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ESTIMATION OF EVAPOTRANSPIRATION UNDER VARIABLE SOIL MOISTURE SITUATION

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SUMMARY

Evapotranspiration (ET) is the amount of water transpired by the plant and evaporated from the soil surface. ET from vegetated surface is a function of several process like radiation exchanges, vapour transport and biological growth operating within a system involving the atmosphere, plants and soils. Models reported by Saxton et al (1974 a,b), Ritchie (1972), Van Keulen(1975), Hanks et al.(1969) etc. are typical examples of this integrated process.

The upper zone of the soil, i.e. the unsaturated zone constitutes the medium between the atmosphere and the saturated ground water system. Soil moisture evaporation occurs under unsteady conditions and results in a net loss of water from the soil i.e. it results in drying. This process involves considerable loss of water, especially in arid regions where these losses can amount of 50 percent or more of total precipitation. A technical note has been prepared for estimation of evaporation rate under quasi steady state conditions. For a given relationship between suction head(ψ) and volumetric soil moisture content(θ) and relationship between hydraulic conductivity $K(\theta)$ and θ, the amount of evaporation loss in a soil and the soil moisture variation at different time steps due to evaporation losses has been estimated.

1.0 INTRODUCTION

Evapotranspiration (ET) is a complex process involving many climatic, soil and plant variables. ET is the amount of water transpired by the plant and evaporated from the soil surface. The upper zone of the soil i.e. the unsaturated zone constitutes the medium between the atmosphere and the saturated ground water system. This zone is very important for the physical and chemical process occuring in the soil-plant system.

The predominant soil factors affecting ET are those that affect the amount of water that are available at the soil surface and to the plants. When the surface layer of the soil is wet, evaporation is governed primarily by atmospheric conditions. However, as this layer dries out, the rate of evaporation decreases very rapidly and is greatly influenced by soil properties such as relative humidity of soil air, the diffusion coefficient, the capillary conductivity and the hydraulic conductivity of the surface layer. These soil properties govern the rate at which water, either in liquid or vapour form is transmitted from lower depths in the profile to the surface. The diffusion coefficient of a soil which is a measure of the diffusion of water vapour under a unit vapour pressure gradient depends primarily on the number, size and distribution of air-filled pores in the soil. That is, the nature, texture, granulation and moisture content of a soil. On the other hand, the capillary conductivity, which characterized the rate of capillary flow is largely a function of the soil moisture content, the size, shape and distribution of the pores.

Transpiration by a plant normally creates a diffusion pressure deficit in the plant roots, so that a potential gradient develops across the root surface which is in contact with the moist soil. This process results in a potential gradient in the soil that causes water to move to the plant. More commonly, soil moisture evaporation occurs under unsteady conditions and results in a net loss of water from the soil i.e. it results in drying. This process involves considerable loss of water, especially in arid regions where these losses can amount to 50% or more of total precipitation (Hide, 1954, 1958).

The soil drying process has been observed to occur in three recognizable stages (Fisher 1923; Pearce et al, 1949).

a) an initial constant rate-stage, which occurs early in the process, while the soil is wet and conductive enough to supply water to the site of evaporation. During this stage evaporation rate is controlled by external meteorological conditions such as radiation wind velocity, air humidity etc. The evaporation rate during this stage might also be influenced by soil surface conditions, including surface reflectivity and the possible presence of a mulch. In a dry climate, this stage of evaporation is generally short and may last only a few hours to a few days.

b) an intermediate falling-rate stage, during which the evaporation rate falls progressively below the potential rate. At this stage, the evaporation rate is limited by the rate at which the gradually drying soil profile can deliver moisture toward the evaporation zone. This stage may persist for a much longer period than the first stage.

c) a residual slow rate stage which is established eventually and which may persist at a nearly steady rate for many days. This stage is often called the vapour diffusion stage and can be important where the surface layer is such that it becomes quickly desiccated (e.g. a loose assemblage of clods).

The transition from the first to the second stage is generally a sharp one, the second stage generally blends into the third stage so gradually that the last two cannot be separated so easily.

During the initial stage, the soil surface gradually dries out and soil moisture is drawn upward in response to steepening evaporation-induced gradients. The rate of evaporation can remain nearly constant as long as the moisture gradients toward the surface compensate for the decreasing hydraulic conductivity(resulting from the decrease in water content). From this point, the moisture gradient toward the surface cannot increase any more, and infact, tend to decrease as the soil in depth loses more and more moisture. Since as the evaporation process continues, both the gradients and the conductivities at each depth near the surface are decreasing at the same time, it follows that the flux toward the surface and the evaporation rate decreases as well.

The length of time the initial stage of drying lasts depend upon the intensity of the meteorological factors that determines atmospheric evaporativity, as well as the conductive properties of the soil itself. When external evaporativity is low, the initial, constant-rate stage of drying can persist longer. This fact has led to the hypothesis that an initially high evaporation rate may in the long run reduce cumulative moisture loss to the atmosphere. This hypothesis was raised in a number of Russian papers and cited by Lemon (1956).

The objective of this study was to estimate the evaporation losses under quasi steady-state conditions in a soil system and to estimate the change in soil moisture content due to evaporation losses in different layers at different time steps.

2.0 REVIEW

Evapotranspiration from vegetated surface is a function of several process like radiation exchange, vapour transport and biological growth operating within a system involving the atmosphere, plants and soil. Several researchers have provided good descriptions of these primary variables which determines ET rates (Tanner, 1957; Penman et al., 1967; Campbell, 1977).

Evapotranspiration varies specially as a result of variations in climate, crops or soils. Climatic variables related to ET tend to be conservative and often do not change rapidly over considerable distance. The variation of crops and soils over a region will need to be treated separately. Fro non-irrigated agriculture water availability to the evaporating plant and soil surfaces also often limits ET. Thus the rate of ET is limited to the diffusion rate of soil water to the soil surface and to the plant roots and through the plant system.

The soil-plant-atmosphere system may be represented schematically as shown in Figure 1. ET is the major component of the water budget after precipitation. The interaction of ET with other components like rooting and soil moisture profiles and the dynamic nature of these many components with time becomes readily apparent as the water budget of this system is computed.

Many methods of estimating ET follow a concept of a vertical water budget within a system as shown in Figure 1.

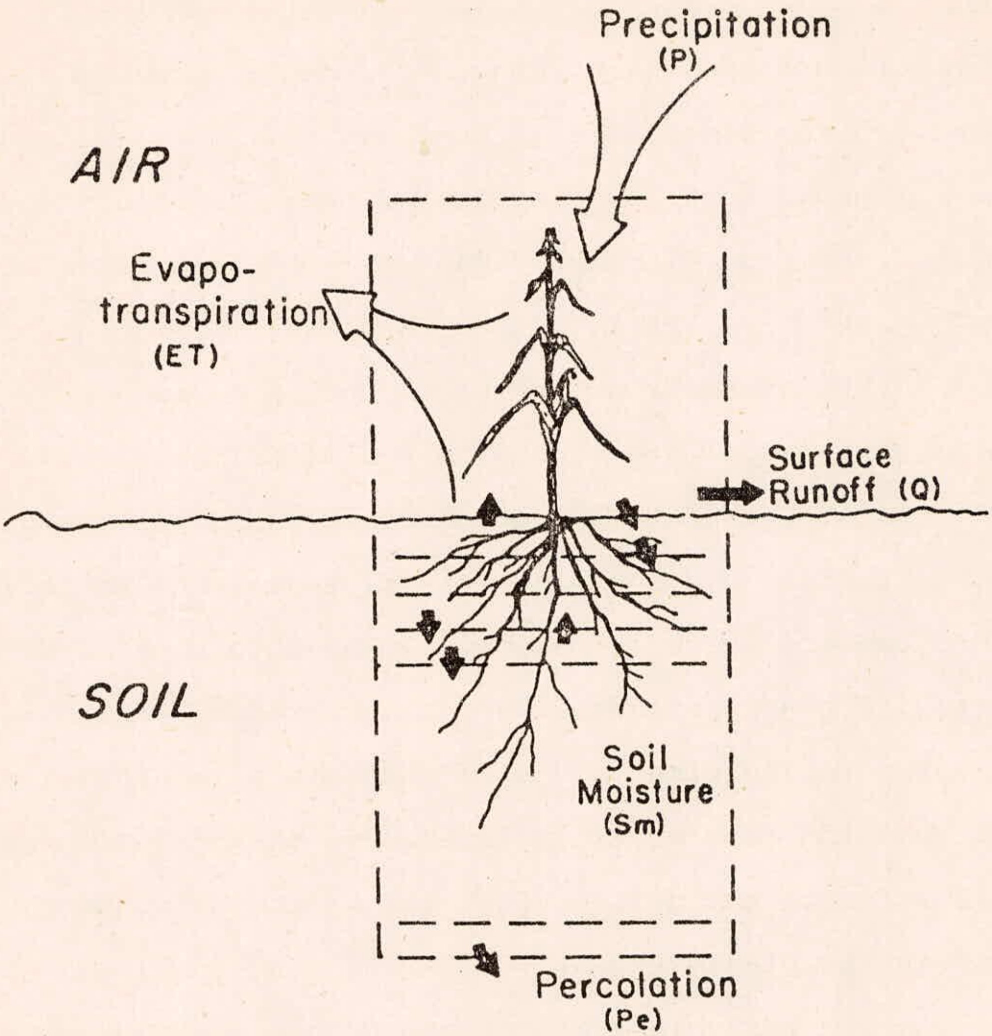

Fig. 1- Schematic of the Soil-Plant- Air System Water Budget

To optimize different schemes of irrigation and water management it is very important to have methods that can estimate evapotransporation under variable soil moisture conditions. In this study, the methods related with plant, water characteristics and soil-water-characteristics have been reviewed for estimation of ET.

2.1 Plant-Water Characteristics

Plant controls a large number of the processes that determine ET rates, either by their use of radient energy or root interaction with available soil water. The effects of plants on ET can be divided into the main categories of a) canopy, b) phenology, c) root distribution, and d) water stress. Many of these interactions of crops with the atmosphere and soil are provided by Monteith (1976), Kramer (1969) and Slatyer (1967).

The water uptake by plants and the mathematical representation of this phenomena have received considerable attention in recent years (Feddes et al., 1976a, 1976b; Hillel and Talpaz,1976; Slack et al.,1977). Saxton et al. (1974b) obtained satisfactory results (evapotranspiration) using nine depth percentage distributions to represent soil water extraction by corn throughout the growing season.

Mustonen and McGuinness (1968) and Baier (1969) summarized several relationships between plant-available soil water and actual/potential transpiration ratio. Some of these relationships were derived for unusual conditions, like deep-rooted crops in sandy soil. Denmead and Shaw (1962) developed basic plant-stress data using a large container study. It is generally agreed that both plant available soil moisture and the atmospheric demand determine what proportion of potential transpiration a plant will achieve. Given a moderate available soil water status, a plant under low atmospheric demand may achieve nearly all of that demand, but the same moisture level and a high atmospheric demand may result in a moisture stress and a significant reduction of transpiration from the potential. A relationship of this process is given in Figure 2, where each curve represents an AET/PET versus plant available soil moisture relationship for a specific atmospheric potential. This approach was applied by Saxton et al. (1974b) to individual 15 cm soil layers. Potential daily demand values associated with curves A through E (low to high PET) were determined for corn and grass crops. Some recent simulations have attempted to treat the movement of water through the soil to the roots and through the roots and canopy as a series of conduits with internal boundary resistance (Slack et al. 1977). During the first congress of Agriculture Engineering held in Geneva, Fleming (1964) published some derived values of soil factors. For using these factors, available soil moisture regime in percentage must be known apart from pan evaporation on day to day basis. The difference between field capacity and wilting point is taken as 100 percent. The table below gives the ratio of actual to potential evapotranspiration as a function of these percentages and the pan evaporation.

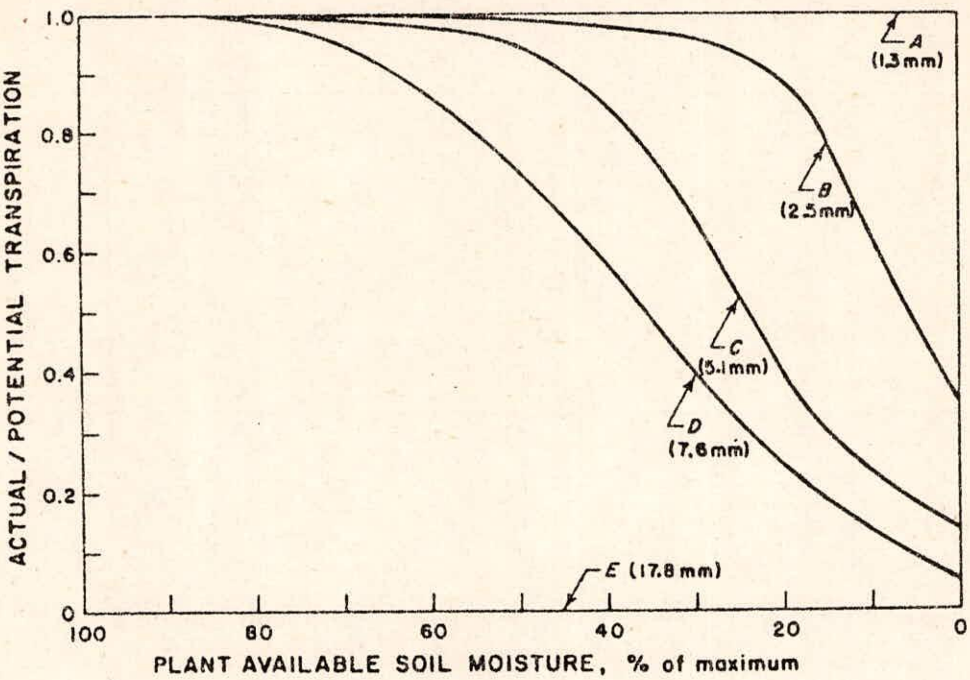

Fig.2 - Moisture Stress Relationships Used to Compute Actual Transpiration. Curves A to E represent Potential ET demand Rates (mm/day) with Values in Parentheses Suggested for Corn in Western Iowa.

Table 1- Fleming's Coefficient of Evapotranspiration

Available soil moisture regime in percentage	Range of free water evaporation in mm			
	0-3.0	3.1-4.0	4.1-6.0	>6.1
100-75	1.0	1.0	1.0	1.0
74-50	1.0	1.0	0.8	0.6
49-25	1.0	0.7	0.5	0.35
24-0	0.5	0.3	0.25	0.15

2.2 Soil-Water-Characteristics

Soil evaporation is often described as occuring at three separate stages beginning with wet soil (Gardner and Hillel 1962; Idso et al. 1974). A relationship incorporating the first two stages for reducing potential soil-water evaporation to actual is shown in Figure 3. (Saxton et al., 1974b). The approach by Hillel (1975, 1977) and Van Bavel and Hillel (1976) provides a more detailed and accurate prediction of soil evaporation but requires significantly more data input and computational time.

A large number of systems have been developed in recent years for actual ET predictions - each has its own requirements and limitations. For application to hydrology, a method should account for climatic, crop and soil variables in some reasonable fashion under a range of moisture regimes.

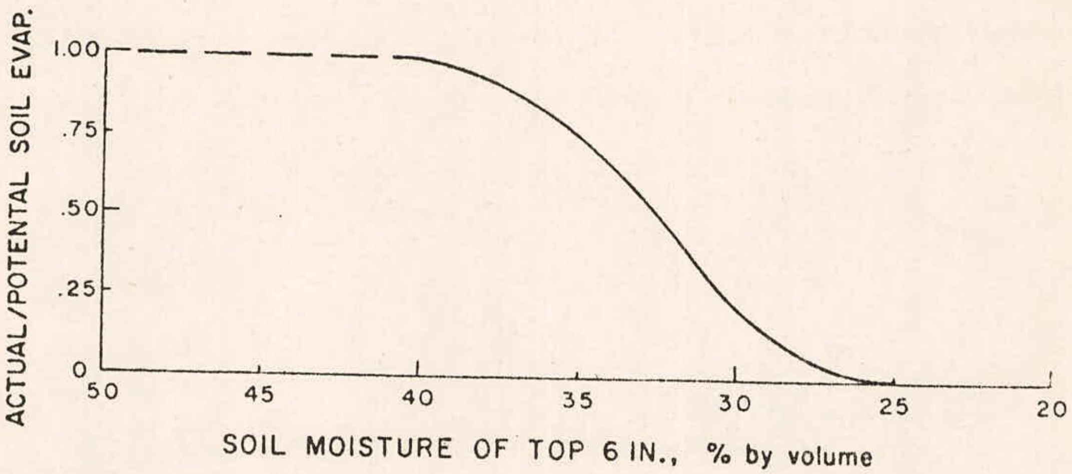

Fig.3 - Soil Evaporation Relationship for Silt loam Soil

Haan (1972) developed a model for simulation of daily ET:

$$E = E_p \ (M/C) \qquad \ldots(1)$$

where,

E = actual ET (mm per day)

E_p = potential ET (mm per day)

M = available soil moisture (mm), and

C = maximum available soil moisture.

Bair and Robertson (1966) reported a somewhat more complex soil moisture budgeting equation for estimation of actual ET;

$$AE_i = \sum_{j=1}^{n} K_j \frac{S_i}{S_j} Z_j \ PE_i \ e^{-w(PE_i - \bar{PE})} \qquad \ldots(2)$$

where,

AE = actual ET (mm per day)

K = coefficient for soil and plant characteristics

S_i = available soil moisture (mm)

S_j = capacity for available water (mm)

Z_j = factor for different types of soil dryness curves

PE_i = potential ET (mm per day)

$\bar{PE}$ = average for month and season, and

w = factor for effects of varying PE rates

The equation (2) is summed for soil layers j for each day i. The coefficient K_j largely depends on plant root in each layer.

A similar single equation approach was applied by Holten et al. (1975):

$$ET = (GI)K\ E_p\ ((S-SA)/S)^x \qquad \ldots(3)$$

where,

ET = actual ET (mm per day)

GI = growth index of crop (percent)

K = ratio of ET to pan evaporation

E_p = pan evaporation (mm per day)

S = total soil porosity (percent)

SA = available soil porosity (percent), and

x = exponent estimated to be 0.10

The GI values reflect crop growth and harvest and are time dependent. The soil storage values S and SA for the root zone approximate water stress.

Soil moisture depletions for irrigation scheduling have been estimated by Jensen et al.(1971) by the relationships

$$E_t = K_c\ E_{tp} \qquad \ldots(4)$$

$$K_c = K_{co}K_a + K_s \qquad \ldots(5)$$

$$K_a = l_n\ (A_{w+1})/l_n\ 101$$

where,

E_t = actual ET (mm per day)

E_{tp} = potential ET (mm per day)

K_c = coefficient representing the combined effects of the resistance of water movement from the soil to the various evaporating surfaces.

K_{co} = mean crop coefficient based on experimental data (soil moisture not limiting)

A_w = remaining available soil moisture, and

K_s = coefficient.

Ritchie (1972) developed a series of equations to represent actual ET then separately calculated soil and plant evaporation. The potential soil evaporation (first-stage drying) was determined by the relationship

$$E = (\Delta/\Delta + \gamma)) \cdot R_n \exp (-0.398 \, LAI) \quad \ldots (6)$$

where

E = potential soil evaporation (mm per day)

Δ = slope of the saturation vapor pressure curve at mean air temperature

γ = phychrometric constant (m bar / ^{o}c)

R_n = net radiation (cal/cm^2/day), and

LAI=leaf area index.

The soil evaporation proceeds at potential rate until soil water transport restraicts the water quantity. The amount of drying before this occurs was determined for each soil. A second stage soil evaporation was computed by the relation

$$E = \alpha t^{\frac{1}{2}} \quad \ldots (7)$$

where,

E = soil evaporation (mm per day)

t = time (days),and

α = a coefficeint to be determined experimentally. Plant transpiration in the Ritchie model was represented by the empirical relation

$$E_p = E_o \ (-0.21 + 0.70 \ LAI^{\frac{1}{2}}) \qquad \text{...(8)}$$

where,

E_p = transpiration (mm per day)

E_o = potential ET (mm per day), and

LAI= leaf area index.

This model was later adopted to field conditions where soil water was limiting. The relationship between E_p and E_o was in the form

$$E_p = E_o \ (1 - (t/t_1)^{\frac{1}{2}}) \qquad \text{...(9)}$$

where,

E_p = transpiration (mm per day)

E_0 = monthly average potential ET (mm per day)

t = time after lower limit of soil water content for potential ET (days) and

t_1 = time to deplete remaining available water after t begins (days).

A comprehensive model to compute daily actual ET was developed and reported by Saxton et al. (1974b). This model, is shown schematically in Figure 4. The more emphasis has been

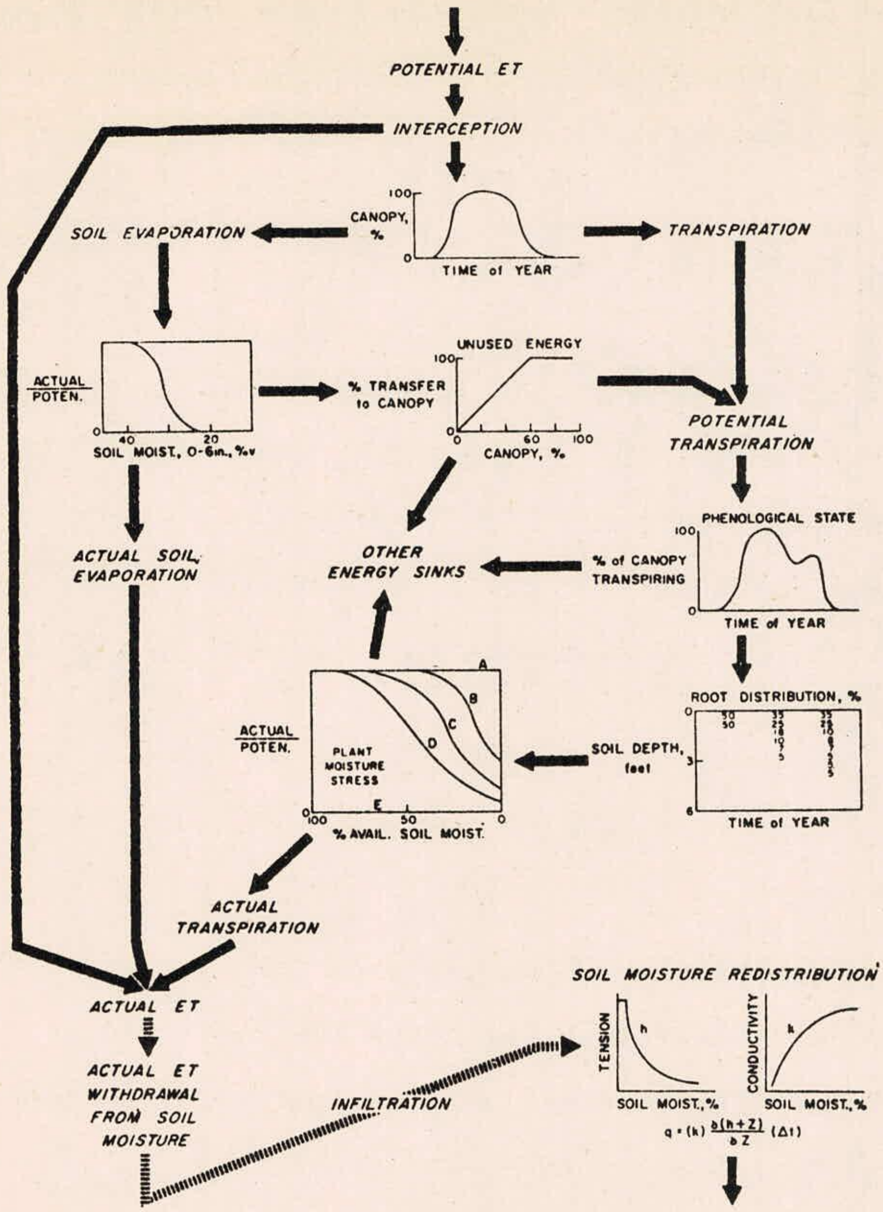

Fig.4 - Schematic Calculation Sequence for Computing Daily Actual ET and Soil Water Changes (Saxton et al., 1974b)

given on graphical representation of various parameters. The amount of interception evaporation, soil evaporation and plant transpiration are combined to provide daily actual ET estimates. As shown in Figure 4 intercepted water at the plant and soil surface is considered to have first use of potential ET energy. The remaining potential ET is divided between soil water evaporation or plant transpiration. Actual soil evaporation is the potential limited by soil water content at the surface except in the very wet range thus representing the traditional two stage drying sequence. For dry soil with a plant conopy a percent of the unused soil evaporation potential is returned to the plant transpiration potential to account for re-radiated energy from the heated soil and air. Actual transpiration is computed through sequential consideration of plant phenology to describe the transpirability of the existing canopy a root distribution to reflect where in the soil profile, the plant is attempting to obtain water, and a water stress relationship which is applied to each soil layer and is a function of plant available water of that soil layer and the atmospheric demand on the plant. The soil water is adjusted by abstracting the daily actual ET from each rooting layer adding daily infiltration computed from daily precipitation minus measured or estimated runoff and estimating soil water redistribution and percolation by a Darcy type unsaturated flow computation. Like most ET methods this method represents a single crop and soil combination for the computed vertical water balance thus watersheds with several crops and/or soils would require multiple applications per daily calculations or average crop and soil representations.

Several models have been developed which describes the ET processes within the soil plant atmosphere system. The soil plant atmosphere model (SPAM) described by Lemon et al. (1973) treats the ET and plant growth characteristics in detail. Van Bavel and Ahmed (1976) described a model of the soil moisture flow and root uptake programmed in CSMP.

2.3 Soil Moisture Hysteries

Jackson (1973) and Jackson et al. (1973) showed that the surface zone soil moisture content fluctuates in a manner corresponding to the diurnal fluctuation of evaporativity, i.e. the soil surface dries during daytime and tends to rewet during nighttime, apparently by sorption from the moisture layer beneath. This phenomenon has been shown by Jackson in Figure 5. The amplitude of the fluctuation decreased with depth and time, and the daily maxima and minima exhibited an increasing phase lag at greater depths.

Hillel (1975, 1976a) developed a dynamic simulation model capable of monitoring the evaporation process continuously through repeated cycles of increasing and decreasing evaporativity. Hillel developed this model in an attempt to clarify the extent to which the diurnal pattern of evaporativity may influence the overall quantity of evaporation and the resulting soil moisture distribution in space and time.

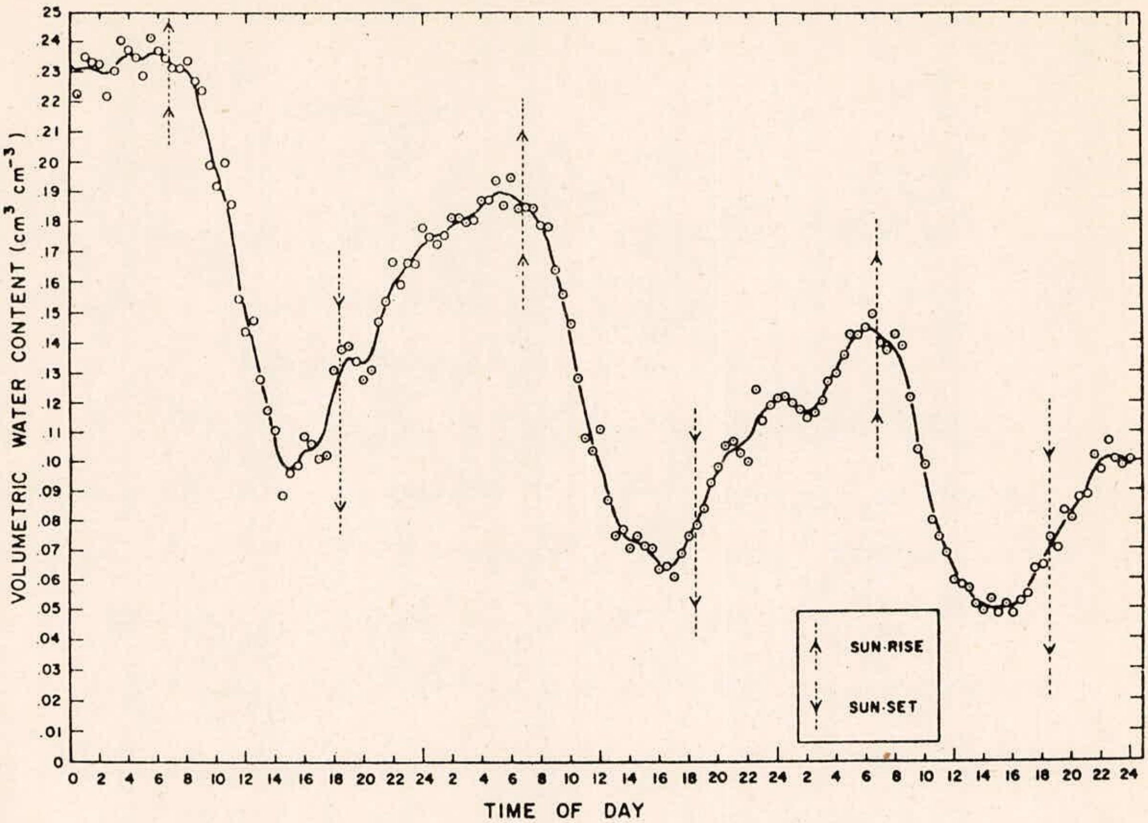

Fig.5 - Time Course of Changing Volumetric Wetness at the Surface (0-0.5 cm) of a Loam Soil During 3 days of Drying , 5-7 days after Irrigation (After Jackson, 1973).

Soil moisture hysteresis was first studied by Haines (1930) and later by Miller and Miller (1956) Youngs (1960b) & many others. More recently several investigators (Rubin, 1966; Bresler, 1969) have studied the effect of hysteresis on soil water dynamics. In principle the hysteresis phenomenon causes a sorbing zone of soil to approach potential equilibrium with a desorbing zone of the same soil while the former is at a lower moisture content, and hence at a lower value of hydraulic conductivity.

A similar study of evaporation from soils of various textures (Hillel and Van Bavel, 1976) showed that differences in soil hydraulic properties can strongly influence cumulative evaporation, with coarse textured soils (sand) evaporating the least and fine textured soils (clay) evaporating the most, under both steady and cyclic evaporativity regimes as shown in Figure 6.

2.4 Tanks and Lysimeters

Evapotranspiration rates are determined from lysimeters by measuring the water loss from tanks (lysimeters) on which plants are grown. ET is determined on lysimeters from measurements of the amount of water applied, outflow and changes in the soil moisture in the tank. Changes in soil moisture level may be obtained by direct moisture sampling (gravimetric, neutron probe etc). In the weighing type lysimeter, the tank is mounted on a self recording scale. In the hydraulic type of lysimeter, the tank is floated and changes in weight are recorded as

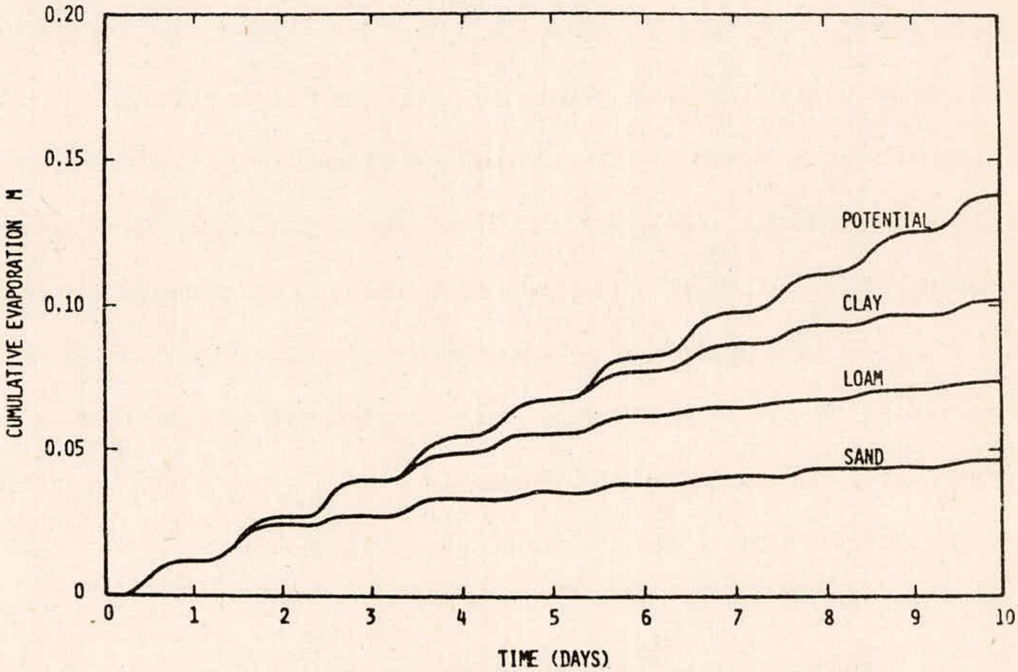

Fig.6 - Cumulative Evaporation During Simultaneous Drainage and Evaporation Form Initially Saturated Uniform Profiles of Sand, Silt and Clay (After Hillel and Van Bavel, 1976).

pressure changes by a manometer.

Some controversy has arised over the use of lysimetric data for determining the amount of evapotranspiration. The differences may exist between the lysimeter and natural conditions in the soil profile, soil moisture regime, plant rooting characteristics, methods of water application and the net energy exchange. However, it is generally conceded that if the installation meet certain minimum standards, they will provide reasonably reliable information on evapotranspiration of plants over short time periods (Linsley et al., 1949 Black et al. 1969) presentated data from a hydraulic balance type of lysimeter in a field experiment. Water flow is represented indirectly through a storage vs drainage relation for the entire lysimeter. Plant water extraction was considered to take place within the lysimeter.

2.5 E_a/E_p and Soil Moisture Deficit Models

Actual evapotranspiration (E_a) is mainly controlled by potential evapotranspiration (E_p), plant factors, stage of growth and stomatal regulation as well as by available water and hydraulic conductivity-moisture tension relationship. E_a reached E_p if conditions permit. E_p can be estimated from meteorological parameters.

For estimation of E_a, some models with limited or approximate applicability are available, relating to E_a/E_p to soil moisture status as shown in Figures 7, 8 and 9 respectively. Figure 7 gives some postulated course of

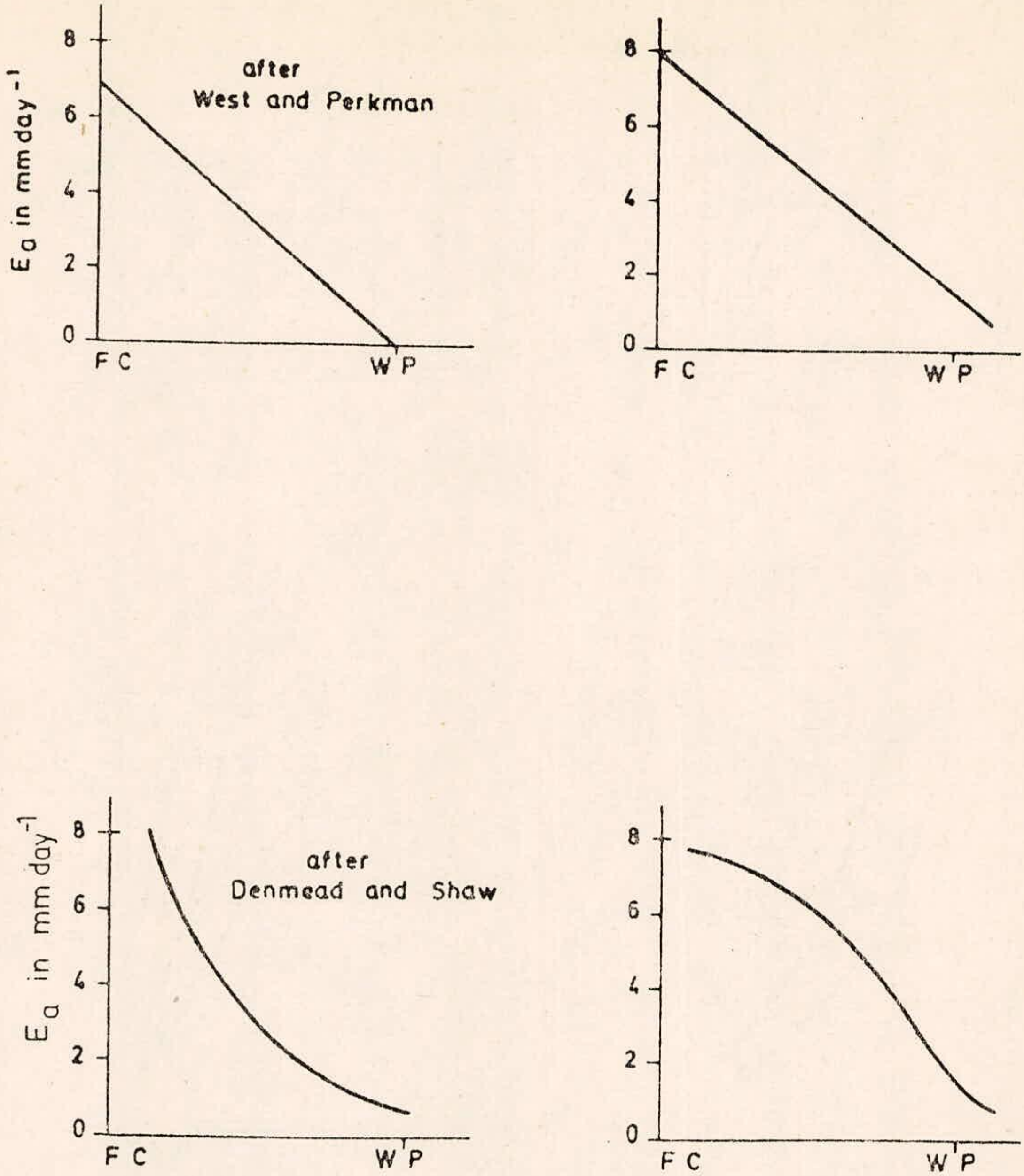

Fig.7 - Some Postulated Curves for Actual Evapotranspiration When the Soil Dries out.

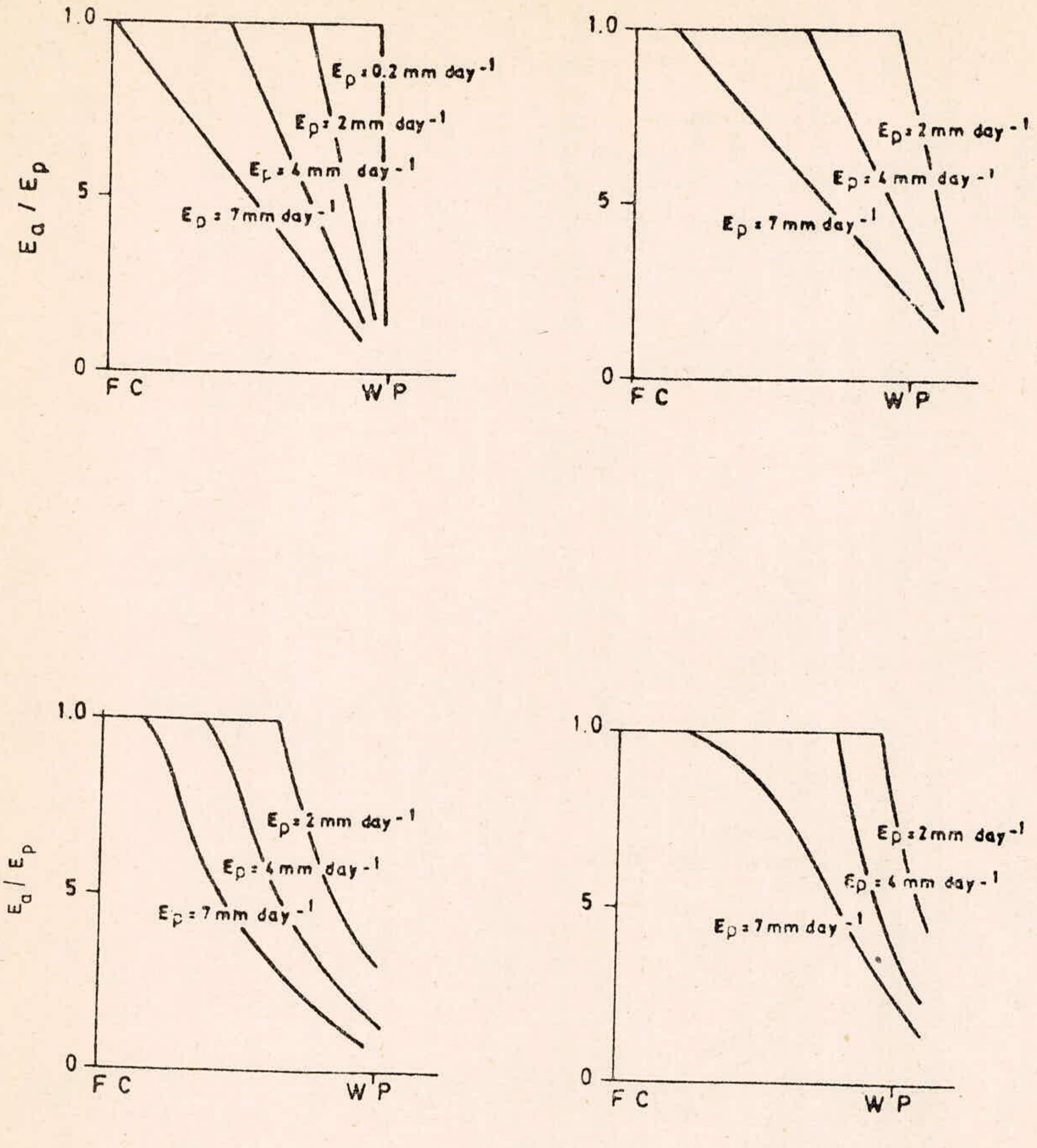

Fig.8 - Reduction of E_a/E_p with Soil Moisture Depletion

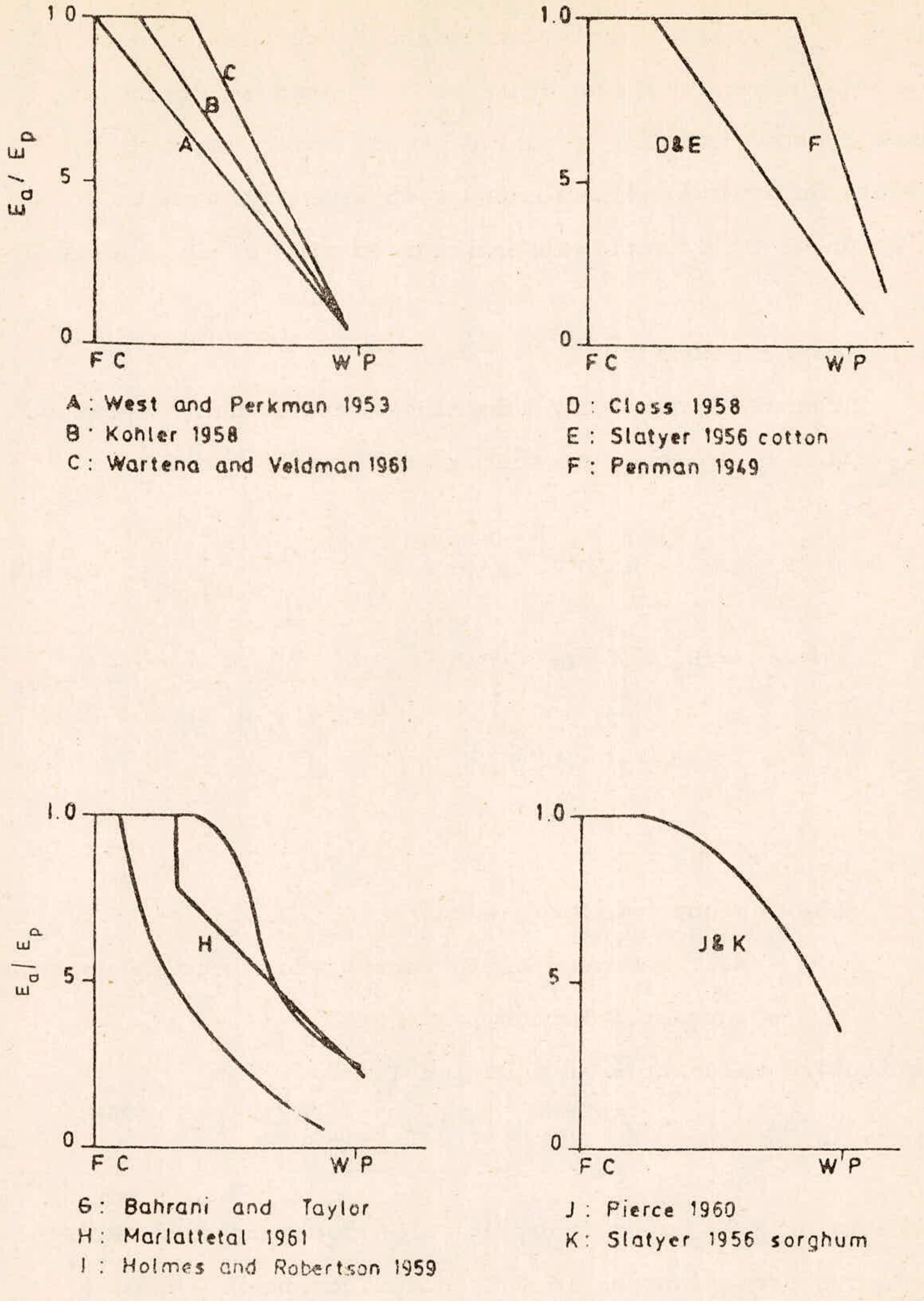

Fig.9 - Reduction of E_a/E_p with Soil Moisture Depletion.

actual evapotranspiration when the soil dries out. Figure 8 shows E_a/E_p curves derived from the E_a curves (Figure 7) by substituting the value of E_p as indicated in Figure 8. Figure 9 shows actual (reported) E_a/E_p ratio and soil moisture deficit which indicates some similarity to the variation of E_a/E_p with soil moisture presented in Figure 8.

2.5.1 Minhas Model

Minhas et al. (1974) developed relationship between E_a/E_p and soil moisture deficit as follows:

$$E_a/E_p = (1 - e^{-\gamma(PAW)})/(B + e^{-\gamma(PAW)}) \qquad \text{...(10)}$$

For $E_a = E_p$ and PAW = PAW_m

$$B = 1 - 2\, e^{-\gamma(PAW_m)} \qquad \text{...(11)}$$

where,

PAW = plant available water

PAW_m = water between field capacity and wilting point, &

γ = parameter fitted from data

Substituting value of B in equation (10):

$$E_a/E_p = (1 - e^{-\gamma(PAW)})/(\, 1-2\, e^{-\gamma(PAW_m)} + e^{-\gamma(PAW)}) \qquad \text{...(12)}$$

The ratio of E_a/E_p is a function of crop weighting factor, increasing from planting to full cover, constant until start of sensscence, then decreasing to harvest. Parameters were fitted from wheat crop data from Delhi by Minhas et al. (1974) and he tested against results from alfalfa crop data of Mustonen and McGuinness (1968). The effect of the ratio of E_a/E_p on available soil water is shown in Figure 10 (by Minhas et al.,1974).

2.5.2 Penman Model

Penman (1949) developed the concept of a 'root constant' and a drying curve to represent the reduction in actual evapotranspiration below the potential rate as a function of soil moisture. The developed relationship can be plotted as a curve of actual soil moisture deficit against potential soil moisture deficit, as shown in Figure 11. This figure forms the basis of a calculation procedure, as follows:

1) locate the current soil moisture deficit on actual soil moisture deficit axis (SMD = 0 at field capacity)

2) If E_p > Rain, assume Rain is evaporated at the potential rate and (E_p - Rain) goes to increase potential soil moisture deficit.

3) From drying curve, read across the corresponding available soil moisture deficit (Figure 11).

4) If ASMD < PSMD (i.e. location on drying curve is beyond Root Constant, RC)

then

E_a = Rain + Δ(ASMD)

and

$E_a/E_p < 1$

if ASMD=PSMD, then

$E_a = E_p$

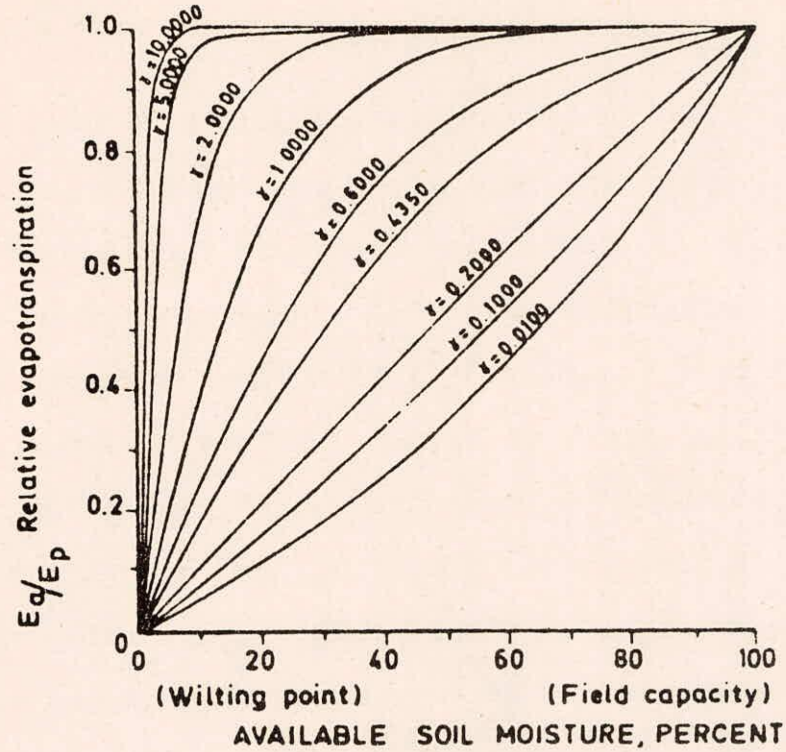

Fig .10 - Family of Curves for Estimating the Relationship Between E_a/E_p and Available Soil Moisture

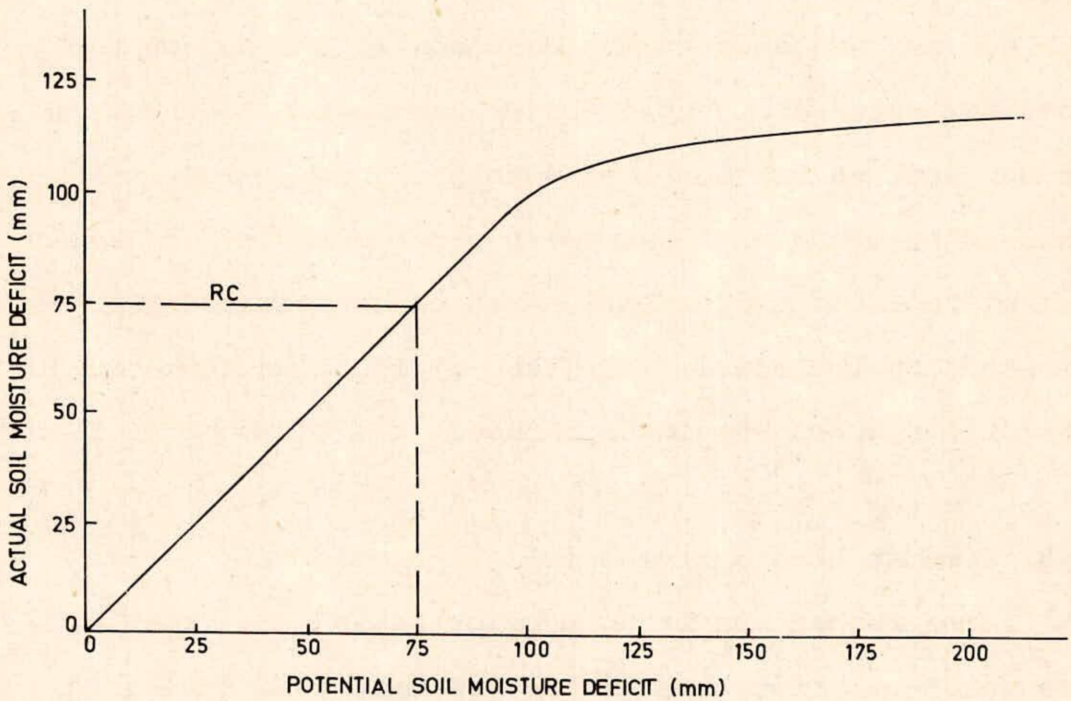

Fig.11 - Relationship Between Actual Soil Moisture Deficit and Potential Soil Moisture Deficit.

5) If Rain $> E_p$, assume $E_a = E_p$ and ASMD is reduced by (Rain $- E_p$). If ASMD < 0, the excess moisture above field capacity is assumed to go as runoff or drainage.

The Penman method was adopted by U.K. Meteorological Office and used as the basis of national service to forecast actual evaporation and soil moisture deficit from 1962. A detailed land use description was used with root constant values estimated from agricultural experience. Changes in land use such as harvest time were included. Comparison of model performance with catchment water balances and lysimeter data indicated that percolation was underestimated (E_a over estimated) by the model, with the implication that the root constant for grass should be reduced from 75 mm to 38-35 mm.

2.5.3 Penman alternative model

The Penman model can be plotted as shown in Figure 12 and several relationships can be postulated in an equivalent form:

The extremes are curves A and B - which show little plant response to dryness and are not generally applicable. Curve C for which $E_a/E_p \propto$ SMD is applicable to sandy loam soil for high evaporative demand). Most curves have 3 stages

i) $E_a = E_p$ until moisture is available at the plant roots to meet the evaporation demand.

ii) E_a/E_p decreases rapidly until the roots are no longer able to withdraw moisture.

iii) a low loss rate associated with diffusion from the

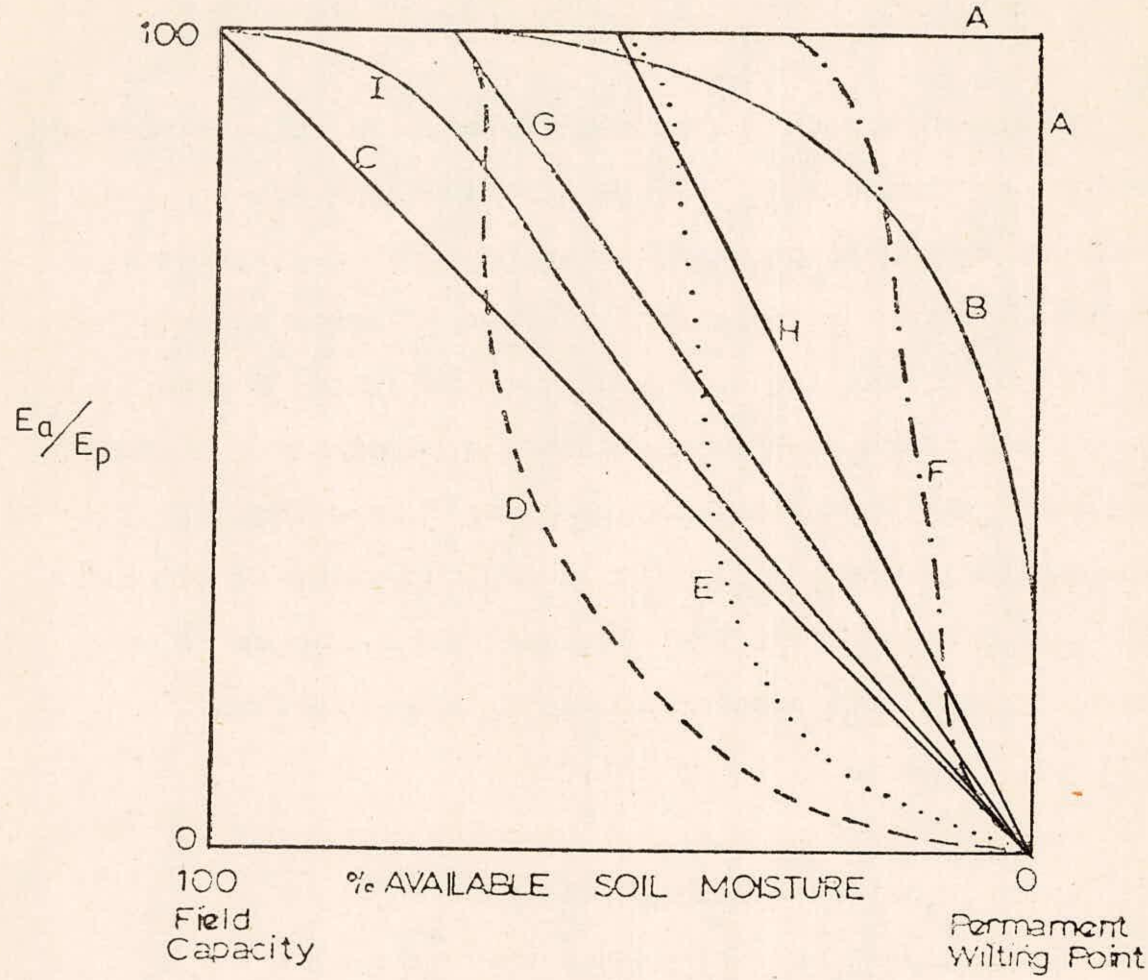

Fig.12 - Postulated Relationship for E_a/E_p as a function of available soil Moisture

soil to the atmosphere.

The form of the curve will be expected to depend on the soil type and plant rooting characteristics. Clay soil will be nearer to D curve while sandy soil will follow F curve.

Alternative models have been formulated using different approximations to the E_a/E_p vs SMD relationship. More complex formulations have been proposed in which water is extracted simultaneously from different soil levels or zones in relation to the rate of E_p and the available soil moisture in each zone (Baier and Robertson, 1966). Each zone is permitted a different form of the E_a/E_p vs SMD relationship, and the effect of E_p rate on the ratio of E_a/E_p is considered. However, in general, there are too few data for the adequate specification of parameter for the more complex models without extensive calibration.

2.5.4 Holmes E_a/E_p vs available moisture model

Holmes (1961) has shown that the ratio of E_a to E_p changes as the soil dries out and the shape of the curve differs both with the type of soil and the drying rate (Figure 13) because plants cannot utilize the total available water (water held in the range between field capacity, FC, and permanent wilting point, PWP, within the root zone).

Unfortunately, it is impossible to define quantitatively shape of curve in terms of the factors affecting it. Undoubtedly the shape reflects the influence of such factor as a) the ability of the media to transmit water in the unsaturated state, b)

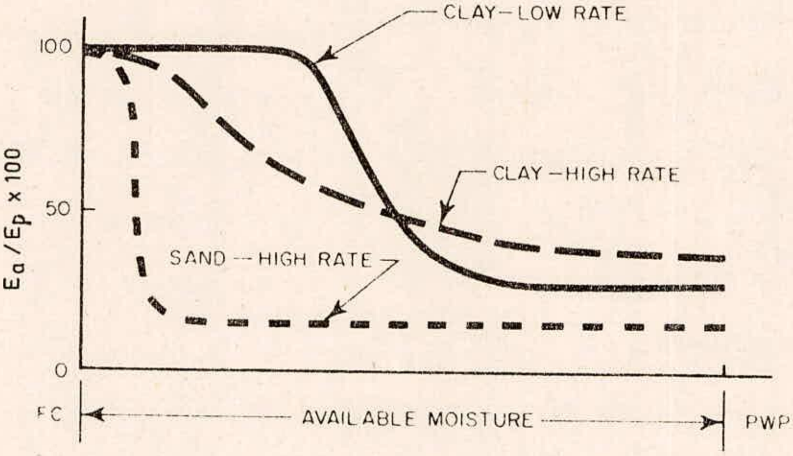

Fig.13 - Schematic Drying Curves Showing Ratio of E_a/E_p plotted with Soil Moisture Content for Different Soils and Drying Rates.

ability of the plant to absorb water and convey it from the roots to leaves and c) the root development and distribution pattern.

From the above discussion it is obvious that an accurate description of the moisture status of a soil as it affects evapotranspiration is not simple; however, it is equally evident that potantial rates must be modulated to account for such factors as soil moisture stress if they are to provide reasonable estimates of E_a rates and the soil moisture status. For example the soil moisture budget proposed by Holmes (1961) divides the soil root zone into two zones, an upper and lower zone as to water availability. He assumed that a) all moisture from the upper zone is evapotranspired by the potential rate b) moisture in the lower zone is withdrawn at a decreasing rate depending on the amount of moisture in the zone (that E_p values are modulated using a factor of less than 1, Holmes using the following factor - first 25 percent of soil moisture evapotranspiration at 0.50 E_p, second 25 percent at 0.20 E_p, next 25 percent at 0.10 E_p, and last 25 percent at 0.05 E_p and c) available water is withdrawn from uppermost moist soil zone before extraction occurs from the lower zone.

3.0 PROBLEM DEFINITION

A shallow ground water table exists at a constant depth 'z' below the ground surface. The soil moisture profile has reached a static equilibrium and upward movement of capillary flow has ceased. At such condition the sum of the suction head and the elevation head at all points above the water table will be a constant. Let the moisture content of the soil at the ground surface change due to atmospheric impact and let $\theta_b(n)$ be the moisture content of the top few centimetre depth of soil which have been recorded at various time, n. Let $\theta_b(n)$ values be less than the corresponding static equilibrium moisture content. Such a specific situation will occur during a dry spell. For the initial and boundary conditions prescribed above it is required to find the upward movement of soil moisture from the water table to the ground surface and the evaporation rate.

4.0 METHODOLOGY

The following assumptions have been made in the analysis:

i) The soil above the water table is homogeneous

ii) The soil has been divided into several zones of equal thickness. Within each zones the soil moisture does not vary with space,

iii) The soil moisture of the top layer and of the other layers do not increase at any time(i.e. there is no rainfall),

iv) The unsteady state has been approximated by quasi steady state conditions. The time span has been discretised into small time steps. Within a time step the suction heads at the top and bottom of a soil layer and the hydraulic conductivity do not change with time. Therefore, the rate of transfer of soil moisture from one layer to the layer above it also do not vary during a time step.

The soil moisture flow from $(i+1)^{th}$ layer to the i^{th} layer during the j^{th} unit time step under the quasi steady state assumption is given by

$$q(i+1, i, j) = K\left(\frac{\theta(i,j) + \theta(i+1,j)}{2}\right) . \frac{\psi(i,j) - \psi(i+1,j)}{z_i - z_{i+1}} - 1 \quad \ldots (13)$$

$K\left(\frac{\theta(i,j) + \theta(i+1,j)}{2}\right)$ represents the hydraulic conductivity at the interface of i^{th} and $i+1^{th}$ layers, $\Psi(i,j)$ is the suction head of i^{th} layer at j^{th} unit time which will be function of θ (i,j). The elevation head z is measured upward from the water table. The rate of soil moisture flow to the top layer from the layer under neath during j^{th} unit time step which is the evaporation rate during the j^{th} unit time step, will be given by

$$q(2,1,j) = K\left(\frac{\theta_b(j) + \theta(2,j)}{2}\right) \times \frac{\Psi(\theta_b(j)) - \Psi(\theta(2,j))}{z_1 - z_2} - 1 \qquad \ldots(14)$$

The soil moisture of the $i+1^{th}$ layer during the $(j+1)^{th}$ unit time step will be given by

$$\theta(i+1,j+1) = \theta(i+1,j) - \frac{q(i+1,i,j)}{\Delta z} + \frac{q(i+2,i+1,j)}{\Delta z} \qquad \ldots(15)$$

5.0 APPLICATION AND RESULTS

The soil data required for estimation of evaporation losses have been obtained from the experimental results of Sonu(1973). The variation of suction head(Ψ) with the volumetric soil moisture content (θ) and variation of hydraulic conductivity $K(\theta)$ with (θ) are shown in Figure 14 and 15 respectively for podure sand. The evaporation losses have been estimated in a soil system where the water table lies 50 cm below the ground surface.

The soil system is divided into 5 layers of thickness 10 cm each. The following steps were used for estimation of evaporation losses:

1) The initial soil moisture at centre of the each layers were measured and given in Table 2. These soil moistures corresponds to the static equilibrium conditions. If the soil moisture of the top layer does not differ from 0.175 which is equal to static equilibrium moisture content, there would not be any evaporation loss from the water table.

2) The soil moisture of the top layer is set to be 0.15 (less than the equilibrium soil moisture).

3) At time step 1, the soil moisture will flow from 2nd layer to top layer. The initial soil moisture of 2nd layer is 0.22. The amount of evaporation has been estimated by using equation(13) and it was found to be 0.414×10^{-2} cm. This is the evaporation loss during first time step.

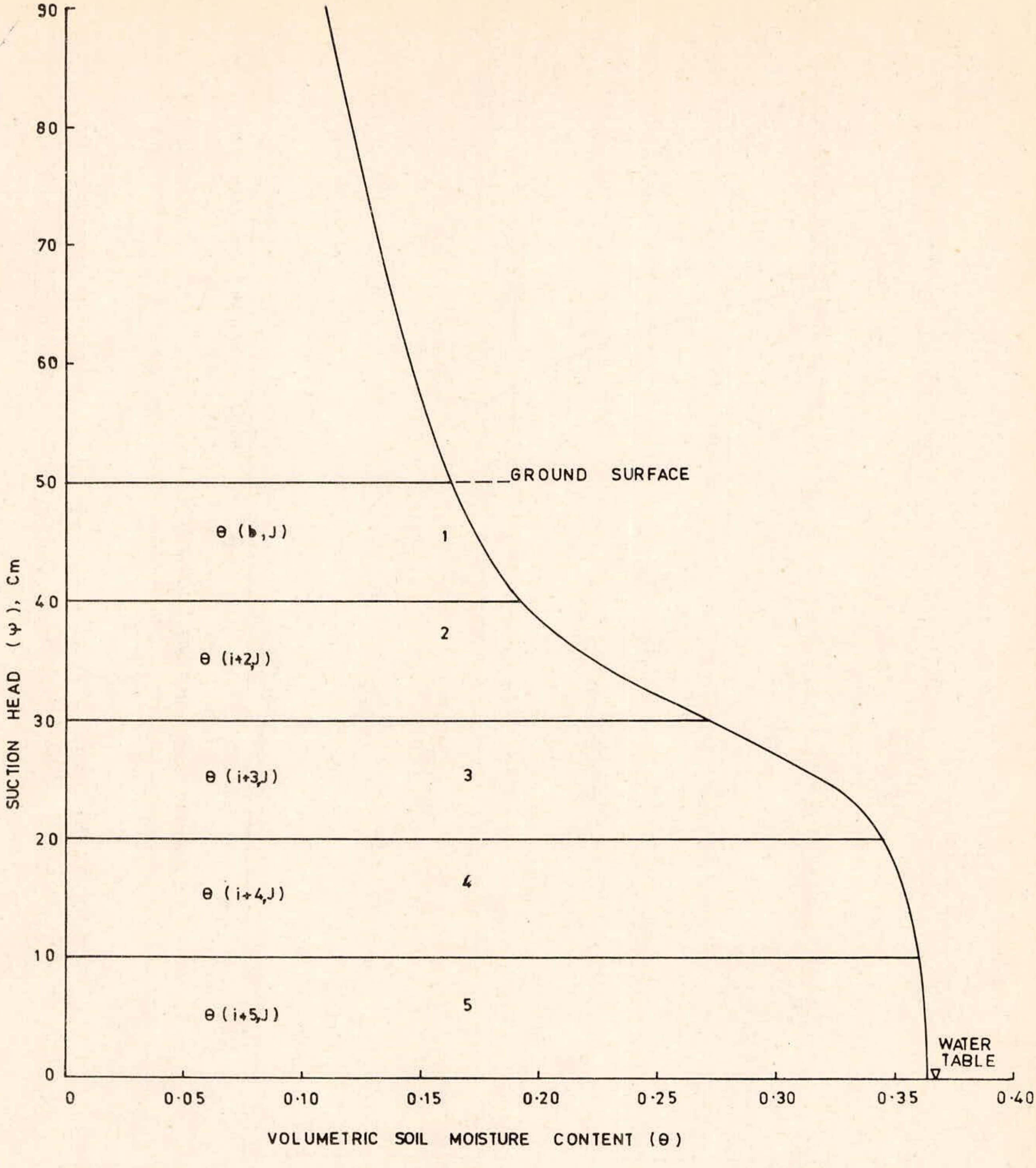

Fig.14 - Variation of Suction Head (Ψ) with Volumetric Soil Moisture Content (Θ) for Poudre Sand

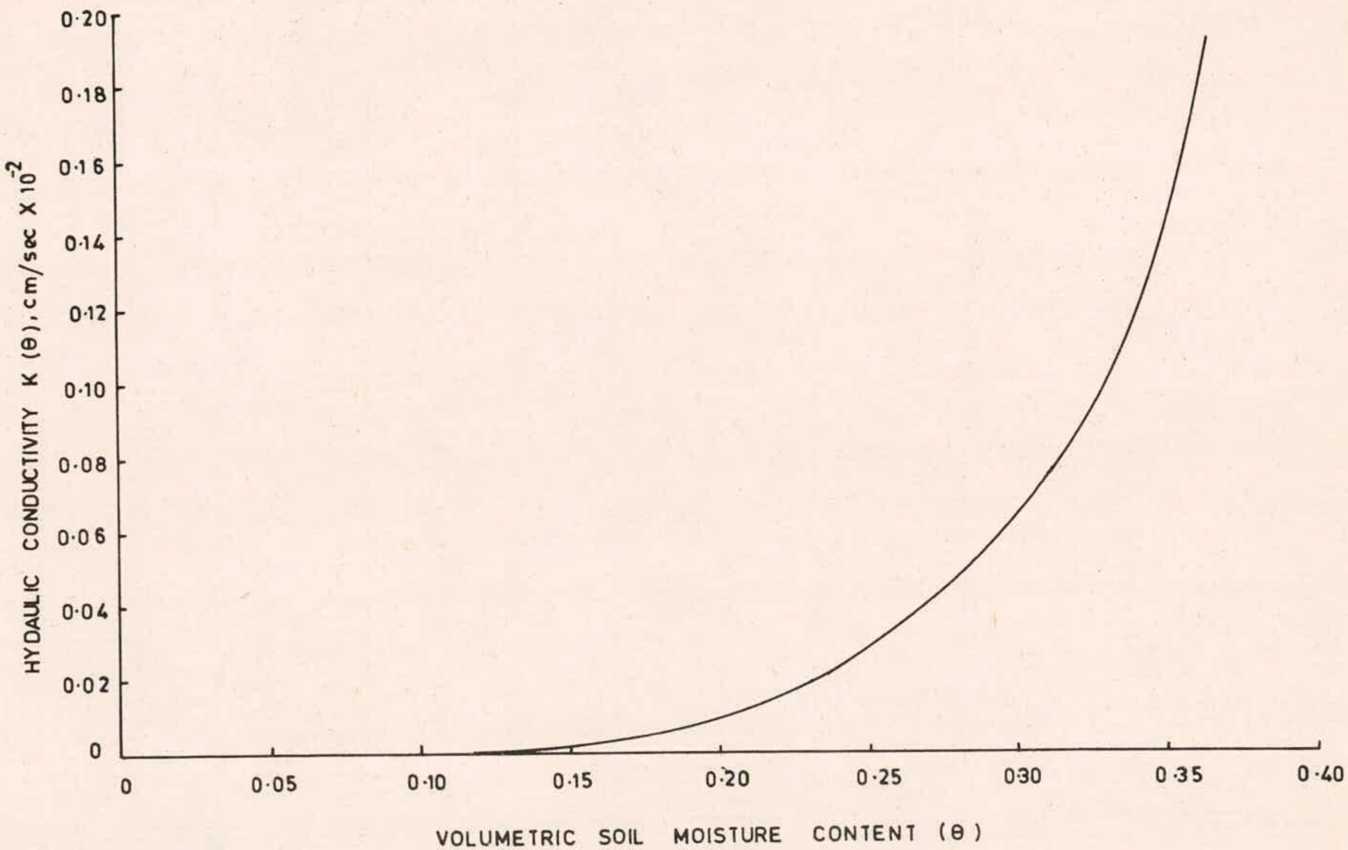

Fig.15 - Variation of Hydraulic Conductivity K(θ) with Volumetric Soil Moisture Content (θ) for Podure Sand.

TABLE 2. ESTIMATION OF EVAPORATION AND SOIL MOISTURE IN A LAYER SOIL AT DIFFERENT TIME STEPS

Layer Nos.	1		2		3		4		5	
	Soil Mois.	Evapora- tion cm	Soil Mois.	Evapora- tion cm	Soil Mois.	Evapora- tion cm	Soil Mois.	Evapora- tion cm	Soil Moisture	Evaporation cm
Layer 1	0.1500		0.1500		0.1500		0.1500		0.1500	
		$0,4140x10^{-2}$		$0.3210x10^{-2}$		$0.3180x10^{-2}$		$0.280 8x10^{-2}$		$0.2754x10^{-2}$
Layer 2	0.2200		0.2158		0.2144		0.2130		0.2119	
		-		$0.1824x10^{-2}$		$0.1752x10^{-2}$		$0.1728x10^{-2}$		$0.1944x10^{-2}$
Layer 3	0.3200		0.3200		0.3181		0.3170		0.3166	
		-		-		$0.0648x10^{-2}$		$0.1308x10^{-2}$		$0.1635x10^{-2}$
Layer 4	0.3525		0.3525		0.3525		0.3518		0.3515	
		-		-				$0.0972x10^{-2}$		$0.1458x10^{-2}$
Layer 5	0.3630		0.3630		0.3630		0.3630		0.3620	

4) During time step 2, the soil moisture of 2nd layer will decrease by an amount 0.414×10^{-2} cm. At this stage, the soil moisture flow will take place from 3rd layer to 2nd layer and from 2nd layer to top layer respectively. The flow rate from 3rd layer to 2nd layer was 0.1824×10^{-2} cm and from 2nd layer to top layer was 0.321×10^{-2} cm.

5) Similarly for different time steps, the flow rates have been estimated from each layer. At each time step, the soil moisture contents of the layers have been updated by using equation(15) and the results are shown in Table 2.

6) It was observed that the flow rate is greater from the top layers as compared to the layers lies underneath it. The evaporation loss decreases with the increase in time step.

6.0 CONCLUSION

Assuming a quasi steady-state condition, the movement of soil moisture from shallow water table has been analysed for a soil for which the relationship between the suction head(Ψ) and volumetric soil moisture content(Θ) and relationship between hydraulic conductivity $K(\theta)$ and volumetric soil moisture content (θ) are available.

This methodology can be conveniently used for prediction of evaporation loss if the soil moisture of the top layer has been recorded and a initial soil moisture profile is known. A sample calculation has been shown for an assumed value of the soil moisture content of the surface layer.

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