resolving time is much closer to the theoretical value than the mass absorption coefficient with no correction, + 0.0014 versus - 0.0045 cm^2/g . Evidently 1.1 μ sec. of the best resolving time was due to errors that caused a non-linearity other than resolving time.

From Table 3 both material and range of counting rate seem to influence the best resolving time as determined by the method described in this report. Materials change the best resolving time as can be seen from the last value for lead (line 4) of 4.0 μ sec. and the value for soil (line 10) of 4.9 μ sec. These two values are relatively free of the

variation due to range of counting rate and their weighted variances are relatively small. The variation in the resolving time due to the absorbing material seems greatest when a metal (e.g., lead) is compared with plastic, water, and soil. Some effect other than resolving time is probably causing this difference.

There is an effect of range of counting rate on the best resolving time in both the data for lead and those for plastic. In the data for lead, the best resolving time was $1.0 \mu\text{sec.}$ (lines 2 and 3) when an intensity resulting from zero thickness of lead was included in the data. When the intensity for zero thickness of lead was removed, the best value for the resolving time was $4.0 \mu\text{sec.}$ (line 4). Since the intensity for zero lead thickness was some 700,000 counts above the next largest intensity, similar data were taken for plastic where the next largest intensity was closer to the intensity through air. In the data for plastic, two effects are noted. First, a lower intensity range of counting rate gave a lower resolving time, $4.6 \mu\text{sec.}$, than a higher intensity range of counting rates, for which the resolving time was $6.6 \mu\text{sec.}$ Data covering the full range gave a resolving time of $5.6 \mu\text{sec.}$ Second, when the intensity for zero thickness of plastic was removed, the resolving time dropped from $5.6 \mu\text{sec.}$ to $5.2 \mu\text{sec.}$ This drop in resolving time was accompanied by a drop in the weighted variance. This indicates that the intensity

for the zero thickness of plastic was not on the same straight line as the rest of the data. Higher counting rates evidently contain either a greater value of resolving time or other sources of non-linearity than lower counting rates. When the technique was used to determine the best resolving time for water and soil, no zero thickness of soil or water was included in the determinations. If the higher range of counting rates have a higher resolving time, the resolving time of water and soil for the minimum weighted variance may contain some effect of averaging the resolving time for the high and low ranges of counting rate.

In Table 4, the mass absorption coefficient of water is almost constant for all lengths and replicates and is $0.0004 \text{ cm.}^2/\text{g.}$ lower than the value in Table 2. This difference would correspond to a $0.2 \text{ } \mu\text{sec.}$ error in the resolving time and probably can be considered insignificant. The constant mass absorption coefficient for several lengths of water shows that a $5.0 \text{ } \mu\text{sec.}$ resolving time is adequate for the range of counting rates normally encountered.

The average mass absorption coefficient of soil from Table 4 is $0.07866 \text{ cm.}^2/\text{g.}$ This is $0.00138 \text{ cm.}^2/\text{g.}$ higher than the value $0.07728 \text{ cm.}^2/\text{g.}$ obtained in Table 2. In this instance, there is one value (0.07711) in Table 4 lower than the value 0.07728 in Table 2, and the variation in the mass absorption coefficient for soil is much larger than for water.

We would assume that the variation in the mass absorption coefficient of soil is due to the small volume actually sampled by the gamma beam. The average value of the mass absorption coefficient for soil in Table 4 would correspond to a 1.0- μ sec. change in the resolving time as determined from Table 2. If the average value in Table 4 is considered correct, then the value of $0.0775 \text{ cm.}^2/\text{g.}$, which would correspond to theory, would occur in the vicinity of 4.1 μ sec. Evidently 0.8 μ sec. of the resolving time is due to other factors than resolving time that cause a non-linearity in the plot of $\ln I$ versus $-x$ or $-\rho x$.

Table 5 shows the importance of a resolving time correction. When water contents calculated from gamma-ray data are compared with gravimetric samples, the water contents corrected for a 5.0- μ sec. resolving time have an average deviation of 0.60% water content but the data uncorrected for resolving time have an average deviation of 3.02% water content. The deviation for the uncorrected values is somewhat biased because the intensity ratio was of wet soil to plastic rather than of wet soil to dry soil. The latter would probably have caused less error because of a lesser resolving time correction. When a resolving time of 4.0 μ sec. is used to correct the gamma-ray data, the water contents are roughly 0.8% less than the water contents corrected for a resolving time of 5.0 μ sec. Thus, whatever causes the

resolving time of plastic, soil, and water to be higher than that of lead is also causing some of the difference in water content. This means that the method described in this study will correct for resolving time and one or more unknown factors that cause the plot of $\ln I/I_0$ versus $-px$ or $-x$ to be nonlinear. With the correction made for resolving time as in Table 5, the gamma-ray attenuation method should give the moisture content for a constant bulk density over a range from 7.67 to 37.06 percent by weight to within ± 1 percent.

Liquid Moisture Diffusivity

Of all the bulk densities recorded in Table 6 for distances from the wetting source of less than 50 cm., only the three bulk densities located at 15, 25, and 35 cm. are at a significantly lower value than the others. The bulk densities for distances of 50 cm. or greater would not have changed the diffusivity relationship because flow did not occur in that region of the column. It is not known what effect the three low density points at 15, 25, and 35 cm. had on the diffusivity function.

From Figure 2, the same curve is not given for the 10 cm. measurement location as for the other three measurement locations which gave very nearly the same curve. There are three possible explanations for deviation of the 10 cm. curve from the single unique curve required by theory. The first reason is that a higher bulk density existed from 0 to 10 cm.

than for points from 15 to 35 cm. (see Table 6). This greater bulk density would have increased the contact of soil particles and may have provided enough small pores to increase the ability of the soil to conduct water at the low moisture contents that existed near the wetting front. This increased ability of the soil to conduct water near the wetting front would have decreased the time required for a given moisture content to reach the 10 cm. distance as compared with the 19 cm. distance. A decreased time would result in an increase in the value of $x/t^{1/2}$ and would cause the observed direction of deviation. The second possible explanation is an unsatisfied boundary condition. If the moisture content increased at $x=0$ after the time water had reached the 10 cm. distance, then water which should have been used to bring the moisture content to its initial constant water content would have moved down the column. This would have decreased the time required for a given moisture content to reach the 10 cm. distance and would result in a deviation to the right from the unique curve as observed in Figure 2. A third possible explanation is that the vacuum which was applied at time zero, in an attempt to get the volume behind the plate filled with water as quickly as possible, changed the resistance to flow for the first few centimeters. Air holes were placed in the surface of the soil container and should have allowed atmospheric pressure to be maintained in the column at all

times, however for a short time the pressure in the first few centimeters may have been lower than atmospheric.

The deviation in Figure 2 is not extreme although Whisler, Klute, and Peters (1968) obtained a curve with less deviation. Whatever the cause of the decreased time required for the water to reach the 10 cm. distance relative to the time taken for the water to reach the other distances, this decreased time would have had to be in the order of 9 min. The diffusivities measured at 10 cm. for later times were assumed correct for this study.

The curve shown in Figure 2 can also be used to explain the frequently observed linear relation between x and $t^{1/2}$ for a horizontal advancing wet front. An observed wetting front results when there is visual change from dry to wet. This visual change will occur in the drier moisture contents; but, the moisture content at which the visual change occurs may vary from one observation to the next. In Figure 2, the value of $x/t^{1/2}$ is essentially constant even over a wide range of moisture contents in the dry range for our soil. Thus for this soil, the position of the observed wetting front for various times will result in a constant value for the relationship $x/t^{1/2}$ even though a constant moisture content was not followed when the wetting front was observed. When $x/t^{1/2}$ is constant, then $x = Ct^{1/2}$ which is the equation of a straight line.

The deviation, of the 10 cm. values of diffusivity in Figure 3 from the values of the diffusivity calculated from the data at the other measurement sites, is striking only at the low moisture contents. In general, a smooth curve can be drawn through the data for four measurement locations. This diffusivity relationship was measured for inflow conditions and will be applied to outflow experiments as was noted in a remark in the literature review. Some results from the literature for an outflow measurement of diffusivity for a Webster soil which can be closely compared with our inflow diffusivities are now considered.

Kunze and Kirkham (1962) used the outflow method of measuring the diffusivity. The Webster soil which they used was a clay loam with 33 percent sand, 34 percent silt, and 33 percent clay (our values for a Webster silty clay loam were 19.85 percent sand, 48.00 percent silt, and 32.16 percent clay). They packed their soil to a bulk density of 1.1 g./cm.^3 (our value was 1.3 g./cm.^3). Their values of diffusivity were almost linear when the logarithm of the diffusivity was plotted against volumetric water content for the range of water content from 45 to 15 percent. At 45, 30, and 15 percent water content their values of diffusivity are smaller than the values in our Figure 3 by factors of 10, 5, and 30 respectively. Despite the difference in soil properties, this gives some idea of the difference between absorption and desorption diffusivities.

Moisture and Temperature Profiles During Evaporation

Wind treated soil columns

For the wind-treated soil columns, the average bulk density from gamma-ray measurements was 0.01 g./cm.^3 greater than that calculated from the weight of soil and the volume of the column (Figure 4). Since gamma-ray bulk densities were calibrated in our experiment with gravimetric bulk densities by the measurement of the mass absorption coefficient (Table 4), the small difference between the average bulk density for the two measurement techniques results from the small soil volume sampled with the gamma beam. The range of bulk density shown in Figure 4 at any one depth is for the most part less than 0.1 g./cm.^3 . This variation is partly due to the thermocouple lead wires which come into the column on the right. The wires were firmly attached to the right wall of the column and evidently kept the soil in the right section of the column from being packed to the same density as the soil in the center and left section of the column except below 10 cm. where the thermocouples were 2 cm. apart. It was not felt that the variation in bulk density would cause any serious problems because below 300 mb. tension (38 percent water content in our case), only small variations in moisture content occur when bulk density is varied by 0.1 g./cm.^3 .

For the wind-treated soil column, variations in initial water content from the average moisture content in Figure 4 for any one depth are ± 1.25 to ± 1.50 percent except at the surface where they are larger. The variation of the deviations of gamma-ray measured moisture content from gravimetric moisture content shown in Table 5 averaged ± 0.6 percent. If it is assumed that little error occurred in the gravimetric moisture determinations shown in Table 5, then ± 0.6 percent is a measure of the variability of gamma-ray determinations and the variability of the gamma-ray measurements themselves account for a major part of the variation in moisture content at any one depth shown in Figure 4. Since the moisture shown in Figure 4 is held at tension of less than 300 mb., some of the variation in moisture content is also due to the variation in bulk density. A correlation of low bulk densities with high moisture contents can be seen in Figure 4. This effect is caused by changes in total porosity due to changes in bulk density.

The variation of moisture content shown in Figure 4 for the soil near the surface is caused by the swelling of soil during wetting. When soil swelling occurs, there is a smaller mass of soil remaining in the original volume. This smaller mass is indicated in gamma-ray measurements. Since the bulk density is assumed constant in order to obtain moisture content from the gamma-ray technique, the lowering in mass is registered as a decrease in moisture content. A

change in bulk density of 0.005 g./cm.^3 relative to the assumed constant bulk density will be registered as a change in water content of 0.4 percent.

Figure 5 shows the average initial and subsequent temperature measurements with depth for the wind treatment. The variation for the points at any one depth was usually less than 0.2°C . A significant minimum temperature should be noted in curves 1, 2, 3, and 4. This minimum temperature is at progressively greater depths and indicates that evaporation is taking place at a zone of sites located at depths of 0, 0.5, 1.0 and 3-4 cm. for times of 21.67, 66.25, 206.92, and 676.92 hrs. respectively. The evaporation at these depths must be occurring at a greater rate than at sites in the near vicinity and has been called the evaporation zone (H. R. Gardner and Hanks, 1966 and Fritton, Kirkham, and Shaw, 1967). The data of Wiegand and Taylor (1962) also show a minimum in the temperature profile when the data are replotted with depth. The data in Figure 5 show that at any one time the wind-treated column was near isothermal conditions, but that the isothermal condition is never achieved even at large times when the evaporation rate is low.

Figure 6 shows the fit of the cumulative evaporation curve from the diffusion theory with the experimental data for the wind treatment. The only problem is that the diffusivity function needed to fit the cumulative evaporation curve was smaller than the measured diffusivity by a factor of 35.0.

This is not a large factor for diffusivity measurements because diffusivity values, as our Figure 3 shows, vary by a factor of 5000 when the diffusivities at low and high moisture contents are compared. Since the shape of the two diffusivity functions is unchanged when the factor is used, the statement made in the literature review that the inflow and outflow diffusivities would tend to increase with increasing water content would seem to be true. Since the measured inflow diffusivity is greater than the diffusivity needed to fit the outflow experimental data, the statement made in the literature review that absorption is faster than desorption is true for our conditions. The difference found here between the ratio of inflow and outflow diffusivities is greater than the difference calculated by W. R. Gardner (1959) and is less than some of the differences measured by Klute, Whisler, and Scott (1964). From the successful fit of the cumulative evaporation data from our wind-treated soil column (Figure 6) it is concluded that the shape of the measured diffusivity curve, as measured, is adequate for the explanation of evaporation from the wind-treated soil column. Data given by W. R. Gardner (1959) indicate that the correct magnitude of the diffusivity can be measured using outflow techniques.

The factor of 35.0, which had to be used in Figure 6 to fit the experimental cumulative evaporation points from the wind-treated soil column, was found by trial and error. This

factor is a measure of some greater resistance in the evaporation soil column compared with the resistance in the inflow soil column which hinders the movement of liquid water to the vaporization zone or the movement of water vapor from the vaporization zone to the atmosphere, or both. Several reasons can be suggested for this greater resistance. The first is the difference between inflow and outflow diffusivities which may vary from a factor of 0 to a factor of 100 (W. R. Gardner, 1959; Klute, Whisler, and Scott, 1964; and Jackson, 1964). A second reason for the resistance factor might be the deviation that was seen in Figure 2. In this figure, the water contents which were measured at the 10 cm. location, and later used to calculate diffusivity using Equation 23, did not fall on the same curve as water contents which were measured at three other locations. This deviation could indicate a violation of the boundary or initial conditions given in Equations 18, 19, and 20 and could have resulted in incorrect diffusivity values. A third reason for a diffusivity resistance factor might be a thin compacted layer at the surface of the soil. It is possible that use of a lead brick which was placed on the soil surface during wetting caused a thin compressed layer of soil to be formed at the surface of the soil. A compressed layer of soil at the surface of the soil column would introduce a true resistance to both liquid and vapor water flow. It should be mentioned that no such layer was

visible during the sectioning of either column. With regard to the linear factor of 35.0, it is noteworthy that this single value served over the full moisture range (6 to 45 percent) and the full time range (0 to 676.92 hr.) for the wind evaporation experiment.

In considering the deviation in Figure 7 of the experimental curves from the calculated, it is recalled that the variation of moisture content from the average at any one depth was approximately ± 1.5 percent in Figure 4. This same value of variation (± 1.5 percent) can be assumed to apply to the curves of Figure 7 except in the upper layers. With this amount of moisture variation in the experimental data, the deviation of points on the experimental curves from the theoretical curves would not be significantly different for the curves 1 and 2 of 21.67 and 66.25 hrs. We conclude from Figure 7 that the isothermal diffusion theory explains the distribution of moisture in a soil column for the early times of a nearly isothermal evaporation of water from a soil column.

Curve 4 (676.92 hrs. in Figure 7) shows moisture content plotted versus depth as determined by two methods. The two methods were possible because at the end of the experiment, 676.92 hrs., a gamma-ray determination could be followed by destructive sampling of the soil for gravimetric moisture determinations. The theoretical curve fails to match the shape of either of the experimental curves over the full 19 cm.

depth of the vertical soil column. For the surface depths from 0 to 7 cm., the theoretical curve needs to be moved to the left from 5 to 0 percent. For the 10 to 19 cm. depths, the theoretical curve needs to be moved to the right from 3 to 2 percent. There is, however, no basis in the diffusion theory, as we know it, to make such adjustments. There are reasons why the theoretical curve of the three curves numbered 4 on Figure 7 departs from the experimental curves. From the minimum shown in the temperature curves of Figure 5, it can be concluded that water changed to the vapor phase at the 4 to 5 cm. depth for the 676.92 hr. time. It can then be speculated that the reason for the incorrect shape of the theoretical curves was the presence of a dry layer at the surface which restricted the upward flow of water. Another speculation which could help explain this incorrect shape is that the water content throughout the soil column had reached the point at which liquid continuity is said to fail. This point is frequently taken as 1 bar (31 percent moisture content in our case). If liquid continuity truly fails at 31 percent moisture content, then water movement from that time on must be at least partially in the vapor phase. Since the vapor pressure of water is still near its saturation value until a moisture content of near the wilting point (15 bars) is reached, there would be little movement of water from the lower layers of the column and the upper layers would tend to

dry at a faster rate than the lower layers. Whatever the explanation, the diffusion equation as it has been used in this study predicts an inadequate moisture distribution when a dry layer exists on the surface of the soil column even though it predicts the cumulative evaporation.

Radiation treated soil column

The variation in bulk density in Figure 8 at any one depth is somewhat less than that shown in Figure 4, and the variation in water content at any one depth is about the same as that shown in Figure 4. The reasons for this variability of bulk density and of moisture content which are given in the discussion of Figure 4 apply to Figure 8 as well.

- The variation of temperature mentioned in the results for Figure 9 for depths near the surface was probably due to an error in placement of the thermocouples. Since the surface was most affected by the swelling of the soil, the thermocouples, even if they were initially located perfectly, were probably moved upward unevenly. An error in location of the surface thermocouples of 2 mm. could have caused the observed 5°C variation in temperature.

Most of the temperature rise observed for the radiation treated soil column shown in Figure 9 occurred during the first day. At all times after the first day, the soil column stayed at a relatively stable temperature distribution and continued to rise in temperature at all depths very slowly.

This temperature stability allowed the approximation of the process as a steady state so that the analysis of Philip and De Vries (1957) could be used.

Even though the surface temperatures are very high, ranging from 75 to 88°C, they exemplify the conditions that would exist if a net radiation of approximately 1.2 cal./cm.² min. were to fall continuously on a soil with little wind movement. There was considerable movement of heat out the sides of the soil column under the radiation treatment, but the horizontal temperature gradient must have been smaller than the value of the variation in thermocouple readings because no consistent outward gradient could be seen in the temperature data. Nevertheless, there could have been some effect of a two dimensional soil in these experiments due to outward moving heat which would mean the temperature at the edges of the columns was cooler than at the center. This would cause water vapor to move toward the outside of the impervious column rather than upward. Since the vertical temperature gradient as seen from the slope of the curves in Figure 9 was much greater than any possible gradient toward the outside of the columns, we will assume a horizontal temperature gradient can be neglected for the purposes of this study.

In Figure 10, the diffusivity resistance factor of 10.0 needed to fit the cumulative evaporation data for the radiation treated soil column is 3.5 times greater than the resistance factor of 35.0 which was used to fit the cumulative

evaporation data for the wind treatment. The difference in the two resistance factors can be explained by the difference in temperature of the soil and its water, in the wind and radiation treatments. To see the effect of the temperature difference, consider the definition of the diffusivity in terms of the hydraulic conductivity and the slope of the soil-water retention curve, $D_{\theta, liq} = K(\partial \psi / \partial \theta)$. Here, the hydraulic conductivity K is temperature dependent because it involves the viscosity of water as given by $K = \rho g k / n$, where ρ is the density of water, g is the acceleration of gravity, k is the intrinsic permeability, and n is the viscosity. The viscosity of water at 25°C is 0.00891 poises and at 75°C is 0.00380 poises. This viscosity decrease results in an increase by a factor of $(0.00891/0.00380 =) 2.34$ in the hydraulic conductivity and thus, in the diffusivity as compared with the diffusivity used in the wind treatment experiment. The factor 2.34 is smaller than the factor of 3.5 which would be needed. Another factor of 1.49 would be needed to completely explain the difference in the resistance factor used for wind and for radiation. This factor of 1.49 will be left as a experimental factor which is due to the difference in packing of the wind and radiation treated soil columns. Thus, the difference in the diffusivity which is used to fit the cumulative evaporation curves of the wind and the radiation treated soil columns can be explained for the most part by the

difference in temperature of the two soil columns during evaporation.

Figure 11 is a graph to show how poor a theoretical fit of our experimental moisture data is achieved if temperature influences other than the linear change in the diffusivity just discussed are not considered. The factor used to adjust the diffusivity is the same as that used to calculate the cumulative evaporation curve (Figure 10) for this same radiation treated soil column. The fit is poor except for the curves labeled '1' because the diffusion theory does not predict the formation of a dry surface layer.

Table 7 indicates the magnitude of the salt movement in the wind and radiation treatments. There is a definite accumulation of salt in the surface 0 to 1 or 0 to 2 cm. layer of soil in the vertical column like that found by Fritton, Kirkham, and Shaw (1967) in another evaporation experiment. Salt can cause changes in gamma-ray measurements because a movement of salt is a movement of mass and the quantity of mass is measured by the gamma-ray technique. A change in salt content relative to the initial salt content of the soil of 0.5 mg./g. can be calculated to be roughly equivalent to a change in water content of 0.05 percent. From the data in Table 7, it can be shown that mass changes due to salt movement cannot cause the observed difference between the gamma-ray and gravimetric determinations presented in Figure 11.

Salt movement can also cause water movement due to osmotic forces. Any errors which have been caused by osmotic forces will be ignored since the changes shown in Table 7 are small.

Moisture Profiles from Heat and Mass Transfer

Figure 12 shows the data from the net radiometer and the heat flow disks. The rise in heat flux passing the disk located near the surface between the 0 and 16 hr. times indicates evaporation is taking place below the level of the heat flux disk almost from the time the heat lamps were turned on. H. R. Gardner and Hanks (1966) found the same effect. The amount of heat moving into the column decreased after the evaporation zone moved below the surface with time similar to the data of H. R. Gardner and Hanks (1966). This decrease is caused by the reradiation of energy due to an increase in surface temperature. The data from the heat flux measurement that showed evaporation was taking place below the surface after the 16 hr. time is consistent with the moisture content data in Figure 11 showing a dry surface layer at the 20.58 hr. time.

In Figure 17, the difference between the gamma-ray and gravimetric moisture contents was probably due to the problem of shrinking of the soil during drying, and the gravimetric curve is probably the most reliable. The calculation of the theoretical curve from Equation 49 involves many assumptions

especially in the calculation of the values of the thermal conductivity, the thermal vapor moisture diffusivity, and the slope of the tension temperature curve. All three of these functions are likely valid because they correspond to experimental observations in De Vries (1963), Philip and De Vries (1957), and Gardner (1955) and Taylor, Evans, and Kemper (1961). The values of the other functions in Equation 49 are more closely connected to experimental observations. The cause of the deviation of the theoretical moisture profile from the experimental moisture profiles for moisture contents above 20 percent at depths below 12 cm. is not known. One possibility is that the liquid moisture diffusivity should decrease more rapidly in the region of 20 to 30 percent moisture content similar to the curve of Kunze and Kirkham (1962) and that this caused the observed deviation.

In any case, there is little question that the heat and mass transfer theoretical curve gives a much better fit of the experimental points than the isothermal diffusion equation where temperature gradients are important (Figures 11 and 17).

CONCLUSIONS

The conclusions which follow are believed to be valid for a Webster silty clay loam soil taken from 0 to 30 cm. of a profile and packed into a soil column to a depth of 20 cm., wet to a moisture content in the range of 80 to 90 percent of saturation, and dried by air movement or radiation.

With regard to the validity of the isothermal diffusion equation when it is applied to prediction of water movement under nearly isothermal (wind) conditions, it is concluded that:

1. The isothermal diffusion equation (Equation 17) can be used to calculate the cumulative evaporation as a function of time (Figure 6).
2. The isothermal diffusion equation can be used to calculate the moisture content distribution until liquid continuity is lost and vapor movement occurs through a surface layer of dry soil (Figures 1 and 7).
3. A diffusivity which is measured as a function of water content using inflow experimental techniques can be used to predict nearly isothermal outflow water movement provided the diffusivity is divided by a resistance factor (Figures 6 and 7).

4. The isothermal diffusion equation will not predict the development of a dry surface layer during evaporation (Figure 7).
5. Profiles of temperature with depth show a minimum temperature indicating the vaporization zone which occurs at an increasing depth as time increases (Figure 5).

With regard to the validity of the isothermal diffusion equation when it is applied to prediction of water movement under non-isothermal (radiation) conditions it is concluded that:

6. Because of compensating vapor movements, the isothermal diffusion equation can be used under non-isothermal conditions to predict cumulative evaporation if the diffusivity function is adjusted to correct for the difference between the temperature at which the diffusivity was determined and the temperature at which the evaporation took place (Figure 10).
7. The isothermal diffusion equation cannot be used to predict the moisture content distribution where there are temperature gradients because the equation cannot predict the development of a dry surface layer which occurs when there are temperature gradients (Figure 11).

With regard to the validity of the heat and mass transfer method of calculating the moisture distribution using an

equation given by Philip (1957), it is concluded that:

8. Heat and mass transfer can be used to predict the development of a surface dry layer during evaporation (Figure 17).
9. Heat and mass transfer can be used to predict moisture distribution during evaporation where temperature gradients are important if enough data is available to characterize the soil sufficiently (Figure 17).

With regard to the gamma-ray attenuation method of measuring moisture content, it is concluded that:

10. The method described to measure the resolving time of the gamma-ray apparatus will result in a resolving time which will act as a calibration of the gamma-ray measured bulk density and moisture content with those measured gravimetrically (Tables 2, 4, and 5).
11. The mass absorption coefficient will be closer to the theoretical value when the mass absorption coefficient is calculated using data corrected for resolving time than for data not corrected for resolving time (Table 2).
12. Water contents calculated from data corrected for resolving time will correlate better with gravimetric water contents than if no resolving time correction is made (Table 5).

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APPENDIX A

The following is an explanation of the undefined symbols used in the program from which the best resolving time (called dead time in the program) was determined.

RO - observed gamma-ray intensity (counts/min.)

RC - gamma-ray intensity corrected for resolving time
(counts/min.)

T - resolving time (min.)

X - thickness of material through which gamma-rays pass (cm.)

NOBS - number of observations

NTIME - number of resolving time calculations

DTIME - constant to control the advance of the resolving
time value (micro-min.)

```

C      CALCULATION OF MINIMUM DEVIATIONS TO FIND BEST DEAD TIME
      REAL INTER
      DIMENSION RC(50),X(50),RC(50)
      READ(1,1)NOBS,NTIME,OTIME
1     FORMAT(2I2,F10.9)
      READ(1,2)(RC(I),I=1,NOBS)
2     FORMAT(9F8.0)
      READ(1,3)(X(I),I=1,NOBS)
3     FORMAT(9F8.4)
      WRITE(3,4)
4     FORMAT('CALCULATION OF MINIMUM DEVIATIONS TO FIND THE
      1BEST DEAD TIME FOR THE GAMMA APPARATUS' / 10X 'WRITTEN AND
      1HARRICK      AUGUST 1967' // 5X 'DEAD TIME(MICRO-SEC.) -NU=AND'
      1 13X 'CN(100)' 12X 'VARIANCE' / 1)
      SX=0
      SXSQ=0
      DO 5 I=1,NOBS
      SX= SX+ X(I)
5     SXSQ= SXSQ + X(I)*X(I)
      DEN= SX*SX - NOBS * SXSQ
      DO 10 ITIME=1,NTIME
      T= ITIME *OTIME + .000001
      SRC=0
      SXR= 0
      DO 6 IOBS=1,NOBS
      RC(IOBS)= RC(IOBS) / ( 1. - T * RC(IOBS))
      RCLOG= ALOG(RC(IOBS))
      SRC= SRC + RCLOG
6     SXR= X(IOBS) * RCLOG + SXR
      SLOPE= (SX* SRC - NOBS * SXR) / DEN
      INTER= (SRC - SLOPE * SX) / NOBS
      VAR= 0
      DO 7 IOBS=1,NOBS
7     VAR= ( EXP( SLOPE+ X(IOBS) + INTER)- RC(IOBS) )**2 / RC(IOBS)
      1 *VAR

```

```
TIME=TIME+.1  
10 WRITE(3,11)TIME,SLOPE,INTER,VAR  
11 FORMAT(14X,F4.1,2E20.6)  
STOP  
END
```

APPENDIX B

The following is an explanation of the undefined symbols used in the program in which the bulk density is calculated from gamma-ray data.

XCOL = inside dimension of column (cm.)

T = resolving time (min.)

USOIL = mass absorption coefficient of soil ($\text{cm}^2/\text{g}.$)

UWATER = mass absorption coefficient of water ($\text{cm}^2/\text{g}.$)

THETAW = air dry water content by weight (g./g.)

EMPTY = gamma-ray intensity through empty column
(counts/min.)

RAIRDR = gamma-ray intensity through air-dry soil in
the column (counts/min.)

PSOIL = bulk density of soil (g./cm^3)

```

C   CALCULATION OF BULK DENSITY FROM GAMMA DATA (EXIT ON MARCH 21, 1966)
   DIMENSION REMPTY (2,22,3), RAINCR (2,22,3), EMPTY1 (2,22,3),
   AIRDR1 (2,22,3), XCOL (2), PSOIL (2,22,3), PAVE (2,22)
   XCOL (1) = 10.75
   XCOL (2) = 10.86
   T = 8.333333E-8
   USGIL = 0.07866
   UWATER = 0.08648
   THETAH = 0.0628
   DO 5 I = 1,2
6   READ (1,2) ((EMPTY (I,J,K), J=1,22), K=1,3)
7   READ (1,2) ((AIRDR (I,J,K), J=1,22), K=1,3)
8   FORMAT (9F8.0)
   DO 5 J = 1,22
   PAVE (1,J) = 0.0
   DO 1 K = 1,3
   EMPTY1 (I,J,K) = REMPTY (I,J,K) / (1.0 - REMPTY (I,J,K) * T)
   AIRDR1 (I,J,K) = RAINCR (I,J,K) / (1.0 - RAINCR (I,J,K) * T)
   PSOIL (I,J,K) = (ALGO (AIRDR1 (I,J,K) / EMPTY1 (I,J,K))) /
1 (-USGIL * XCOL (1) - UWATER * XCOL (1) * THETAH)
   PAVE (1,J) = PAVE (1,J) + PSOIL (I,J,K)
1 CONTINUE
   PAVE (1,J) = PAVE (1,J) / 3
5 WRITE (3,3) 1,J, (PSOIL (1,J,K), K=1,3), PAVE (1,J)
3 FORMAT (' COLUMN', 5X, 12, 5X, 'DEPTH', 5X, 12, 5X, 4(F7.4, 5X))
STOP
END

```

APPENDIX C

The following is an explanation of the undefined symbols used in the program in which the water content is calculated from gamma-ray data. Symbols which are identical to those in Appendix B will not be repeated.

RWET - gamma-ray intensity through wet soil in the column (counts/min.)

PWATER - density of water (g./cm.^3)

THETAV - water content by volume ($\text{cm.}^3/\text{cm.}^3$)


```

C   CALCULATION OF VOLUMETRIC WATER CONTENT FROM GAMMA DATA (SPITTON 4/5/68)
      DIMENSION KEMPTY (2,22,3), RAIROD (2,22,3), EMPTYI (2,22,3),
      RAIRODI (2,22,3), XCCL (2), PSOIL (2,22,3),
      IRWET (2,22,3), WETI (2,22,3), THETAIV (2,22,3), THETAII (2,22)
      XCCL (1) = 10.75
      XCCL (2) = 10.80
      T = 6.333333E-8
      USOIL = 0.07806
      UWATER = 0.08668
      PWATER = 0.997044
      THETAII = 0.0628
      DO 5 I = 1,2
9 READ (1,2) ((KEMPTY (I,J,K), J=1,22), K=1,3)
7 READ (1,2) ((RAIROD (I,J,K), J=1,22), K=1,3)
4 READ (1,2) ((IRWET (I,J,K), J = 1,22), K = 1,3)
2 FORMAT (9F8.0)
      DO 5 J = 1,22
      THETAII (I,J) = 0.0
      DO 1 K = 1,3
      EMPTYI (I,J,K) = KEMPTY (I,J,K) / (1.0 - KEMPTY (I,J,K) * T)
      AIRODI (I,J,K) = RAIROD (I,J,K) / (1.0 - RAIROD (I,J,K) * T)
      WETI (I,J,K) = IRWET (I,J,K) / (1.0 - IRWET (I,J,K) * T)
      PSOIL (I,J,K) = (ALOG (AIRODI (I,J,K) / EMPTYI (I,J,K))) /
1 (-USOIL * XCCL (1) - UWATER * XCCL (1) * THETAII)
      THETAIV (I,J,K) = ((ALOG (WETI (I,J,K) / AIRODI (I,J,K))) -
1 (UWATER * XCCL (1) * PSOIL (I,J,K) * THETAII)) / (-UWATER * PWATER *
1 XCCL (1))
      THETAII (I,J) = THETAII (I,J) + THETAIV (I,J,K)
1 CONTINUE
      THETAII (I,J) = THETAII (I,J) / 3
5 WRITE (3,3) I,J, (THETAIV (I,J,K), K = 1,3), THETAII (I,J)
3 FORMAT (' COLUMN', 5X, 12, 5X, 'DEPTH', 5X, 12, 5X, 4(7.4, 5X))
      STOP
      END

```

APPENDIX D

The following is an explanation of the undefined symbols used in the program in which soil water diffusivity was calculated (Equation 23). Symbols which are identical to those in Appendixes B and C will not be repeated.

DIFFUS = soil water diffusivity ($\text{cm.}^2/\text{min.}$)

TIME = cumulative time after initial wetting (min.)

N = number of gamma-ray intensity measurements used

Y = distance from porous plate (cm.)

BTIME = cumulative time for the first gamma-ray intensity measurement used at a location (min.)

```

C      CALCULATION OF SOIL WATER DIFFUSIVITY USING THE TECHNIQUE OF
C      KRISLER ET. AL. SSSAP 3213-11(1968) (KRIFFEN 4/19/66)
      DIMENSION KALY (1233), THETAV (1333), DIFFUS (1333),
      LTIME (1333), ASUM (1333)
      N = 279
      THETAH = 0.0027
      T = 8.333333E-8
      USOIL = 0.07866
      UWATER = 0.06668
      PWATER = 0.996373
      ASUM (1) = 0
      READ (1,3) XCCL, RAIROA, Y, KEMPTY, BTIME
3  FORMAT (F4.2, F8.0, F3.0, F8.0, F8.4)
      TIME (1) = BTIME * 5.0 / 60.0
      AIRONI = RAIROA / (1.0 - RAIROA * T)
      EMPTYI = KEMPTY / (1.0 - KEMPTY * T)
      PSOIL = (ALOG(AIRONI / EMPTYI)) /
11-USOIL*XCCL - UWATER * XCCL * THETAH)
      READ (1,2) (RHET (J), J=1,N)
2  FORMAT (9F8.0)
      DO 1 J=1,277
      ARETI = (6.*RHET (J)) / (1.0 - 6.*RHET (J)*T)
      BRETI = (6.*RHET (J+1)) / (1.0 - 6.*RHET (J+1)*T)
      CRETI = (6.*RHET (J+2)) / (1.0 - 6.*RHET (J+2)*T)
      THETAH (J) = ((ALOG (ARETI / AIRONI)) -
1UWATER * XCCL * PSOIL * THETAH (J) / (1-UWATER * PWATER *
1XCCL))
      THETAH (J+1) = ((ALOG (BRETI / AIRONI)) -
1UWATER * XCCL * PSOIL * THETAH (J) / (1-UWATER * PWATER *
1XCCL))
      THETAH (J+2) = ((ALOG (CRETI / AIRONI)) -
1UWATER * XCCL * PSOIL * THETAH (J) / (1-UWATER * PWATER *
1XCCL))
      TIME (J+1) = TIME (J) + 13.2754/60.0
      TIME (J+2) = TIME (J+1) + 13.2754/60.0

```

```

      ASUM ( J+1)=ASUM ( J) + (1.0/SCRT (TIME (J+1)) +
1 1.0/SCRT (TIME (J+1))) * ((THETAV (J+1) - THETAV (J+1))/2.0)
      DIFFUS ( J) = ( 0.25)* ((Y *Y )/(4 ( TIME (J) + TIME (J+1))/2)
1 * (SQRT((TIME (J) + TIME (J+1))/2))) *
1 ((TIME(J+1) - TIME ( J) )/ ( THETAV (J+1) - THETAV ( J))) *
1 ASUM ( J)
1 WRITE (3,4) DIFFUS ( J),THETAV (J+1), TIME (J+1) ,ASUM (J)
4 FORMAT ( E15.5, F13.8, F12.6, E15.5)
      STOP
      END

```

APPENDIX E

The following is an explanation of the undefined symbols used in the program in which moisture distribution was calculated from the isothermal diffusion equation (Equation 31).

NMAX = number of water contents from 0 to saturation

K = number of depths

N = NMAX

CONSL = unused constant

TIMEC = unused constants

T = array of volumetric water content in steps of 1 percent from 0 to saturation ($\text{cm.}^3/\text{cm.}^3$)

P = array of water pressure (negative) corresponding to the volumetric water content array (cm. of water)

D = values of the summed diffusivity multiplied by DELW corresponding to volumetric water content array ($\text{cm.}^2 \text{ cm.}^3/\text{min. cm.}^3$)

W = initial water content distribution from bottom to top of column ($\text{cm.}^3/\text{cm.}^3$)

H = initial pressure (negative) distribution from bottom to top of column (cm. of water)

DELT = initial value of delta time = 0.01 (hr.)

DELW = water content interval = 0.01 ($\text{cm.}^3/\text{cm.}^3$)

CONS7 = unused constant

CONSH = unused constant

TTT = 0.5 = constant which controls type of difference solution (see Richmyer, 1957, p. 94)

CONT - unused constant

DELX - depth interval = 1 (cm.)

GRAVY - gravitational term = DELX

TIME - initial value of cumulative time = 0.00 (hr.)

CUMMC - initial value of cumulative evaporation = 0.00 (cm.)

CONB - unused constant

CONQ - interval of cumulative evaporation from which DELT
is estimated = 0.1 (cm.)

CONBB - initial value of hydraulic conductivity to be
stored at all depths (cm./hr.)

CONCC - initial value of inverted water capacity to be
stored at all depths (cm. of water/cm.³/cm.³)


```

C      EVAPORATION OF SOIL WATER FROM A SOIL COLUMN OF FINITE LENGTH
C      HANKS AND GARDNER 1965, SSSAP 29:490-499, IFAITON 5/8/69)
      DIMENSION A(70), B(70), C(70), Y(70), G(70), Z(70), P(70),
      1 H(70), K(70), O(70), V(70),      1(70), DE1(70), DE2(70), TIMEC (70)
500 READ(1,501)NMAX
      IF(NMAX)113,113,502
501 FORMAT(12)
502 DO 504 I=1,NMAX
      P(I)=0.0
504 O(I)=0.0
      READ (1,116)K,      A,      CONSI
116 FORMAT (2I3,F4.3)
      READ(1,1000)TIMEC(111),11=1,16)
1000 FORMAT (8F10.2)
      11 =1
      KK=K
      KKK=K-1
      68 READ (1,117)(T(I),I=1,NMAX)
117 FORMAT (2X,26F3.2)
      89 READ(1,119)(P(I),I=1,N)
119 FORMAT (7F11.0)
      91 READ(1,121)(O(I),I=1,N)
121 FORMAT(6E10.3)
      87 READ(1,91)(K(I),I=1,KK)
      81 FORMAT (16F5.4)
      82 READ (1,121) (H(I),I=1,KK)
      92 READ(1,115)DELT,DELT,CONST,CONSH,ITT,CCAT,DELT,
      1 GRAY,TIME,CUNNC,CONS,CONE,CUNR, CONCC
115 FORMAT (6F5.2,F5.0,F5.2,2F5.4,2F5.3,F7.4,F7.0)
      DE1(1)=21
      DE2(1)=20
      DO 149 I=2,K
      DE1(I) =DE1(I-1)-DELT
149 DE2(I) = DE2(I-1)- DELT
      WRITE(3,254)(1,T(I),P(I),O(I),K(I),H(I),I=1,NMAX)

```

```

254 FORMAT (1H,12,F6.2,F11.0,E11.1,F8.4,F11.1)
DO 93 I=1,KK
53 V(I)=W(I)
THO = 1.0-TIT
PII=0.0
CONSB = CONSB / 10.0
DO 1000 I=1,K
B(I) = CONSB
1000 C(I) = CACC
DO 55 I=1,KKK
65 PII= PII+W(I)
OXSQB = DELX* DELX
43 DO 40 I = 1,K
40 C(I)=I(I)
DO 4012 J = 1,70
IF (H(2) - T(J)) 4011,4011,4012
4012 CONTINUE
4011 DIFFA = (H(2) - T(J-1)) * (C(J) - C(J-1))/DELX + C(J-1)
DO 41 I=1,KKK
IF (H(I)-H(I+1)) 710,712,710
710 DO 44 J=1,70
IF(H(I+1)-T(J)) 47,47,44
44 CONTINUE
47 DIFFB=(H(I+1)-T(J-1)) * (C(J)-C(J-1))/DELX + C(J-1)
72 DIFF = DIFFA - DIFFB
B(I) = (DIFF / (H(I)-H(I+1))) * 60.
B(I) = B(I) / 10.0
WRITE (3,3000) DIFF, DIFFA, DIFFB, B(I)
3000 FORMAT (1H,5E15.5)
54 DIFFA = DIFFB
712 DO 52 J=1,70
IF(V(I+1)-T(J)) 54,54,52
52 CONTINUE
54 C(I) = (P(I) - P(J-1))/DELX
41 CONTINUE

```

```

1 DO 2 I=1,K
2 A(I) = (DELT/DKSCD)* Q(I)
  A(I) = 0.0
  WRITE (3,3001) (A(I), I= 1,K)
3001 FORMAT (1H,3E15.5)
  KKKK = K -2
  Y(1) = 0.0
  Z(1) = H(1)
3 DO 5 I=1,KKKK
4 TEMPA = A(I)*C(I)+111
  TEMPZ = TEMPA*Y(I)
  TEMPY = ((11A(I+1)+ A(I))* C(I) +111)+1.03-TEMPZ
  Y(I+1) = (111+1)*111*C(I)/TEMPY
  TEMPH = 11(G(I+2)-G(I+1))+THO-(GRAVY)*A(I+1)
  TEMPU = (TEMPH +11(G(I)-G(I+1))*111C)+GRAVY*A(I)*C(I)+ G(I+1)
  Z(I+1) = ((TEMPA*Z(I))+ TEMPU)/TEMPY
  *R11= (3,3002) TEMPA, TEMPZ, TEMPY, TEMPH, TEMPU
3002 FORMAT (1H,5E15.5)
5 CONTINUE
  I= K-2
  12 H(I+1) = H(I+2) + Y(I+1) + Z(I+1)
2000 WRITE (3,2001) I, H(I+1), Y(I+1), Z(I+1),H(I+2)
2001 FORMAT (2H, 12, 4F15.3)
10 IF(I-1) 14,14,5
  6 I=I-1
  GO TO 12
14 DO 170 I=1,KK
26 DO 28 J=1,70
  IF (H(I) )-P(J)) 29,29,28
28 CONTINUE
29 WATER = (H(I) )-P(J-1))/(P(J)-P(J-1))*DELTA+11J-1)
39 V(I) =WATER
37 N(I) =WATER
170 CONTINUE
74 TIME =TIME + DELT

```

```

76 JJ=1
   JJJ=10
   WRITE (3,151) TIME
151 FORMAT (1H0,12H TIME(HOURS)=,F9.4)
157 WRITE (3,153)(DELT(I),DELT(I),I=JJ,JJJ)
153 FORMAT(1H , 9HDELT(CH),F5.1,1P-,F4.1, 19(Fc.1,1d-,F4.1))
   WRITE (3,155)(G(I),I=JJ,JJJ)
155 FORMAT (1H ,8H G VOL ,10F11.4)
   WRITE (3,158)(H(I),I=JJ,JJJ)
158 FORMAT (1H ,9HHCAG (CH),10F11.1,/)
   IF (JJJ-K) 167,165,169
167 JJ=JJ+10
   JJJ=JJJ+10
   IF (JJJ-K) 157,157,159
159 JJJ=X
   GO TO 157
165 CONTINUE
   WRITE (3,168)
168 FORMAT (1H0,31HACC TIME(HR)   ACC TIME(MIN)
13HDELTA TIME(HR)   ACC EVAP (CH) ,
14HEVAP RATE 1 (CH/HR)   EVAP RATE 2(CH/HR))
19 CONTINUE
   TIME1= TIME+60.0
20 SUM=0.0
21 DO 22 I=1,KKK
22 SUM=H(I )+SUM
23 SUMC =(PIT-SUM)+DELX
   PIT=SUM
25 CUMPC= CUMPC  +SUMC
   WFR2= SUMC /DELT
   IF (WFR2 - 0.154)1012,1012,1013
1013 TIME = TIME +((WFR2/0.154)*DELT)- DELT
   WRITE (3,151) TIME
1012 IF (X(K) + 5000000.)1002,1002,1005
1005 IF (X(K2)1001,1004,1004

```

```

1004 IF (WFR2 - 0.154)1002,1002,1001
1002 W(K) = - 5000000.
      W(K) = 0.0200
      II=II+1
      GO TO 1003
1001 W(K) = -311.
      W(K) = 0.3853
1003 WRITE (3,13) TIME, TIME1, DELT, CUNNC,WFR,WFR2
      13 FORMAT (1H ,F10.4,2F15.4,F17.4,F19.4,F22.4)
      IF (II-2)1010,1011,1010
1011 DELT = 0.02
      GO TO 110
1010 DELTY = DELT
      IF (WFR2)1006,1006,1007
1006 DELT = 2.0* DELT
      GO TO 110
1009 DELT = 2.0* DELTY
      GO TO 110
1007 DELT = CUNC/WFR2
      IF (DELT - 24.0)110,110,1008
1008 IF (TIME -24.0) 1009,1009,1100
1100 DELT = 22.0
      110 IF (W(2)-0.15) 500,150,150
      150 GO TO 43
      113 WRITE(3,114)
      114 FORMAT (1H ,10HNMAX IS 00)
      112 STOP
      END

```

APPENDIX F

The following is an explanation of the undefined symbols used in the program from which the moisture distribution is calculated from the heat and mass transfer equation (Equation 49).

E = evaporation rate (cm./sec.)

Q = soil heat flux (cal./cm.² sec.)

S = gas constant of water vapor (erg./g. °C)

DELO = water content interval (cm.³/cm.³)

N = number of water content intervals between 0 percent and saturation

CONDA = thermal conductivity of dry air (cal./cm. sec. °C)

CONDW = thermal conductivity of water (cal./cm. sec. °C)

CONDS = thermal conductivity of soil mineral (cal./cm. sec. °C)

GG = acceleration due to gravity (cm./sec.²)

XS = volumetric content of soil (cm.³/cm.³)

ALPHA = tortuosity factor for the diffusion of gases in soils

THETAO = volumetric water content of the surface layer of soil (cm.³/cm.³)

TEMP = temperature array in steps of 10 (°C)

CONDV = unused constants

HEATL = latent heat of vaporization (cal./g.)

SURTEN = surface tension (dynes/cm.)

RHOO = density of saturated water vapor (g./cm.³)

PRESWV = partial pressure of water vapor (mm. of Hg)

DENSW = density of water (g./cm.^3)

T = temperature distribution of soil column ($^{\circ}\text{C}$)

O = water content array in steps of DELO ($\text{cm.}^3/\text{cm.}^3$)

P = pressure (negative) array corresponding to the water content array (cm. of water)

DOLIQ = liquid water diffusivity array corresponding to the water content array (cm./min.)


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C  CALCULATION OF MOISTURE AND TEMPERATURE PROFILES ACCORDING TO
C  J.R. PHILIP (1957) EVAPORATION, AND MOISTURE AND HEAT FIELDS IN THE
C  SOIL. JOURNAL OF METEOROLOGY 14:354-366. (PRITTON 6/23/68)
C  DIMENSION F(53), Q(53), TK(25), VPCDEF(20), VISC(10), AVISC(20),
1  TEMP(10), CONDV(10), HEATL(10), SURTEN(10), RHOD(10),
1  PRESV(10), DENS(10), Q(53), F(53), DOLIQ(53), C(53), D(53),
1  DTLEQ(53), T(25), XH(53), A(53), RHQ(53), DQVAP(53), KA(53), CONQ(53),
1  DIVAP(53), DT(53), GCW(53), ARHCO(20), CONQV(20), AHEATL(20),
1  ASUR(20), APKE(20), ADENSH(20), BETA(20), STCCF(20), QATH(20), V(20)
C  DIMENSION PP(53), COOLQ(53)
C  E = 0.8072E-06
C  Q = -0.453E-03
C  S = 4.0150E 06
C  DELQ = 0.01
C  N=53
C  NN = N-1
C  CONDA = 0.0015E-03
C  CONDK = 1.57E-03
C  CONDS = 5.1E-03
C  CG = 980.6
C  XS = 0.48
C  ALPHA = 0.66
C  THETA = 0.02
C  READ (1,1) (TEMP(1), CONDV(1), HEATL(1), SURTEN(1), RHOD(1),
1  PRESV(1), DENS(1), VISC(1), I = 1,10)
1  FORMAT (F4.0, E10.2, F9.5, F10.5, E15.5, F10.4, F10.1, F10.4)
C  READ (1,2) (T(J), J=1,20)
2  FORMAT (10F7.3)
C  DO 3 J=1,20
C  DO 4 I=1,10
C  IF (T(J)-TEMP(I)) 3,5,4
4  CONTINUE
5  TERPOL = (T(J) - TEMP(I-1))/10.0
C  ARHCO(J) = TERPOL * (RHOD(I)-RHOD(I-1)) + RHOD(I-1)
C  BETA(J) = (RHOD(I) - RHOD(I-1))/10.0

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CONDV(J) = TERPOL * (CONDV(1) - CONDV(1-1)) + CONDV(1-1)
ANSATL(J) = TERPOL * (HEATL(1) - HEATL(1-1)) + HEATL(1-1)
ASUR(J) = TERPOL * (SURTEN(1) - SURTEN(1-1)) + SURTEN(1-1)
APRE(J) = TERPOL * (PRESHV(1) - PRESHV(1-1)) + PRESHV(1-1)
ADENSH(J) = TERPOL * (DENSH(1) - DENSH(1-1)) + DENSH(1-1)
AVISC(J) = TERPOL * (VISC(1) - VISC(1-1)) + VISC(1-1)
STCOEF(J) = ((SURTEN(1) - SURTEN(1-1)) / 10.0) / ASUR(J)
STCOEF(J) = STCOEF(J) * 30.0
VPCOEF(J) = (PRESHV(1) - PRESHV(1-1)) / 10.0
TK(J) = 1(J) + 273.0
WRITE(13,201) ANSG(J), BETAJ(J), CONDV(J), ANSATL(J), ASUR(J),
1 APRE(J), ADENSH(J), STCOEF(J), TK(J), AVISC(J)
201 FORMAT(' ',10E12.3)
3 CONTINUE
READ(1,117) (G(I), I=1,N)
117 FORMAT(2X,26F3.2)
READ(1,119) (PP(I), I=1,N)
119 FORMAT(7F11.0)
READ(1,121) (DOLIQ(I), I=1,N)
121 FORMAT(8E10.3)
DO 7 J=1,20
DO 122 I=1,N
P(I) = PP(I) / (1.0 - STCOEF(J) * (TK(J) - 27.0))
DOLIQ(I) = DOLIQ(I) * 0.8545 / AVISC(J)
COLIQ(I) = DOLIQ(I) / 35.0
122 COLIQ(I) = DOLIQ(I) / 60.0
DO 6 I=1,N
6 C(I) = (DOLIQ(I) * (P(I) + P(I+1)) / 2.0) - (P(I-1) * P(I)) / 2.0
C(1) = C(2)
C(N) = C(NN)
DO 66 I=1,N
66 B(I) = C(I) * DOLIQ(I)
WRITE(3,16) J, T(J)
16 FORMAT('0',13,F10.5)
SUM = 0.0

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DATA(J) = 0.217*((TK(J)/273.0)+1.28)
V(J) = (780.0)/(120.0 - APRE(J))
CONDV(J) = ((AMTAL(J) + DATA(J) * V(J))/(5 + TK(J)) * VPCDEF(J)
1*1733.22
WRITE (3,400) CONDV(J)
400 FORMAT ('0',E12.3)
DO 17 1=1,N
OTLQ (1) = B(1) * STCDEF(J) * P(1)
A(1) = 1.0 - XS - Q(1)
IF (1-11300.300,301)
300 RHO(1) = 0.0
GO TO 302
301 RHO(1) = ARHO(J) * EXP ((P(1)*GG)/(5 + TK(J)))
302 DOWAP(1) = (DATA(J) * V(J) * ALPHA * A(1) * GG * RHO(1)) / (5 + TK(J) *
1 C(1))
XW(1) = Q(1)
XA(1) = 1.0 - XS - XW(1)
IF (XW(1) - 0.30) 8,9,9
8 CONDV8 = CONDV(J) * ALOG10 (1.0 + (XW(1)/0.0422))
CONDAV = CONDA + CONDV8
R = 0.035 * XW(1) * (0.333 - 0.035)/0.39
GO TO 200
9 CONDAV = CONDA + CONDV(J)
R = 0.333
200 DELTSA = ((1.0/3.0)*((2.0/(1.0+((CONDS/CONDAV)-1.0)* 0.125))+11.0/(
1 1.0 +((CONDS/CONDAV)- 1.0)* 0.75)))
DELTWA = ((1.0/3.0)* ((2.0/11.0 +((CONDAV/CONDAV)- 1.0)* 0.125)) +
1 (1.0/11.0 +((CONDAV/CONDAV)- 1.0)* 0.75)))
DELTSH = ((1.0/3.0)* ((2.0/11.0 +((CONDS/CONDAV)- 1.0)* 0.125)) +
1 (1.0/11.0 +((CONDS/CONDAV)- 1.0)* 0.75)))
DELTAW = ((1.0/3.0)* ((2.0/11.0 +((CONDAV/CONDAV)- 0.1)* R
1 (1.0/11.0 +((CONDAV/CONDAV)- 1.0)* (1.0-2.0*R))))
IF (XW(1) - 0.05) 10,10,11
10 COND (1) = ((XA(1) * CONDAV + XS * CONDS * DELTSA + XW(1) * CONDAV +
1 DELTWA)/(XA(1) + XS * DELTSA + XW(1) * DELTWA))*1.25

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GO TO 12
11 COND(1) = 1/XW(1)*CONCH + XS* CONDS*DELTSW + XA(1)*CONCAV*DELTAM
1 1/(XW(1) + XS* DELTSW + XA(1)* DELTAM)
IF (1-7)12,500,12
500 DO 501 K = 1,6
COND(K) = COND(1) + X4(K) * (COND(7)-COND(1))/ 0.06
501 WR(1E13,502) COND(K)
502 FORMAT(' COND IS CALCULATED',E20.3)
12 DELTAM = 1.0/(XA(1) + X4(1)* DELTWA + XS * DELTSA)
IF (X4(1) - 0.31) 13,13,14
13 FA = XA(1) + XW(1)
GO TO 12
14 FA = XA(1) + (XA(1)/0.210)* XW(1)
15 DTVAP(1) = DELTAM * BEIA(J) * V(J) * OATH(J) * FA
1 * (RHO(1)/ARKH(1))
DT(1) = DTVAP(1) * DTCLQ(1)
DOW(1) = DOVAP(1) * DOLIQ(1)
WRITE (3,170)G(1),F(1),DOLIQ(1),F(1),DTCLQ(1),DOVAP(1),COND(1),
1 DTVAP(1),DT(1),DOW(1),OATH(J), RHO(1)
170 FORMAT(' ',F6.2, 11E10.3)
17 CONTINUE
DO 20 I = 1,53
IF (1-1)14,14,20
20 CONTINUE
19 F(1) = (COND(1)* DOW(1) + DT(1)*ADENSH(J)*AHEATL(J)*DOVAP(1))/
1 (COND(1)* (E + D(1)) - DT(1) * Q)
G(1) = (COND(1+1) * DOW(1+1)*DT(1+1)* ADENSH(J)*AHEATL(J) *
1 DOVAP(1+1)
1 1/(COND(1+1)* (E + D(1+1)) -DT(1+1)* Q)
SUM = SUM + (G(1) * F(1))/2.01 * (D(1+1) - D(1))
IF (J-1)24,24,403
400 IF (SUM - (J-1))21,22,23
21 IF (SUM + 0.1 - (J-1)) 24,22,22
23 IF (SUM - 0.1 - (J-1)) 22,22,24
24 WRITE (3,25) 1,1, SUM,F(1),G(1)

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29 FORMAT (' .13,3E20.10')

I=I+1

IF (I-51) 27,27,28

27 GO TO 19

22 WRITE (3,26) I+1,5CM,F(1),G(1),C(I+1)

26 FORMAT ('0'.13,3E20.10, F10.3)

28 WRITE (3,29)

29 FORMAT ('H 7HI 11 54')

7 CONTINUE

STOP

END