Workshop on Ground Water Balance
(Jan. 23-26, 1989)

EVAPOTRANSPIRATION LOSSES

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INTRODUCTION

Evapotranspiration (ET) is the combined process by which water is transformed from the earth surface to the atmosphere. It includes evaporation of liquid from soil and plant surface plus transpiration of liquid water through plant tissue.

ET is the major component of the water budget after precipitation. The interaction of ET with other component like rooting and soil moisture profiles and dynamic nature of these many components with time make evaluation more difficult. The amount of liquid water and the energy to vaporize it will vary both in space and time over the watershed surface.

ET flux moves large quantities of water from the soil back to atmosphere. In humid zones, about 750 to 900 mm/yr of water is vaporized. In sub humid areas, 550 to 700 mm/yr commonly evaporate from vegetated surfaces. In drier regions where evaporative demands are even higher, most if not all of the precipitation is returned to the atmosphere through this process. Accurate prediction of ET is required for hydrologic models e.g. U.S.G.S. model, TVA model, USDAHL model, CREAMS model, SHE model etc.

ET amount varies with the time and space because temperature, precipitation, irrigation practices, humidity, wind velocity varies with time and space.

DEFINITIONS

Definitions of some of the important terms which are referred frequently are given below :

Evaporation

It is the process during which water changes to vapour. This is one of the basic components of hydrologic cycle by which water changes to vapour through the absorption of heat energy.

Transpiration :

It is the process by which water vapour leaves the living plant body and enters the atmosphere. It involves continuous movement of water from the soil into the roots, through the stem and out through the leaves to the atmosphere.

Evapotranspiration (ET) :

It is the process through which water is transferred from the earths surface to the atmosphere. It is the amount of water evaporated from soil and plant surfaces and transpiration of liquid water through plant tissues. It is generally expressed as equivalent depth of water per unit area as mm or cm.

Potential Evapotranspiration (PET)

The ideas of Potential Evapotranspiration was first put forth by Thornthwaite in 1944 after two decades of study of the moisture factor in climate. The water loss taking place from extensive vegetation cover under the ideal condition of continuously adequate soil moisture is termed as Potential Evapotranspiration (PE). It is a theoretical moisture loss which applies to any place but is only attained in reality where there is no shortage of moisture and where other requirements are met.

Actual Evapotranspiration (AE_T)

It is necessary to distinguish between evapotranspiratic under conditions of variable soil moisture and potential evapotranspiration. Actual evapotranspiration represents the amount of water loss actually taking place in the form of evaporation and transpiration. When precipitation is greater than potential evapotranspiration, AE_T equals PE_T. Under these conditions, there is sufficient moisture so that evaporation can proceed unhindered. When precipitation is less than potential evapotranspiration, AE_T equals precipitation plus any soil moisture which evaporates and transpires.

3.0 PROCESS OF EVAPOTRANSPIRATION

Evapotranspiration is a complex process involving many climatic, soil and plant variables. It constitutes two processes namely evaporation and transpiration. The upper zone of the soil (i.e., the unsaturated zone) constitutes the medium between the atmosphere and the saturated ground water system.

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, plant and soil. A schematic representation of the soil-plant-atmosphere system is shown in Figure 1.

Evapotranspiration varies as a consequence of variations in climate, crops and soils. Unlike precipitation meteorological parameters influencing evapotranspiration do not vary widely from place to place. However, the variation in crops and soils does influence to a large extent, the variation in ET.

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.

Plant controls a large number of the processes that determine ET rates, either by their use of radiant 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).

3.1 Evaporation

The process of evaporation of water in nature is one of the basic components of the hydrologic cycle by which water changes to vapour through the absorption of heat energy. The essential requirements for evaporation process are the source of heat to vaporize the liquid water and the presence of a concentration gradient of water vapour between the evaporating surface and the surrounding air . The source of energy for evaporation may be in solar energy, in the air blowing over the surface or in the underlying surface itself.

Evaporation can occur only when the vapour concentration at the vaporating surface exceeds that in the overlying air. Dalton (1982) stated that evaporation is a function of the difference in the vapour pressure of the water and the vapour pressure of the air which may be written as follows :

$$E = (e_s - e_d) . F(u) \qquad .... (1)$$

where

E = evaporation

e_s = saturation vapor pressure at the temp of evaporating surface, mm Hg

e_d = saturation vapour pressure at the dew point temperature, mm Hg, and

$F(u)$ = a function of the horizontal wind velocity.

The total heat content is the sum of the sensible heat, depending on temperature and latent heat depending on vapour pressure. The rate of evaporation is thus influenced by solar radiation, air temperature, vapourpressure, wind and to certain extent atmospheric pressure. Since solar radiation controls the net energy available for evaporation, evaporation varies with latitude, season, time of day and sky condition.

The rate of evaporation from a saturated soil surface is approximately the same as that from an adjacent water surface of the same temperature. As the soil begins to dry, evaporation decreases and its temperature rises to maintain the energy balance. Eventually evaporation ceases since there is no effective mechanism for transporting water from appreciable depths. Thus, the rate of evaporation from soil surfaces is limited by the availability of water.

3.2 Transpiration

Only a small portion of the water absorbed by the root system of a plant remains in the plant tissues. Almost all the water is lost to the atmosphere through transpiration. Transpiration is basically an evaporation process. However, transpiration by plants and trees is modified by the plant structure, stomatal behaviour coupled with the physical processes controlling evaporation.

The factors affecting transpiration may be physiological or environmental. Important physiological factors are density and behaviour ofsstomats, extent and character of protective covering, leaf structure, and plant diseases. The essential environmental factors include temperature, solar radiation, wind and soil moisture when the permanent wilting percentage is reached. Rainfall intercepted by vegetation is evaporated, thereby consuming some energy which otherwise would be available for transpiration. Plant type becomes an important factor in controlling transpiration when available soil moisture is limited. As the upper layers of the soil dry out, shallow rooted species can no longer obtain water and wilt, while deep rooted species continue to transpire until the soil moisture at greater depths is reduced to the wilting point.

4.0 MEASUREMENT AND ESTIMATION OF EVAPOTRANSPIRATION

The actual transfer of water from the ground to the atmosphere by evaporation and transpiration from various crops under different atmospheric conditions may be calculated from measurements of such quantities as the vertical gradient of water vapour and the transfer properties of atmosphere near the ground. But satisfactory instrumentation is elaborate and costly. Simple measurements of the loss of water from small tanks containing either soil and vegetation or water may be made under various conditions, but the correct interpretation of the results is doubtful unless the surface of the sample is homogeneous with the surroundings.

By the use of moisture controlled soil tanks it has been found that empirical relationships can be established between the evaporation from adequately watered land and the meteorological conditions. These relationships vary slightly from one crop to another and apply only in the special case where the potential transpiration of the crop is fully realised.

Two forms of empirical relationship have been used, one established directly between the consumptive use of water by irrigated crops and the meteorological conditions and the other established between the consumptive use and the evaporation from a hypothetical standard surface, calculated from consideration of the energy available for the evaporation of water.

4.1 Thornthwaite's Formula

The best known of the direct relationships is that of Thornthwaite (1948), who concluded that the waterneed of a particular crop was directly dependent on the air temperature and the length of day light. Thornthwaite gave the following relationship to estimate potential ET based on mean monthly temperature

$$PE_T = Ct^a$$

where

PE_T = potential evapotranspiration (cm/month)

t = mean monthly temperature (°C)

C and a are coefficients varying with temperature

$$a = 0.000\,000\,675\, I^3 - 0.0000771\, I^2 + 0.01792\, I + 0.49239 \quad \ldots (3)$$

I = annual heat index = i = monthly heat index ... (4)

$$i = (t/5)^{1.514}$$

The relationships he established between these quantities for a standard vegetative cover of turf varied slightly from place to place but could be fitted approximately by a family of curves distinguished by an index I which varied with the form of the normal annual cycle of temperature. Curves relating the water need of pasture in a standard month of 30 days of 12 hours to the mean air temperature are shown in Figure 2 for selected values of Thornthwaites index I.

The Thornthwaite's method has been widely used for estimates of monthly PE_T the world over. In India Prof. V P Subrahmanyam and his associates at the Andhra University used Thornthwaite's PE for computation of water balance by Thornthwaite's method.

The method being entirely dependent on temperature was found to overestimate the PET in summer months and underestimates PET in winter months.

4.2 Penman's Formula for Potential Evapotranspiration (PE)

According to Penman, potential evapotranspiration is the amount of water transpired in unit time by a short te green crop, completely shading the ground, of uniform height and never short of water'. The formula developed by Penman for estimating PE over vegetative cover is

$$PE = \frac{\left[R_A(1-r)\left(a+b\frac{n}{N}\right) - \quad (0.56-0.092\sqrt{\bar{e}_d})\left(0.10+0.90\frac{n}{N}\right)\right] + 0.35(e_a-e_d)\left(1+\frac{u}{100}\right)}{\overline{\quad} + 1} \qquad \ldots (5)$$

where

PE = potential evapotranspiration in mm per day

R_A = Incident radiation outside the atmosphere on a horizontal surface expressed in mm of evaporable water per day

r = reflection coefficient or albedo

$\frac{n}{N}$ = ratio of actual hours of sunshine to theoretical duration of sunshine

σ = Stefan - Boltzmen constant

a and b = constants

T = mean temperature in degrees absolute

σT^4 = black body radiation at mean temperature

e_a = saturation vapour pressure in mm of mercury

e_d = actual mean vapour pressure in mm of mercury

U = wind speed at 2 metres above ground in miles per day

= rate of change of saturation vapour pressure with temperature in mb. per degree Centigrade

= Psychrometric constant in mb. per degree Centigrade

$R_A(1-r)(a+b\frac{n}{N})$ represents the incoming shortwave radiation and $\sigma T^4(0.56-0.092\sqrt{e_d})(0.10+0.90\frac{n}{N})$ the outgoing longwave radiation. Reflection coefficient r (albedo) is taken as 0.25 for vegetation on the basis of Monteith's (1959) measurements. The aerodynamic term $0.35\,(1+\frac{U}{100})\,(e_a-e_d)$ is the drying power of the atmosphere based on saturation deficit, air movement and extra roughness of vegetation cover compared to water surface. The factor $\frac{\Delta}{r}$ makes allowance for the relative significance of net radiation and aerodynamic terms.

Three modifications have been introduced in the original equation by Rao et al (1971). These are

(i) Correction for elevation of the station :

In the original derivation of the equation, $\frac{\Delta}{\gamma}$ is taken to depend on mean air temperature only. Recent studies have shown that this factor depends on altitude. A correction is, therefore, applied by multiplying $\frac{\Delta}{\gamma}$ by $\frac{Po}{Ph}$ where Po is standar sea level pressure and Ph station level pressure.

(ii) Correction for latitude of the station :

In the term for incoming radiation, $(0.29 \cos \phi + 0.52\frac{n}{N})$ is used as suggested by Glover and McCulloch (1958) for $(a+b\frac{n}{N})$, ϕ being the latitude of the station. This term takes into account the obliqueness of sun-rays crossing the atmosphere and is, therefore, considered better for studying stations distributed over wide latitudinal belts. Radiation measurements made at stations in India gave values for 'a' ranging from .27 to .38 and for 'b' from .33 to .44. Values of 'a' obtained is thus slightly greater and 'b' slightly smaller than given by Glover and McCulloch.

(iii) Sunshine data :

As sunshine data normals are not available for all the stations, $\frac{n}{N}$ is taken uniformly for all stations as equal to (1-m), where m is the cloudiness expressed as a decimal fraction

The formula thus modified becomes

$$PE = \frac{\frac{(P_o)}{P_H} \_\_ \; R_A(1- \quad) \; 0.29\cos\phi + 0.52(1-m) - \sigma T^4(0.56-0.092\sqrt{e_d}) \; 0.10+0.9 \; (1-m) \; + 0.35(e_a-e_d)}{\frac{(P_o)}{P_H} \_\_ +1} \quad \ldots (6)$$

Normal monthly and annual PET of about 300 stations in and near India according to the modified formula were computed by Rao et al (1971). The isopleths of annual PET are shown in Figure 3.

4.3 Saxton Model

A comprehensive model to compute daily actual ET was developed and reported by Saxton et al (1974). This model lays emphasis on a graphical representation of various parameters. The amount of interception evaporation, soil evaporation and plant transpiration are combined to provide daily ET estimates. The model is shown in Figure 4.

The energy available is first utilised for evaporating the intercepted water. The remaining energy is used for soil evaporation and plant transpiration. The actual evaporation is limited by the available soil water content.

For dry soil with a plant canopy a part of the unused energy for soil evaporation is available to the plant transpiration to account for reradiated energy 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 the drawing of water from soil profile, water stress relationship expressed as a function of the available water capacity of the particular soil layer and the atmospheric demand.

The soil water is adjusted by abstracting the daily actual E_T from each rooting layer, adding daily infiltration computed from daily precipitation minus measured or estimated runoff and percolation.

4.4 Use of Lysimeters

Lysimeters are tanks filled with soil in which crops are grown under natural conditions to measure the amount of water lost by evapotranspiration. Soil conditions inside the lysimete are maintained to be very nearly the same as those in the surroundings. Lysimeters are grouped in three categories (i) non-weighing, constant water table type, which provides reliable data in areas where a high water table normally exists (ii) non weighing percolation type, in which changes in water stored in the soil are determined by sampling or neutron probe methods and the rainfall and percolation are measured, and (iii) weighing type, in which changes in soil water are determined either by weighing entire unit with a mechanical scale, or by electronic weighing balance.

While making measurements of ET from a particular crop species, it is essential to see that the root development should not be inhibited by the limited dimensions of the lysimeter.

There are at present over 40 lysimeters maintained by the Indian Meteorological Department, agricultural universities and those under ICAR. Their locations are shown in Figure 5.

5.0 ESTIMATING EVAPOTRANSPIRATION FOR CROPS

The crop evapotranspiration depends on several crops and environmental conditions such as climate, soil moisture, the type of crop, stage of growth and the extent to which plants cover the soil. Hence, in order to have accurate estimates of crop ET it is necessary to take into account these factors. If estimates of potential E_T for a reference crop are available, the estimation of E_T for specific crops can be made using

$$E_T = K_c E_t$$

where

K_c = crop coefficient determined experimentally

E_T = actual evapotranspiration

E_t = potential or reference evapotranspiration.

Experimentally derived crop coefficients, reflect the physiology of the crop, the degree of the crop cover, and the reference ET. Factors affecting the values of crop coefficient are mainly the crop characteristics, crop planting or showing data, rate of crop development, length of growing season and climatic conditions.

6.0 WATER TABLE INRELATION TO EVAPOTRANSPIRATION

Evaporation and transpiration from irrigated soil causes the water table to rise within a few feet of root zone of crops. The upward flow from water tables can have a substantial effect on the fluctuation of ground water storage. The rate of evaporation may be controlled by either the capacity of the atmospheric environment to evaporate water or the capacity of the soil to transmit water to the surface. The rise of water in the soil from a free water surface i.e. water table has been termed as capillary rise.

When a water table is present, soil water generally does not attain equilibrium over in the absence of vegetation, since the soil surface is subject to the evaporation due to atmospher conditions, If the soil and esternal conditions as constant, that is, if the soil is of stable structure, the water table is at a constant depth and over a given period of time the evaporative power of atmosphere remain constant, then in time, a steady state flow situation can develop from water to atmosphere in a soil.

A shallow ground water table may be present at a constan or variable depth. Where a ground water table occurs close to the surface, steady state flow may take place from the saturate zone beneath, through the unsaturated layer to the surface.

In the absence of a shallow ground water table, however, the loss of water at the surface and resulting upward flow of water would cause the soil to dry. Moore (1939) based on investigations of upward flow from water tables concluded that when the water table is at greater depth the finer soil supported high evaporation rate.

Where the water table is near the surface, the suction at the soil surface is low and the evaporation rate is determined by external conditions. However, as the water table becomes deeper and the suction at the soil surface increases, the evaporation rate approaches a limiting value irrespective of the evaporative power of the atmosphere.

7.0 Estimation of ET under variable Soil Moisture Situation

7.1 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.

7.2 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 :

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

$K \frac{(\theta(i,j) + \theta(i+1,j)}{2}$ 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 $\theta(i,j)$. 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

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

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

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

7.3 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 cvariation of hydraulic conductivity $K(\theta)$ with (θ) are shown in Figure 14 and 15 respectively for podure sand. The evaporation losses have beei 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 1. 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 beer estimated by using equation (13) and it was found t(be 0.414×10^{-2} cm. This is the evaporation loss during first time step.

4) During time step 2, the soil moisture of 2nd layer wi: decrease by an amount 0.41×10^{-2} cm. At this stage, the soil moisture flow will take place from 3rd laye: to 2nd layer and from 2nd layer to top layer respecti' The flow rate from 3rd layer to 2nd layer was 0.1824x cm and from 2nd layer to top layer was 0.321×10^{-2} c:

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 1.

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.

8.0 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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