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ESTIMATION OF EVAPOTRANSPIRATION

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1988-89

PREFACE

Evapotranspiration is one of the important elements of the hydrologic cycle. In its most simple form, evapotranspiration is the loss of moisture from a soil by evaporation and plant transpiration. Within the hydrologic cycle, evapotranspiration is next to precipitation in the quantity of water involved, and in fact they become nearly identical in semiarid and arid climates. Through its magnitude and close alliance with plant growth, evapotranspiration is a highly influential process to be considered and predicted for estimates related to hydrology.

The complexity of the evapotranspiration system which involves driving and controlling factors by the climate, soil, and vegetation, makes it one of the more difficult processes within the hydrologic cycle to measure, understand, and predict. The purpose of this technical note is to provide a concise summary of factors controlling evapotranspiration and to describe methods of measuring and estimating evapotranspiration. Various methods of estimating evapotranspiration are considered and conclusions are drawn concerning their usefulness and applicability.

This report entitled 'Estimation of Evapotranspiration' is a part of the research activities of 'Ground Water Assessment' division of the Institute. The study has been carried out by Shri Chandra Prakash Kumar, Scientist 'B' under the guidance of Dr.G.C.Mishra, Scientist 'F'. The manuscript has been typed by Shri A.K. Chatterjee, P.A. and Shri V.K. Srivastava, L.D.C. and the tracings have been prepared by Shri R.K. Garg, Tracer.

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Date: January 10, 1990

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ABSTRACT

Next to rainfall, evapotranspiration is the most important term in the water balance of catchment areas. The subject of evapotranspiration, which includes evaporation of water from land and water surfaces and transpiration by vegetation, is becoming increasingly more significant. Therefore, it is of interest to develop and test methods for estimating evapotranspiration.

A knowledge of evapotranspiration is necessary in planning and operating water resources development. Evapotranspiration data are essential for estimating water requirements for irrigation, and are useful for estimating municipal and industrial water needs, rainfall disposition, safe yields of ground water basins, water yields from mountain watersheds, and stream-flow depletions in river basins. Actual measurements of evapotranspiration under each of the various physical and climatic conditions of any large area are time-consuming and expensive. Thus, rapid and reliable methods are needed by hydrologists for estimating evapotranspiration.

To make a fair estimate of the evapotranspiration losses, one must conduct the extensive field surveys, and use one or more of the appropriate calculation methods. A common approach to predict evapotranspiration is to estimate the potential atmospheric energy available at the plant and soil surfaces, then determine the proportion of this energy utilized for conversion of liquid water to vapour depending on the water availability or rate of transmission from within the soil profile. Several methods to estimate the potential evapotranspiration have been developed and tested. They contain one or more atmospheric variables, or an indirect measurement of them, which are often combined with a representation

of the surface conditions and interactions. Some are based on the physics of combining the vertical radiation budget with turbulent boundary flow over the land surface. More empirical methods based on solar radiation, air temperature, air humidity, pan evaporation, or some combination of these have proven to be practical. Several models of the complete dynamic evapotranspiration process have been developed in recent years which vary considerably in complexity from single equations with empirical coefficients to very detailed physical representations. The utility of each method depends upon its requirements for input data, location calibration, and expected accuracy.

The present report deals with the major system processes which determine evapotranspiration. The report provides a summary of : theory of evaporation and evapotranspiration; measuring techniques for evapotranspiration; estimating evapotranspiration from meteorological data; and the recent numerical model studies.

1.0 INTRODUCTION

1.1 General

Evapotranspiration is the loss of water into the atmosphere through evaporation from all surfaces containing water, including evaporation from free-water surfaces, soil, and man-made surfaces and transpiration from plants. Although evapotranspiration is the second largest component in the hydrologic cycle (precipitation is the largest) and has undergone considerable investigation, it is still the most obscure part of the hydrologic cycle. There are two basic reasons for this obscurity; first, no instrumentation exists that can truly measure evaporation from a natural surface and second, evaporation and transpiration are highly sensitive to microclimatic influences and show considerable spatial variability. Therefore, one should estimate evaporation from a controlled surface(e.g., from evaporation pans and lysimeters) directly or by estimating indirectly as a residual amount in an equation involving many other natural processes (e.g., the water budget and energy budget). The magnitude of evapotranspiration from one plant-soil community to another often differs greatly, even under similar meteorological conditions. This is one reason why many evapotranspiration estimation techniques are not particularly accurate. To be understood as a complex, site-specific process, the estimation of evapotranspiration must involve a great amount of detailed measurement of the evaporating surface and the surrounding meteorological conditions. However, the complexity and expense of such investigations prevent them from being useful for many applications.

Transpiration is a natural plant physiological process whereby water is taken from the soil moisture storage by roots and passes through the plant structure and is evaporated from cells in the leaf called stomata.

Part of the water is utilized by photosynthesis in the manufacture of plant tissue. The quantity bound chemically is minor compared to that evaporated from the surface. It is logical that there should be a good correlation between transpiration and pan evaporation.

The watershed on which the plants grow also looses water by evaporation from the soil as well as from any streams or other water surfaces. Evapotranspiration is defined as the water vapour produced from the watershed as a result of the growth of the plants of the watershed and evaporation from other surfaces. The actual evapotranspiration can be determined by analysis of concurrent rainfall and runoff records from a watershed.

There is an important difference between evapotranspiration and free surface evaporation. Transpiration is associated with plant growth; therefore, evapotranspiration only occurs when the plant is growing. This results in diurnal and seasonal variations which are superimposed on the normal annual free water surface evaporation. Most plants transpire vigorously during the daylight hours (pineapple and cactus are exceptions). Plants transpire vigorously during their major growth season. The intensity of transpiration for wheat for example is greatest at the time when the grain is being formed. When the grain matures, the plant turns yellow and transpiration ceases. Transpiration by a deciduous tree occurs during the growing season. During the dormant seasons, the tree looses its leaves and transpiration ceases. Some perennial plants transpire during the entire non-freezing season. Blue grass lawns will begin to transpire when the mean daily temperature begins to exceed 35°F. Thus any correlation between lake evaporation and transpiration must also be conditioned to the normal growth season of the plants.

One of the practical applications of estimates of evapotranspiration is in the design of irrigation systems for artificially supplying water during seasons of deficient water for continuing plant growth. The terms Potential Evapotranspiration and Consumptive Use have evolved for this purpose. Potential evapotranspiration, PET, is defined as the evapotranspiration which would occur if there was always an adequate water supply available to a fully vegetated surface. This term implies an ideal water supply to the plants. If the water supply available to the plants is less than PET, the deficit will be drawn from soil moisture storage until about 50 percent of the available supply is utilized. With increasing moisture stress, the actual evapotranspiration, ET, will decrease below the PET until the wilting point is reached when evapotranspiration ceases.

1.2 An Overview of Evapotranspiration Principles

Evapotranspiration requires energy inputs, water availability, and transport processes from the surface into the atmosphere. The flux of water vapour is largely limited by one or more of these requirements. Several researchers have provided good descriptions of these primary variables which determine ET rates. Eagleson (1978) provided an extensive mathematical review which integrated the principles of ET into hydrologic predictions.

Because ET is a phase change of water, large energy inputs are required. At a nominal value of 580 cal/g, a daily ET of 5mm will require the equivalent of 4480 kg/ha coal. Solar radiation usually supplies 80 to 100 percent of this energy and is often the factor limiting ET. For nonirrigated agriculture, water availability to the evaporating plant and soil surfaces also often limits ET. Thus, the rate of ET is limited to the conduction rate of soil water to the soil surface and plant roots, and through the plant system. In these circumstances, absorbed radiant energy (incoming minus reflected) in excess of that required to transform the available water is dissipated primarily by an increase of sensible heat in the air, soil, and plant canopy.

Evapotranspiration varies spatially 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 or significantly over considerable distance. However, generalizations cannot be made because local elevations, aspects, orographic effects, and cropping patterns can cause large ET changes. The variation of crops and soils over a region in question will need to be treated either by separate considerations of major combinations or broad scale averages. Some spatial averaging is implicit in every ET estimate and the user must acknowledge and quantify the effects with respect to the application.

Of the water budget components, ET is usually the largest after precipitation. The interaction of ET with the 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, whether for hydrologic models or irrigation scheduling, follow a concept of a vertical water budget within a system. In general, the procedure is to first estimate or measure a potential for ET based on meteorological factors, then compute the amount of that potential that is utilized by the actual ET processes, given the current status of the plant and soil water related characteristics. To apply such a procedure requires variables that describe : (a) the potential ET; (b) plant water relationships ; and (c) soil water evaporation relationships.

Soil Water Balance

There are several situations where flow of water in soils is important to the study of average evapotranspiration rates. These are vertical unsaturated drainage from the root zone, upward flow from a shallow, saturated zone, and radial flow to plant roots.

The various sources of water and fluxes affecting the soil water balance are shown in figure 1 . The total quantities affecting the soil water over a given time period can be represented by the following major components.

$$\Delta w_s = w_i + R_e - w_{et} - w_d \qquad \ldots(1)$$

where Δw_s is the change in soil water content, w_i is the irrigation water, R_e is the effective rainfall (rainfall entering the soil or retained on the foliage or soil surface), w_{et} is the total quantity vaporized in evaporation from soil and plant surfaces including transpiration, and w_d is the total quantity drained from the effective root zone and not recovered by capillary rise from subsoils. In most areas, some deep percolation is essential since this is the only practical way of maintaining a favourable salt balance in the root zone.

Potential Evapotranspiration

Thornthwaite defined potential evapotranspiration as "the water loss which will occur if at no time there is a deficiency of water in the soil for the use of vegetation". However, most investigators have assumed that potential evapotranspiration is equal to lake evaporation as determined from National Weather Service Class A pan records. This is not theoretically correct because the albedo (amount of incoming radiation reflected back to the atmosphere) of vegetated areas and soils, ranges as high as 45%. As a result, potential evapotranspiration is somewhat less than

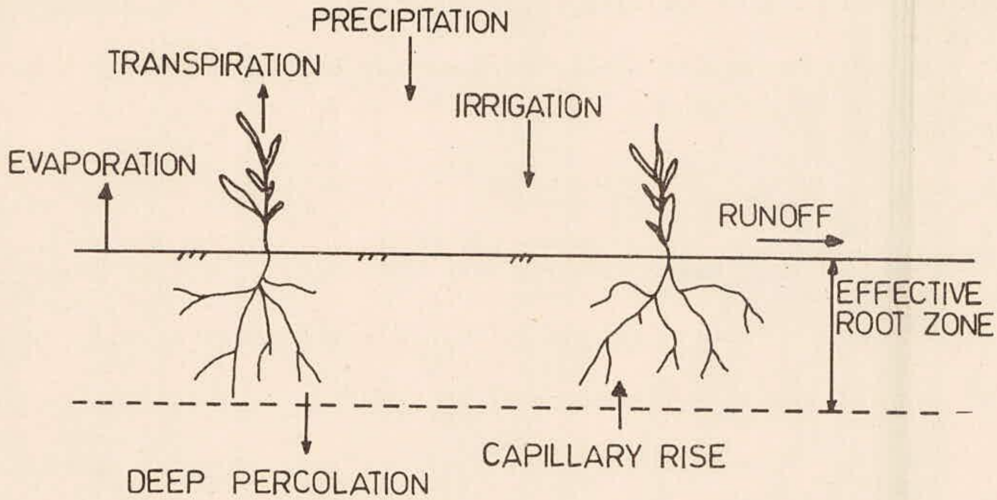


FIGURE 1 - COMPONENTS OF SOIL WATER BALANCE

free water surface evaporation.

Consumptive Use

Though both terms consumptive use and evapotranspiration are interchangeably used but in reality consumptive use includes water used by plant for its metabolic activities in addition to evaporation and transpiration losses. But as the water used in actual metabolic processes is insignificant (less than 1% of ET), the term consumptive use is taken equivalent to evapotranspiration.

Reference Evapotranspiration

In many types of evapotranspiration evaluations, it is common to first estimate a reference (or index) evapotranspiration and then adjust this amount empirically to estimate the actual evapotranspiration. A common adjustment factor is a crop coefficient, K_c, such that

$$ET = K_c E_r \qquad \ldots(2)$$

where E_r is the reference evapotranspiration rate having the same units as ET. The idea of a reference evapotranspiration was primarily introduced by Thornthwaite (1948) with his concept of 'potential evapotranspiration'. Potential evapotranspiration is generally viewed as the evapotranspiration that would occur from a surface with an unlimited moisture supply, depending only on climatic factors. Its definition is imprecise for use with transpiration, because the maximum expected transpiration will also be a function of the nature of the transpiring surface and will depend upon variations in the atmosphere-plant-soil interaction. For this reason, many investigators have adopted the terms unstressed evapotranspiration or reference evapotranspiration in place of potential evaporation. Thus, potential evapotranspiration is viewed as a type of reference evapotranspiration.

Virtually all reference values are estimates either of the potential evaporation or of a maximum expected evapotranspiration rate. Most of the methods used for estimating actual evapotranspiration can also apply to the reference evapotranspiration, provided that the area under observation has sufficient water at all times. Therefore, the lysimetry, energy budget, and combination methods are all applicable.

To model ET for numeric predictions, it is necessary to examine the entire ET system and to define the principle cause-and-effect relationships. The required accuracy for the objectives at hand and data availability must be decided. The temporal and spatial delineations must be considered. While ET proceeds continuously in strong diurnal fashion, integrated values of daily or weekly values are often most useful and expedient. And similarly, regardless that each small unit of the surface is unique and provides an ET rate different from others, there is much commonality across fields and watersheds in both conservative atmospheric inputs and land surface response. Therefore, single ET estimates are frequently applied to fairly large areas with useful results. Thus, although the ET process is complex, it can be defined, described, and predicted with an accuracy which will provide useful estimates and guidance for research.

1.3 Purpose and Scope

The purpose of this report on "Estimation of Evapotranspiration" is to provide a concise summary of factors controlling evapotranspiration and to describe methods of measuring and estimating evapotranspiration. The report provides the practicing engineer with a summary of: theory of evaporation and evapotranspiration; sources of evapotranspiration data; and its utilization.

2.0 EVAPORATION AND TRANSPIRATION

2.1 General

Evaporation is important in all water resource studies. Water will evaporate from land, either from bare soil or soil covered with vegetation, and also from trees, impervious surfaces like roofs and roads, open water and flowing streams. The rate of evaporation varies with the colour and reflective properties of the surface (the albedo) and is different for surfaces directly exposed to, or shaded from,solar radiation. Some of the important meteorological factors affecting evaporation are discussed below.

2.2 Meteorological Factors Affecting Evaporation

2.2.1 Solar radiation

Evaporation is the conversion of water into water vapour. It is a process that is taking place almost without interruption during the hours of daylight and often during the night also. Since the change of state of the molecules of water from liquid to gas requires an energy input (known as the latent heat of vaporisation), the process is most active under the direct radiation of the sun. It follows that clouds, which prevent the full spectrum of the sun's radiation reaching the earth's surface, will reduce the energy input and, therefore will reduce the process of evaporation.

2.2.2 Wind

As the water vaporises into the atmosphere, the boundary layer between earth and air becomes saturated and this layer must be removed and continually replaced by drier air if evaporation is to proceed. The movement of the air in the boundary layer depends on wind and so the wind speed is important.

2.2.3 Relative humidity

The third factor affecting evaporation is the relative humidity of the air. As the air's humidity rises, its ability to absorb more water vapour decreses and the rate of evaporation reduces. Replacement of the boundary layer of saturated air by air of equally high humidity will not maintain the evaporation rate;evaporation will continue only if the incoming air is drier than the air that is displaced.

2.2.4 Temperature

An energy input is necessary for evaporation to occur. It follows that if the ambient temperatures of the air and ground are high, evaporation will take place more rapidly than if the surrounding temperature is low. Since the capacity of air to absorb water vapour increases as its temperature rises, so air temperature has a double effect on how much evaporation takes place, while ground and water temperatures have single direct effects.

2.2.5 The nature of evaporating surface

Evaporation from a particular surface is directly related to the opportunity for evaporation (availability of water) provided by that surface. For open bodies of water, evaporation opportunity is 100%, while for soils it varies from a high of 100% when the soil is highly saturated-for example, during storm periods - to essentially zero at stages of very low moisture content. Other types of surfaces provide diverse degrees of evaporation opportunity and, except in rare cases, these will almost always vary widely with time.

2.3 Soil Water Evaporation

The evaporation of water from the soil surface involves the same basic physics as any other evaporation, i.e. a liquid-to-vapour phase change limited by the availability of either water or energy and upward vapour transport. For bare soil or fields with incomplete canopies, this component of the evapotranspiration process is highly important and can involve significant quantities of water. Given that bare soil is readily exposed to the atmosphere due to lack of vegetative cover, water availability frequently limits the soil water evaporation rate. Water can only be supplied by recent rainfall or by upward movement through capillary action.

Soil water evaporation is often described as occurring at three separate stages beginning with wet soil. In the first stage, the drying rate is limited by and equals the evaporative demand (available energy). During the second stage, water availability progressively becomes more limiting. The third stage is described as an extension of the second stage but is limited to a more constant rate.

The approach by van Bavel and Hillel (1976) of simultaneous heat and water flux at the soil surface and within the soil profile provides a more detailed and accurate prediction of soil evaporation but requires significantly more data input and computational time. An intermediate, and perhaps more feasible approach, may be to consider intercepted and ponded water at or near the soil surface for stage one drying and upward unsaturated flow for stage two.

Considerable effort has been made to explain many influences on soil evaporation, like mulches, residues, wetting methods, crust formation and tillage. Each of these effects can be significant, but the cause can usually be attributed to one of the physical limitations.

Separating the atmospheric potential reaching the soil surface from that going to the vegetation, and accounting for the moisture availability limitations, hydrologic predictions can be made for first estimates of soil evaporation which would require some calibration.

2.4 Methods of Estimating Evaporation

2.4.1 Water budget or storage equation approach for estimating evaporation from a basin

This method consists of drawing a balance sheet of all the water entering and leaving a particular catchment or drainage basin. If rainfall is measured over the whole area on a regular and systematic basis then a close approximation to the amount of water arriving from the atmosphere can be made. Regular stream gauging of the streams draining the area, and accurately prepared flow-rating curves, will indicate the water leaving the area by surface routes. The difference between these two can be accounted for in only three ways:

(i) by a change in the storage within the catchment, either in surface lakes and depressions or in underground aquifers;

(ii) by a difference in the underground flow into and out of the catchment;

(iii) by evaporation and transpiration.

The storage equation can be written generally as

$$E = P + I - U - O - \Delta S \qquad \ldots(3)$$

where, E = evapotranspiration,
P = total precipitation,
I = surface inflow(if any),
U = underground outflow,
O = surface outflow,
ΔS = change in storage(positive for increase).

If the observations are made over a sufficiently long time, the significance of ΔS, which is not cumulative, will decrease and can be ignored if the starting and finishing points of the study are chosen to coincide as nearly as possible with the same seasonal conditions. The significance of U can not be generalised but in many cases can be assigned second-order importance because of known geological conditions that preclude large underground flows. In such cases a good estimation of evapotranspiration becomes possible and the method provides a means of arriving at first approximations.

2.4.2 Energy budget method

The energy budget method illustrates an application of the continuity equation written in terms of energy. It has been employed to compute the evaporation from oceans and lakes. The equation accounts for incoming and outgoing energy balanced by the amount of energy stored in the system. The accuracy of estimates of evaporation using the energy budget is highly dependent upon the reliability and preciseness of measurement data. Under good conditions, average errors of perhaps 10% for summer periods and 20% for winter months can be expected.

The energy budget equation for a lake may be written as,

$$Q_o=Q_s-Q_r+Q_a-Q_{ar}+Q_v-Q_{bs}-Q_e-Q_h-Q_w \qquad \ldots(4)$$

where,

Q_o = increase in stored energy by the water,

Q_s = solar radiation incident at the water surface,

Q_r = reflected solar radiation,

Q_a = incoming long-wave radiation from the atmosphere,

Q_v = net energy advected (net energy content of incoming and outgoing water) into the water body,

Q_{ar} = reflected long-wave radiation,

Q_{bs} = long-wave radiation emitted by the water,

Q_e = energy used in evaporation,

Q_h = energy conducted from water mass as sensible heat, and

Q_w = energy advected by evaporated water.

All the terms are in calories per square centimeter per day. Heating brought about by chemical changes and biological processes is neglected. The transformation of kinetic energy into thermal energy is also excluded. These factors are usually very small, in a quantitative sense, when compared with other terms in the budget if large reservoirs are considered. As a result, their omission has little effect on the reliability of results.

During winter months when ice cover is partial or complete, the energy budget only occasionally yields adequate results because it is difficult to measure reflected solar radiation, ice-surface temperature, and the areal extent of the ice cover. Daily evaporation estimates based on the energy budget are not feasible in most cases because reliable determination of changes in stored energy for such short periods is impractical. Periods of a week or longer are more likely to provide satisfactory measurements.

In using the energy budget approach, the required accuracy of measurement is not the same for all variables. For example, errors in measurement of incoming long-wave radiation as small as 2% can introduce errors of 3 to 15% in estimates of monthly evaporation, while errors of the order of 10% in measurements of reflected solar energy may cause errors of only 1 to 5% in calculated monthly evaporation.

To permit the determination of evaporation by equation (4), it is common to use the following relationships:

$$B = \frac{Q_h}{Q_e} \qquad \ldots(5)$$

where, B is known as Bowen's ratio, and

$$Q_w = \frac{C_p Q_e (T_e - T_b)}{L} \qquad \ldots(6)$$

where,

C_p = the specific heat of water (cal/g-$^{\circ}$C),

T_e = the temperature of evaporated water ($^{\circ}$C),

T_b = the temperature of an arbitrary datum usually taken as 0°C,

and L = the latent heat of vaporization (cal/g).

Introducing these expressions in equation (4) and solving for Q_e, we obtain

$$Q_e = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o + Q_v}{1 + B + C_p (T_e - T_b)/L} \qquad \ldots(7)$$

To determine the depth of water evaporated per unit time, the expression

$$E = \frac{Q_e}{\rho L} \qquad \ldots(8)$$

may be used, where

E = evaporation (cm^3/cm^2-day), and

ρ = the mass density of evaporated water (g/cm^3).

The energy budget equation thus becomes,

$$E = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o + Q_v}{\rho [L(1+B) + C_p (T_e - T_b)]} \qquad \ldots(9)$$

The Bowen ratio can be computed using

$$B = 0.61 \frac{p}{1000} \frac{(T_o - T_a)}{(e_o - e_a)} \qquad \ldots(10)$$

where,

p = the atmospheric pressure (mb),

T_o = the water surface temperature (^oC),

T_a = the air temperature (^oC),

e_o = the saturation vapour pressure at the water surface temperature (mb), and

e_a = the vapour pressure of the air (mb).

This expression circumvents the problem of evaluating the sensible heat term, which does not lend itself to direct measurement.

2.4.3 Empirical formulae

Many attempts have been made to produce satisfactory formulae for the estimation of evaporation. These are usually for evaporation from an open water surface. Evaporation, if it is to take place, presupposes a supply of water. Whatever the meteorological conditions may be, if there is no water present then there can be no evaporation. Accordingly, estimating methods using meteorological data work on the assumption that abundant water is available; that is, a free water surface exists. The results obtained therefore are not necessarily a measure of actual but of potential evaporation. Often these two are the same, as for example, in reservoirs where a free water surface exists. When evaporation from land surfaces is concerned, the loss of water in this way clearly depends on availability: rainfall, water-table level, crop or vegetation, and soil type all have an influence, which can be expressed by applying an empirical factor, usually less than unity, to the free water surface evaporation. There are two cases that should be considered:

(i) when the temperature of the water surface is the same as the air temperature;

(ii) when the air and water surface temperatures are different.

Case (i) rarely occurs and is empirically treated by an equation of the form

$$E_a = C(e_s - e)f(u) \qquad ...(11)$$

where,

E_a = open water evaporation per unit time (for air and water temperature the same t^oC) in mm/day,

C = an empirical constant,

e_s = saturation vapour pressure of the air at t^oC(mm mercury),

e = actual vapour pressure in the air above (mm mercury), and

u = wind speed at some standard height.

An empirical equation of the above type which has been used, is

$$E_a = 0.35(e_s - e)(0.5 + 0.54\ u_2) \qquad ...(12)$$

u_2 denotes wind speed in m/s at a height of 2m, E_a is in mm/day. Case (ii) is the one that normally occurs and the following formula has been used:

$$E_o = C(e'_s - e)f(u) \qquad ...(13)$$

in which e'_s is the saturation vapour pressure of the boundary layer of air between air and water, whose temperature t'_s is not the same as either air or water and is virtually impossible to measure. Accordingly empirical formulae have been developed in the form of equation (11), which work fairly well for specific locations where the constants have been derived.

Such a formula, derived for the Ijsselmeer in The Netherlands, and only applicable to it under similar conditions, is

$$E_o = 0.345(e_w - e)(1 + 0.25\ u_6) \qquad \ldots(14)$$

where,

E_o = evaporation of the lake in mm per day,

e_w = saturation vapour pressure at temperature t_w of the surface water of the lake in mm mercury,

e = actual vapour pressure in mm mercury, and

u_6 = wind velocity in m/s at a height of 6 m above the surface.

2.4.4 Penman's theory

The following nomenclature is used :

E_o = evaporation from open water (or its equivalent in heat energy),

e_w = saturation vapour pressure of air at water surface temperature t_w,

e = actual vapour pressure of air at temperature t

= saturation vapour pressure at dew-point t_d,

e_s = saturation vapour pressure of air at temperature t,

e'_s = saturation vapour pressure of air at boundary layer temperature t'_s,

n/D = cloudiness ratio = actual/possible hours of sunshine,

R_A = Angot's value of solar radiation arriving at the atmosphere,

R_c = sun and sky radiation actually received at the earth's surface on a clear day,

R_I = net amount of radiation absorbed at surface after reflection, and

R_B = radiation from the earth's surface.

In 1948 Penman presented a theory and formula for the estimation of evaporation from weather data. The theory is based on two requirements, which must be met if continuous evaporation is to occur. These are:

(i) there must be a supply of energy to provide latent heat of vaporisation;

(ii) there must be some mechanism for removing the vapour, once produced.

<u>The energy supply</u>

During the hours of daylight there is a certain measurable amount of short-wave radiation arriving at the earth's surface. The amount depends on latitude, season of the year, time of day and degree of cloudiness. Assuming there were no clouds and a perfectly transparent atmosphere, the total radiation to be expected at a point, has been given in tabular form by Angot, and is reproduced in table 1 as values of R_A.

If R_c=short-wave radiation actually received at the earth from sun and sky and n/D = ratio of actual/possible hours of sunshine, then Penman gives (for Southern England)

$$R_c = R_A(0.18+0.55\ n/D) \qquad \ldots(15)$$

Kimball has given the following formula for Virginia, USA

$$R_c = R_A(0.22+0.54\ n/D) \qquad \ldots(16)$$

and Prescott has given the following relation for Canberra, Australia

$$R_c = R_A(0.25+0.54\ n/D) \qquad \ldots(17)$$

Thus even on days of complete cloud cover (n/D=0), about 20 percent of solar radiation reaches the earth's surface, while on cloudless days about 75 percent of radiation gets through.

Part of R_c is reflected as short-wave radiation; the exact amount depends on the reflectivity of the surface, or, the reflection coefficient r.

Table 1 - Angot's Values of Short-Wave Radiation Flux R_A at the outer Limit of the Atmosphere (in g cal/cm^2/day) as a Function of the Month of the Year and Latitude.

Latitude (degrees)	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
N 90	0	0	55	518	903	1077	944	605	136	0	0	0	3540
80	0	3	143	518	875	1060	930	600	219	17	0	0	3660
60	86	234	424	687	866	983	892	714	494	258	113	55	4850
40	358	538	663	847	930	1001	941	843	719	528	397	318	6750
20	631	795	821	914	912	947	912	887	856	740	666	599	8070
Equator	844	963	878	876	803	803	792	820	891	866	873	829	8540
20	970	1020	832	737	608	580	588	680	820	892	986	978	8070
40	998	963	686	515	358	308	333	453	648	817	994	1033	6750
60	947	802	459	240	95	50	77	187	403	648	920	1013	4850
80	981	649	181	9	0	0	0	0	113	459	917	1094	3660
S 90	995	656	92	0	0	0	0	0	30	447	932	1110	3540

[Source: David Brunt (1944)]

If R_I = the net amount of radiation absorbed, then (for Southern England)

$$R_I = R_c(1-r) = R_A(1-r)(0.18+0.55\ n/D) \qquad \ldots(18)$$

In turn, some of R_I is re-radiated by the earth as long-wave radiation, particularly, at night when the air is dry and the sky clear. The net outward flow R_B may be expressed empirically as

$$R_B = \sigma T_a^4(0.47-0.077\sqrt{e})\ (0.20+0.80\ n/D) \qquad \ldots(19)$$

where,

σ = Lummer and Pringsheim constant

= 117.4×10^{-9} g cal/cm^2/day,

T_a = absolute earth temperature = t^oC+273, and

e = actual vapour pressure of air in mm mercury.

Hence the net amount of energy finally remaining at a free water surface ($r=0.06$) is given by H, where

$$\begin{aligned} H &= R_I - R_B \\ &= R_c - rR_c - R_B \\ &= R_c(1-r) - R_B \\ &= R_A(0.18+0.55\ n/D)\ (1-0.06) - R_B \end{aligned}$$

Therefore,

$$H = R_A(0.18+0.55\ n/D)\ (1-0.06)-(117.4 \times 10^{-9})$$
$$T_a^4(0.47-0.077\sqrt{e})\ (0.20+0.80\ n/d) \qquad \ldots(20)$$

This heat is used up in four ways, that is

$$H = E_o+K+S+C \qquad \ldots(21)$$

where,

E_o = heat available for evaporation from open water;

K = convective heat transfer from the surface;

S = increase in heat of the water mass(that is, storage);

and C = increase in heat of the environment(negative advected heat).

Over a period of days and frequently over a single day, the storage of heat is small compared with the other changes, and the same is true of environmental storage, so that with a small degree of error,

$$H = E_o + K$$

Vapour removal

It has been shown that evaporation may be represented by

$$E_o = C(e'_s - e)f(u) \qquad \ldots(22)$$

but that e'_s can not be evaluated if air and water are at different temperatures. Penman now made the assumption that the transport of vapour and the transport of heat by eddy diffusion are essentially controlled by the same mechanism (that is, atmospheric turbulence), the one being governed by (e'_s-e), the other by (t'_s-t). To a close approximation therefore

$$\frac{K}{E_o} = \beta = \frac{\gamma(t'_s-t)}{e'_s-e} \qquad \ldots(23)$$

where,

γ = psychrometer constant

= 0.66 if t is in ^oC and e in mbar.

Now since

$$H = E_o + K = E_o(1+\beta)$$

then

$$E_o = \frac{H}{1+\beta} = \frac{H}{1+\gamma\dfrac{(t'_s-t)}{(e'_s-e)}} \qquad \ldots(24)$$

Now eliminate (t'_s-t) by substitution, since

$$(t'_s-t) = \frac{(e'_s-e_s)}{\Delta} \qquad \ldots(25)$$

where,

e_s = saturation vapour pressure at temperature t, and

Δ = slope of vapour pressure curve at t,
= tan α (see figure 2)

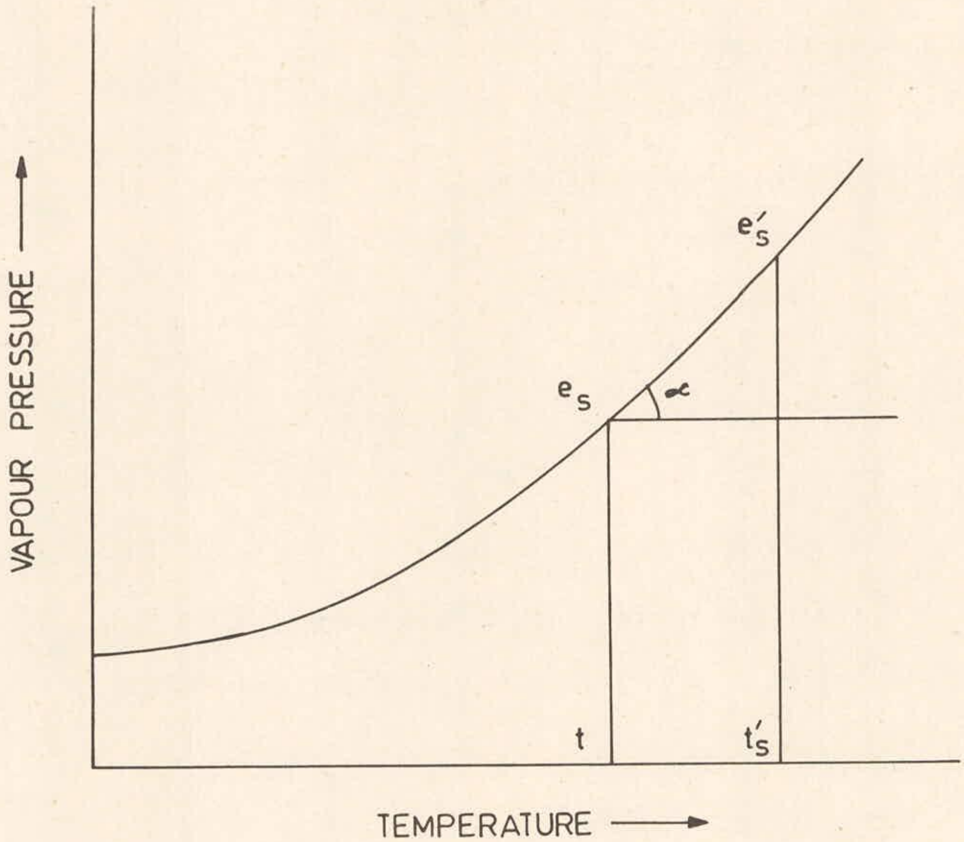


FIGURE 2 - SATURATION VAPOUR PRESSURE CURVE

This is reasonable since t_s' is never very far from t.

Hence,

$$E_o = \frac{H}{1 + \frac{\gamma}{\Delta} \cdot \frac{e_s' - e_s}{e_s' - e}} \qquad \ldots(26)$$

Now e_s must be eliminated.

Since, $\quad e_s' - e_s = (e_s' - e) - (e_s - e) \qquad \ldots(27)$

and from equation (11),

$$E_a = C(e_s - e)f(u)$$

and from equation (13),

$$E_o = C(e_s' - e)f(u)$$

then

$$\frac{E_a}{E_o} = \frac{e_s - e}{e_s' - e} \qquad \ldots(28)$$

where, E_a = evaporation (in energy terms) for the hypothetical case of equal temperatures of air and water.

Then by the values of equations (27) and (28) into equation (26),

$$E_o = \frac{H}{1 + \frac{\gamma}{\Delta} \left[\frac{(e_s' - e) - (e_s - e)}{e_s' - e}\right]}$$

and

$$E_o = \frac{H}{1 + \frac{\gamma}{\Delta}\left(1 - \frac{E_a}{E_o}\right)}$$

from which

$$E_o = \frac{\Delta H + \gamma E_a}{\Delta + \gamma} \qquad \ldots(29)$$

Δ has values obtained from the saturation vapour pressure curve, typically as shown

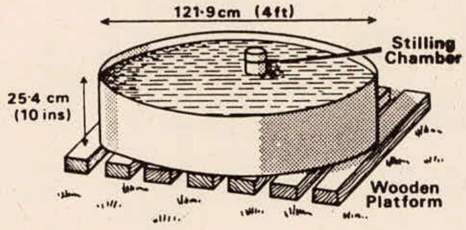
$t = 0^oC$	$\Delta = 0.36$
10	0.61
20	1.07
30	1.80

Referring to equations (12) and (20) for E_a and H respectively, it can be seen that E_o is now computed from standard meteorological observations of mean air temperature, relative humidity, wind velocity at a standard height and hours of sunshine. The formula has been checked in many parts of the world and gives very good results. Being based on physical principles it is of general application and gives values that should serve for most project studies until supplemented by actual evaporation measurements.

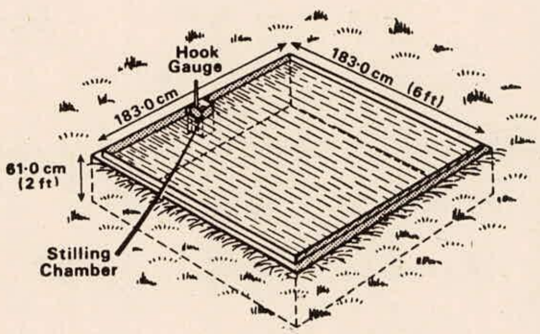
2.4.5 Direct measurement of evaporation by pans

Whenever possible, direct observations of evaporation should be made. The instrument used for this is the evaporation pan (figure 3). In Britain the standard pan is 1.83 m square and 610 mm deep filled to a depth of 550 mm and set in the ground so that the rim of the pan projects 76 mm above the surrounding ground. In the USA the standard or class A pan is circular 1.22 m in diameter and 254 mm deep, filled to a depth of 180 mm set on a timber grillage with the pan bottom 150 mm above ground level. A third type of pan is sometimes used in the United Kingdom, the Peirera pan, which is circular like the class A pan but deeper and sunk in the ground with a 3 in. air space surrounding it. The class A pan has a greater daily range of temperature than the square one, but is usually homogeneous whereas the water in the square pan may stratify. Doubling the wind run may increase evaporation by upto 20 percent.

The relatively small capacities and shallow depths of pans in comparison to lake and river volumes and their situation at or near the land surface allows proportionately greater amounts of advected heat from the atmosphere to be absorbed by the water in the pan through the sides and bottom, than by natural open water. Pan evaporation is therefore



U.S. Weather Bureau Class A Pan



British Meteorological Office Pan

FIGURE 3 - EVAPORATION PANS

usually too high and a pan coefficient has to be applied. These coefficients range from 0.65 to less than 1, depending on the dimensions and siting of pan. Generally the standard British pan has a coefficient about 0.92 and the U.S. Weather Bureau Class A pan about 0.75.

There are difficulties in using pans for the direct measurement of evaporation, because of the difference in the boundary conditions.

2.4.6 Atmometers

These are devices that can give direct measurement of evaporation. A water supply is connected to a porous surface and the amount of evaporation over a designated time period is given by a measure of the change in the volume of water. Thus $E_o = \Delta S$. It is essential to have a continuous instrument exposure to ensure consistent observations and it has been found satisfactory to have atmometers set in a well ventilated screen as is used for exposing thermometers to register air temperature. Atmometers are simple, inexpensive and easy to operate, but care must be taken to see that the porous surfaces from which the evaporation takes place are kept clean. Two types are described here:

The 'Piche' evaporimeter consists of a glass tube, 14 mm in diameter and 225 mm long with one end closed. A circular disc 32 mm diameter of absorbent blotting paper is held against the open end by a small circular metal disc with a spring collar. The evaporating surface area is 1300 mm^2 and this is fed constantly by the water in the tube hung up by its closed end. The tube is graduated to give a direct reading of evaporation (E_o) over a chosen time period, usually a day. The measurement in millimetres is related to the evaporating surface of both sides of the paper. The tube holds an equivalent of 20 mm of evaporation, the water is replenished when necessary. When the Piche evaporimeter is exposed in a standard temperature screen, the annual values have been found to be approximately equivalent to the open water evaporation from a US Class A pan. This type of instrument is used widely in the developing countries of Africa and the Near East.

In the 'Bellani' atmometer, the porous surface is provided by a thin ceramic disc, 85 mm in diameter. This is attached to a graduated burette holding the water supply. As with the Piche evaporimeter, the difference in burette reading over a specified time gives the measure of evaporation.

2.4.7 Evaporation from shallow and deep lakes

In the case of a deep lake it is necessary to consider the amount of heat energy that is stored and the net energy that is advected into the lake from inflow and outflow. Kohler et al. (1955) suggest that the following equation may be used to obtain evaporation from a deep lake using estimated amounts from shallow lakes:

$$E'_L = E_L + \alpha_L (R_v - R_\theta) \qquad \ldots(30)$$

where E'_L and E_L = evaporation from deep and shallow lakes, respectively,

α_L = proportion of advected energy into the lake used for evaporation,

R_v = net energy advected to the lake as measured from inflow and outflow, and

R_θ = change in stored energy in the lake (measured).

2.5 Selection of a Method for Determining Evaporation

Any of the methods described can be employed to determine evaporation. Usually, instrumentation for energy budget method is quite expensive and the cost to maintain observations is substantial. For these reasons, the water budget method and use of evaporation pans are more common. The pan method is the least expensive and will frequently provide good estimates of annual evaporation. Any approach selected is dependent, however, upon the degree of accuracy required. As the ability to evaluate the terms in the water budget and energy budget improves, so will the resulting estimates of evaporation.

2.6 Transpiration

Plants play a significant and dynamic role in the evapotranspiration process, particularly on areas largely used for agricultural crops, which are sown, harvested, and cultivated with considerable variation in both time and space. For well-vegetated surfaces, most of the radiant energy is absorbed and dissipated by the leaf surfaces. Thus, the plants have primary control of the amount utilized for latent heat through stomatal control, water availability, and root proliferation. For agricultural applications, it is quite important to represent the plant functions because vegetation in many climates seldom transpires at a potential rate. This may either be the result of plant control or water availability. Recently developed ET methods have combined these plant effects and descriptions through system models of ET from crops with incomplete canopies and limited water where transpiration and soil evaporation are calculated separately. Plant characteristics for ET methods can be divided into the main categories of (a) canopy cover, (b) phenology, (c) roots, and (d) water stress. While there are many interactions among these categories, they represent major considerations for computational purposes and provide a useful framework for model representations.

2.6.1 Canopy

The dynamic development, maturation, and decay of crop canopies significantly influence plant transpiration effects. For annual crops, like corn or cotton, the canopy very rapidly develops from nothing to nearly full soil cover and then matures and is harvested. The canopy of any particular day largely determines the amount of intercepted solar radiation or adsorbed advection utilized for plant transpiration. Hydrologic models must provide a representation of this dynamic plant behaviour.

A direct approach is to plot the canopy growth curve versus time to represent the percent of ground shading throughout the year. To define crop canopy curves, it is required to have a knowledge or observations of normal planting dates, emergence times, rate of development, tasseling or blooming dates, harvest dates, regrowth or residue conditions. To represent canopy as average daily soil shading primarily is to partition the radiant energy between plant and soil, thus modifications need to be considered if advection is expected to play a significant role. Although not highly accurate, an empirical canopy curve based on local knowledge of crop growth will often adequately represent crop canopy effects.

Recent research on crop effects has used the ratio of leaf area divided by soil surface area as a leaf area index (LAI) to relate measured evapotranspiration to effective canopy. This measure compares canopy effects of different crops, although it has not been entirely satisfactory among crops of widely differing canopy architecture. For almost all cases, the actual-to-potential evapotranspiration (ET/PET) ratio approaches 1.0 as the LAI approaches 3.0. Although LAI values relate closely to the ET/PET ratio and provide a direct measure of crop canopy, they are difficult to predict, and estimates of canopy cover as a percent of the soil shading may yet be the most practical.

2.6.2 Phenology

The phenological development of plants often modifies a plant's ability to transpire. As a crop matures, its need for water and ability to transpire are diminished. Because phenological changes may occur independent of the crop canopy present, this effect must be introduced as a modifier. As with canopy, a time distribution graph of the relative ability of a plant canopy to transpire will often be adequate for hydrology models. The effect being represented is the transpiration ability of the canopy existing at any time as compared with that of a fully transpiring, equal canopy. Crop maturation is the principal cause for loss of ability, but drying of leaves from stress of heat, moisture,or insects may also cause modifications.

Crop residues pose a special canopy situation since they intercept radiation but have no ability to utilize that energy for evaporation unless intercepted precipitation is present. This can be considered a special case of crop phenology where the crop has lost all ability to transpire, or the residues may be considered as part of the soil evaporation process. Nevertheless, the residues change characteristics over time through decay and destruction by tillage and must be dynamically represented.

The combined effects of plant canopy and phenology have often been represented by crop coefficient curves, particularly for irrigation evapotranspiration estimates. To use this approach, we must carefully note the derivation and intended application of these coefficient curves because they are empirically determined and include specific conditions and assumptions. Typical crop coefficient curves are reported by Jensen (1973). Doorenbos and Pruitt (1977) provided a method to develop curves for many crops under a wide range of growing season lengths and climates.

2.6.3 Roots

Crop roots are as important as canopy in the process of connecting soil water with atmospheric energy and the resulting transpiration. However, root distribution and their effectiveness are more difficult to study and quantify. Some information is available on most crops and basic relationships are presented in texts.

Many evapotranspiration models have simply considered depth of maximum rooting as a predetermined parameter or fitted coefficient. For more physically related modelling, the time and depth of rooting density is required, especially for annual crops which establish a complete new root system each year. Only in this way can soil water profiles and their interaction with the root profiles be modelled. This becomes very important when large differences in water contents exist with soil depth. Crop transpiration may be severely limited if the rooting patterns do not coincide with available moisture in the soil profile.

The water uptake by plants and the mathematical representation of this phenomenon have received considerable attention in recent years. In addition to representing time and depth of rooting densities, several approaches for calculating the resistance of water flow to the root, then through the root, stems, and leaves have been proposed. Root age and location seem to be quite important. It can be speculated that the deeper, less dense roots may be more effective for water uptake because they are younger and usually in wetter soil. Studies in this level of detail have not yet provided additional prediction capability; thus, including basic root-density dynamics for interaction with available soil water may be the most sophistication now warranted.

Most crops have a genetic rooting characteristic that will provide estimates of root density distributions with depth and time; and, in turn, estimates of water extractions. Older roots are less efficient in water extraction, thus root quantity distributions must be modified. Characteristic rooting patterns can be significantly modified by the soil root environment like dense layers, poor aeration, very dry or wet, and chemicals. Some deep-rooted plants, like alfalfa or trees, may extract water from deep, wet layers or shallow ground water.

2.6.4 Water stress

The lack of available plant water and/or high evaporative demands will cause most plants to biologically react by closing stoma, reducing transpiration, and reducing assimilation and metabolic reactions. Continued stress results in leaf drop and tissue death. While this process has long been observed, there is yet considerable controversy and lack of definitive predictive relationships. Many studies have been conducted and reported, but the results are quite variable, probably because of crop, soil, climatic or technique differences. It is generally agreed that both plant-available soil moisture (or soil water pressure) and the atmospheric demand determine that proportion of potential transpiration a plant will achieve, i.e., the actual/potential ratio. 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 moisture stress and a significant reduction of transpiration from the potential.

Water stress relationships will require calibration for each soil depth used to define available soil water, each crop, and perhaps for each soil. Soil water pressure (capillary suction) may be a better measure than plant available water. Additional effects that have been investigated, like soil conductivity near the roots and plant water pressure, may eventually reduce the need for calibration. 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 and boundary resistances, but this approach will require further development and testing before it can be readily applied.

3.0 UNSATURATED ZONE AND EVAPOTRANSPIRATION

3.1 General

Proper water management of any region with competing interest implies an evaluation of the various measures that are available. The hydrological part of this evaluation can be made with the aid of mathematical-hydrological models. These models concern the entire system of surface water-ground water-soil water-evapotranspiration.

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 physical processes occurring in the soil-plant system. Hydrological measures taken will affect the processes in the unsaturated zone. In the following the soil-physical aspects of the unsaturated zone will be treated with respect to ground water movement due to capillary action and evapotranspiration.

3.2 Unsaturated Zone

3.2.1 General theory

Water in soil moves from a point where it has a high energy status to points where it has a lower one. The energy status of water is the water potential which consists of several components. Potentials are defined relative to the reference status of water at atmospheric pressure and elevation datum zero. In hydrology, potential is usually expressed as energy per unit weight of soil water, with the dimension of length, i.e. cm and potential is then denoted as 'head'. When dealing only with the matric head, h_m, arising from local interacting forces between soil and water and gravitational head, Z, arising from the gravitational force, total (hydraulic) head, H, can be expressed as:

$$H = h_m + Z(\text{cm}) \qquad \ldots(31)$$

where the vertical coordinate Z is considered positive in upward direction. Under conditions of atmospheric pressure and non-swelling soils the matric head can be denoted as pressure head.

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

$$\theta = f(h) \qquad \ldots(32)$$

To describe the flow of water in soil systems, it is customary to use Darcy's law. For one dimensional vertical flow, the volumetric flux q ($cm^3.cm^{-2}.d^{-1}$) can be written as:

$$q = -K \frac{\delta H}{\delta Z}(\text{cm.d}^{-1}) \qquad \ldots(33)$$

where K is the hydraulic conductivity ($cm.d^{-1}$).

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

$$K = f(\theta) \text{ or } K = f(h) \qquad \ldots(34)$$

Substitution of equation (31) into equation (33) yields:

$$q = -K(h) \left(\frac{\delta h}{\delta Z} + 1\right) \qquad \ldots(35)$$

In order to get a complete mathematical description, the continuity principle (Law of Conservation of Matter) can be applied:

$$\frac{\delta \theta}{\delta t} = - \frac{\delta q}{\delta Z}(d^{-1}) \qquad \ldots(36)$$

Substitution of equation (35) in equation (36) yields:

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta Z}\left[K(h)\left(\frac{\delta h}{\delta Z} + 1\right)\right] \qquad \ldots(37)$$

To avoid the problem of the two dependent variables θ and h, the derivative of θ with respect to h can be introduced, which is known as the differential soil water capacity C.

$$C = \frac{d\theta}{dh}(cm^{-1}) \qquad \ldots(38)$$

Writing

$$\frac{\delta \theta}{\delta t} = \frac{d\theta}{dh} \cdot \frac{\delta h}{\delta t} \qquad \ldots(39)$$

and substitution of equation (38) into equation (37) yields the one-dimensional equation for water flow in heterogeneous soils.

$$\frac{\delta h}{\delta t} = \frac{1}{C(h)} \frac{\delta}{\delta Z}\left[K(h)\left(\frac{\delta h}{\delta Z} + 1\right)\right] \qquad \ldots(40)$$

3.2.2 Steady-state capillary rise

The water use of crops is affected by the upward flow from the relatively shallow ground water table. This upward flow is usually termed capillary rise.

Integration of equation (35) in steady-state conditions yields the relationship between flux, q, pressure head, h, and vertical coordinate, Z. For special types of $K(h)$-functions analytical solution of the integration is possible. Solution by numerical integration, however, is always possible, both for homogeneous and heterogeneous (layered) profiles. Therefore equation (35) can be written in a finite difference notation as:

$$\Delta Z = - \frac{1}{1 + q/K(h)} \Delta h \qquad \text{...(35a)}$$

and applied to each layer separately.

3.2.3 Non-stationary flow

In case of a non-stationary flow, $\delta h/\delta t \neq 0$ and equation(40) is valid. Solution of this equation for non-homogeneous soil profiles is only possible with numerical and analog models.

Writing equation (40) in a finite difference form yields:

$$\frac{\Delta h}{\Delta t} = \frac{1}{C(h)\,\Delta Z} \Delta[K(h)(\frac{\Delta h}{\Delta Z} + 1)] \qquad \text{...(41)}$$

The functional relationships $C(h)$ and $K(h)$ now must be 'averaged' over time and space. Because of the high non-linearity in the $C(h)$ and $K(h)$ functions, the maximum time step to be allowed during the computation is relatively small.

For a unique solution of h with respect to time and space, initial and boundary conditions must be applied. As initial condition either the pressure head as a function of depth Z must be given

$$h(Z, t=0) = h_o(Z) \qquad \text{...(42)}$$

or the moisture content as a function of the depth Z

$$\theta\,(Z, t=0) = \theta_o(Z) \qquad \text{...(43)}$$

must be applied.

Boundary conditions at the top and the bottom of the unsaturated zone can be specified in three different ways:

(i) Dirichlet condition: the pressure head is specified as a function of time.

$$h(Z=Z_B,t) = h_B(t) \qquad ...(44)$$

(ii) Neumann condition: the flux is specified as function of time.

$$q(Z=Z_B,t) = q_B(t) \qquad ...(45)$$

(iii) Functional relationship: (Cauchy condition) the flux is a function of the dependent variable h at $Z=Z_B$.

$$q(Z=Z_B,t) = f(h_B,t) \qquad ...(46)$$

Use of this type is only possible in iterative computations. In general at the top and at the bottom of the unsaturated zone different types of boundary conditions can be used at the same time. Even during the process of computation the type of boundary conditions may be changed. For the unsaturated zone the boundaries are constituted by the soil surface and the phreatic surface. Through these boundaries relations can be established with the atmosphere and the saturated zone.

3.2.4 Saturated groundwater system

The phreatic surface acts as "lower" boundary of the unsaturated zone. The connection with the saturated zone is usually established by the flux-ground water table depth relationships. In the more traditional approach of evaluation of water management measures, changes in the water table depth are calculated without considering the actual response of the unsaturated zone and mostly an unique relationship between mean ground water table depth during the growing season and evapotranspiration/crop yield is taken (figure 4). With the use of numerical models, it is

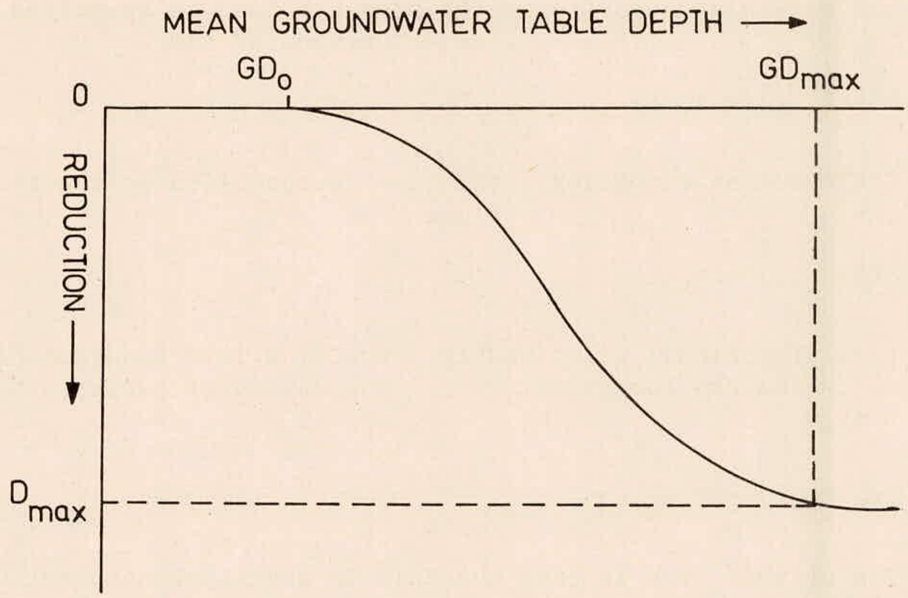


FIGURE 4 - EFFECT OF GROUNDWATER TABLE DEPTH, GD, DURING THE GROWING SEASON ON CROP YIELD REDUCTION

GD_o = lowest groundwater table depth where reduction is approximately zero;

GD_{max} = highest groundwater table depth where reduction is approximately at its maximum

[Source : Berg (1979)]

possible to consider both the saturated and unsaturated zone as one continuum.

3.2.5 Water uptake by plant roots

Water uptake by the roots is represented by a sink term, which is added to the continuity equation (36):

$$\frac{\delta\theta}{\delta t} = - \frac{\delta q}{\delta Z} - S \qquad (d^{-1}) \qquad \ldots(47)$$

where S represents the volume of water taken up by the roots per unit bulk volume of soil in unit time ($cm^3.cm^{-3}.d^{-1}$).

Feddes et al. (1978) considers the sink term as a function of the soil water pressure head, h. By definition the integral of the sink term over the rooting depth, Z_r, equals the actual transpiration rate.

$$T = \int_{Z=0}^{Z=Z_r} S(h)dZ \qquad (cm.d^{-1}) \qquad \ldots(48)$$

For optimal soil moisture conditions, T=PT and

$$S(h) = S_{max} = \frac{PT}{Z_r} \qquad (d^{-1}) \qquad .(49)$$

In non-optimal soil water conditions water uptake is reduced according to

$$S(h) = \alpha(h)\ S_{max} \ ; \qquad 0 \leq \alpha \leq 1 \qquad \ldots(50)$$

An example of the shape of $\alpha(h)$ is shown in figure 5. Little is known about the anaerobasis point, (h_1, figure 5), at which deficient aeration condition exists. In any case, it will depend on type of crop, temperature and type of soil.

For the h-range ($h_2 \rightarrow h_3$) where transpiration is at potential rate, different values of h_2 and h_3 are used. Some authors apply a fixed range, while others found a varying range, depending on the evaporative demand

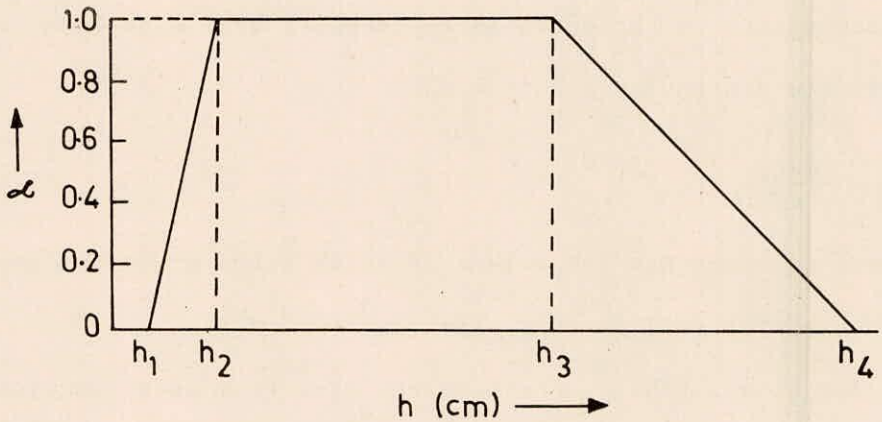


FIGURE 5 - RELATION BETWEEN SINK TERM VARIABLE α , $[\alpha(h) = S(h)/S_{max}]$ AND SOIL WATER PRESSURE HEAD, h

h_1 = anaerobiosis point;

h_2 = lowest value of h where $S = S_{max}$;

h_3 = highest value of h where $S = S_{max}$;

h_4 = wilting point

[Source : Feddes et al.(1978)]

of the atmosphere. For the wilting point (h_4) a value of h is taken to be -16,000 cm.

Incorporation of the sink term into equation (41) yields:

$$\frac{\Delta h}{\Delta t} = \frac{\Delta}{C(h)\Delta Z}[K(h)(\frac{\Delta h}{\Delta Z} + 1)] + \frac{S(h)}{C(h)} \qquad ...(51)$$

The numerical procedure for solution of equation (51) is essentially the same as for equation (41).

3.3 Integrated Approach to Evaluate Effects of Changes in Soil Water Conditions on Evapotranspiration

To optimize different schemes of water management, it is important to evaluate effects of water management upon evapotranspiration,crop production, water quality, composition of vegetation etc. Different approaches can be used to evaluate the effects of changes in soil water conditions on evapotranspiration. They range from methods that are based on semi-empirical relationships to physical-mathematical models that describe the system in a very detailed way. In the following an overview of the various categories of methods, now a days in use, has been presented.

3.3.1 Empirical methods

This category of methods has the advantage that they are simple to apply, but the disadvantage that they are not applicable to other areas than for which they were developed. Visser (1958) gives the yield depression for seven soil classes with various mean depths of the water table during the growing seasons (figure 6). Yield depressions at high water tables can be ascribed to lack of aeration of the soil, depressions at deeper water tables are due to shortage of water. Since both conditions may occur dependent on climatological circumstances, it is evident that the curves may undergo a horizontal shift to the right in wet years and

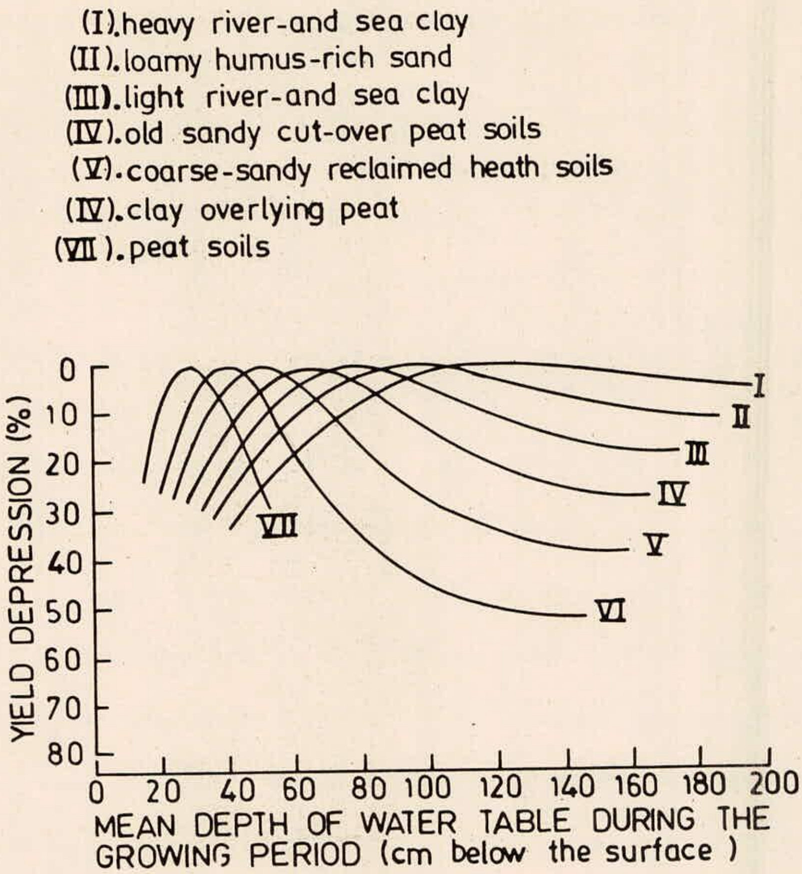


FIGURE 6 - THE EFFECT OF THE MEAN DEPTH OF THE GROUNDWATER TABLE DURING THE GROWING SEASON ON FINAL CROP YIELD FOR SEVEN GROUPS OF SOILS

[Source : Visser (1958)]

a shift to the left in dry years. These curves are reflecting the behaviour of the unsaturated zone and its effect through evapotranspiration on crop production. Hence the unsaturated zone is used here as a 'black box'. It should be remarked that attempts have been made to adopt the curves to changed circumstances.

3.3.2 Parametric models

This type of models is still empirical but make use of some properties of the unsaturated zone.

Grootentraast (see Berg, 1979) determined the yield depression of grassland on sandy soils due to an artificial drawdown of the groundwater table according to the principle presented in figure 4. For a number of years the relative evapotranspiration/yield is computed for conditions where no water table is present. This gives the maximum yield depression D_{max}. This point corresponds with a theoretical ground water table depth GD_{max} where the contribution from capillary rise approximately is zero. Then the ground water table depth GD_o is determined at which a reduction in evapotranspiration/yield never will occur and the points GD_{max} and GD_o are connected by a smooth curve. With this method a number of properties as the soil moisture retention curve and the hydraulic conductivity implicitly are taken into account. One of the most important parameters is the water storage ST of the unsaturated zone. Thornthwaite and Mather (1955) calculated the actual evapotranspiration ET according to

$$ET = - \frac{d\ ST}{dt} (cm.d^{-1}) \qquad \ldots(52)$$

under the assumption that

$$ET = c\ ST\ (cm.d^{-1}) \qquad \ldots(53)$$

and taking at $t = t_o$: $ST = ST_{t_o}$ and $ET = PET$, it follows that

$$c = \frac{PET}{ST_{t_o}} \qquad \ldots(54)$$

$$\text{so, } ET = \frac{ST}{ST_{t_o}} \, PET \; (\text{cm.d}^{-1}) \qquad \ldots(55)$$

Equation (55) applies to (vegetated) soil with deep ground water tables (no capillary rise). A reservoir model that takes into account the influence of the ground water table has been developed by a Working Group of ICW (1979).

For a unit area the water balance of the unsaturated zone over a certain period can be written as:

$$ST_t = ST_o + Q - ET \quad (\text{cm}) \qquad \ldots(56)$$

where

ST_t, ST_o is water storage of soil profile at $t = t$ and $t = 0$ respectively,

Q is water supplied from outside the system (infiltration, sprinkling, seepage),

ET is actual evapotranspiration.

Evapotranspiration ET depends on ST as follows:

$$ET = C \; PET \qquad \ldots(57)$$

$C = 1$ for $0 \leqq pF \leqq 3.2$

$C = \dfrac{ST - ST_{4.2}}{ST_{3.2} - ST_{4.2}}$ for $3.2 < pF < 4.2$

$C = 0$ for $pF = 4.2$

The value of ST is derived as

$$ST = \frac{ST_o + ST_t}{2} \qquad \ldots(58)$$

Because ST_t is not known beforehand, the procedure of calculation starts with a first estimation of ST_t, i.e. $ST_t = ST_o$. From equation (56) ET is computed. This value is substituted in equation (57) and a new value of ST_t is obtained. If this value differs significantly from the initial ST_t value, the computing process is repeated. Otherwise the next timestep is taken.

In the above mentioned two methods the reduction factor is independent of the magnitude of (potential)evapotranspiration. One can imagine that at low potential evapotranspiration rates reduction in evapotranspiration occurs at lower values of available soil water than at high potential evapotranspiration rates. Therefore some authors (e.g. Federer, 1979)found a relation between the reduction factor, and the ratio of available soil water (or availability factor) and potential (evapo)transpiration.

3.3.3 Physical-mathematical models

This group of models is based on physical processes and mathematical relationships.

(a) Steady-state models

These models can be applied when calculating percolation profiles or steady state capillary rise (evaporation) from the ground water table towards the root zone or the soil surface. Numerical integration of equation (35a) yields the relationship between q,z and h. An illustration of the effect of a disturbing coarse sand layer in a fine sandy soil on the capillary flux from the ground water table towards the root zone is presented in figure 7. From a computational point of view there is no problem in handling non-homogeneous soil profiles.

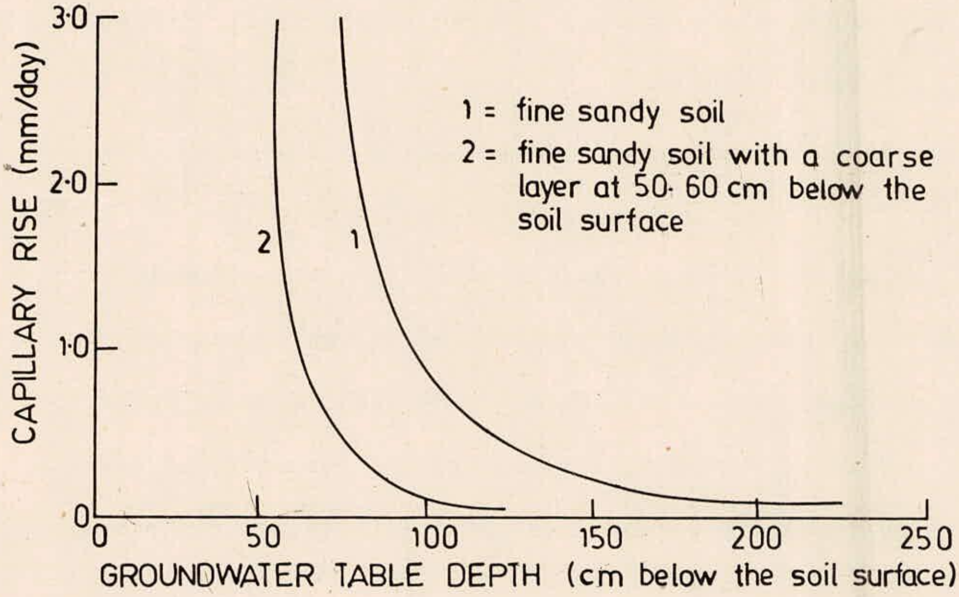


FIGURE 7 - RELATION BETWEEN GROUNDWATER TABLE DEPTH, GD, AND CAPILLARY FLUX, q, IN A HOMOGENEOUS (1) AND A HETEROGENEOUS PROFILE (2)

(The pressure head at the lower boundary of the root zone is - 1000 cm. in both cases)

[Source : Van Bakel (1981)]

(b) Pseudo steady-state models

A non-stationary process of water flow can be approximated by a subsequent series of steady-state flow situations. The pressure head profiles $h(z,q)$ can be converted into water content profiles $\theta(z,q)$ through the pF-curves. The $\theta(z,q)$ profiles can be integrated over a certain soil depth yielding water storage $ST(q)$. Changes in water storage of the soil profile are found from the difference in two subsequent steady-state situations, $ST(q_2)-ST(q_1)$.

Rijtema (1965) used a pseudo-steady state approach to compute relative evapotranspiration during the growing season. He distinguished zones in the soil profile: the root zone and the subsoil. In the root zone all water is taken to be transported through the roots, i.e. no vertical gradient over this zone exists. The subsoil is considered as a homogeneous single system. The model computes actual evapotranspiration over a certain period from the water balance of the root zone:

$$ET = P+Q-\Delta ST \qquad \text{(cm)} \qquad \ldots(59)$$

where,

ET is evapotranspiration,

P precipitation (infiltration),

Q capillary rise from the subsoil, and

ΔST change in available soil water, $\Delta ST = ST(q_2)-ST(q_1)$.

It is assumed that the evapotranspiration rate is at its potential value when the tension in the root zone is smaller than a certain limit h_1. When this limit is exceeded, a reduction in evapotranspiration occurs.

(c) Transient models

The non-stationary process of water flow can be approximated by a numerical solution of equation (51). In fact this approach is the most simple one as it needs less restricting assumptions. The traditional disadvantage of a large amount of computing time inherent to this approach becomes less and less significant as recent trends in computer technology are directed towards faster computers.

Feddes et al. (1978) developed a computer program based on equation (51) with the sink term described according to equation (49) and (50), for two layered soil profiles. Theoretical results predicted by the model were compared with a field experiment in which red cabbage was grown on a heavy clay soil in the presence of a water table. Water balance studies were performed with a specially designed non-weighing type lysimeter.

In figure 8 graphs of cumulative flow are given. They are (i) the measured cumulative evapotranspiration ($ET_{water\ balance}$) as obtained from the lysimeter, (ii) the cumulative transpiration T_{plant}^{comp} as computed with the model by integration of the sink term over depth, (iii) the cumulative soil evaporation E_{soil}^{comp} derived from the computed terms of the water balance.

The above mentioned model applies to one-dimensional vertical flow.

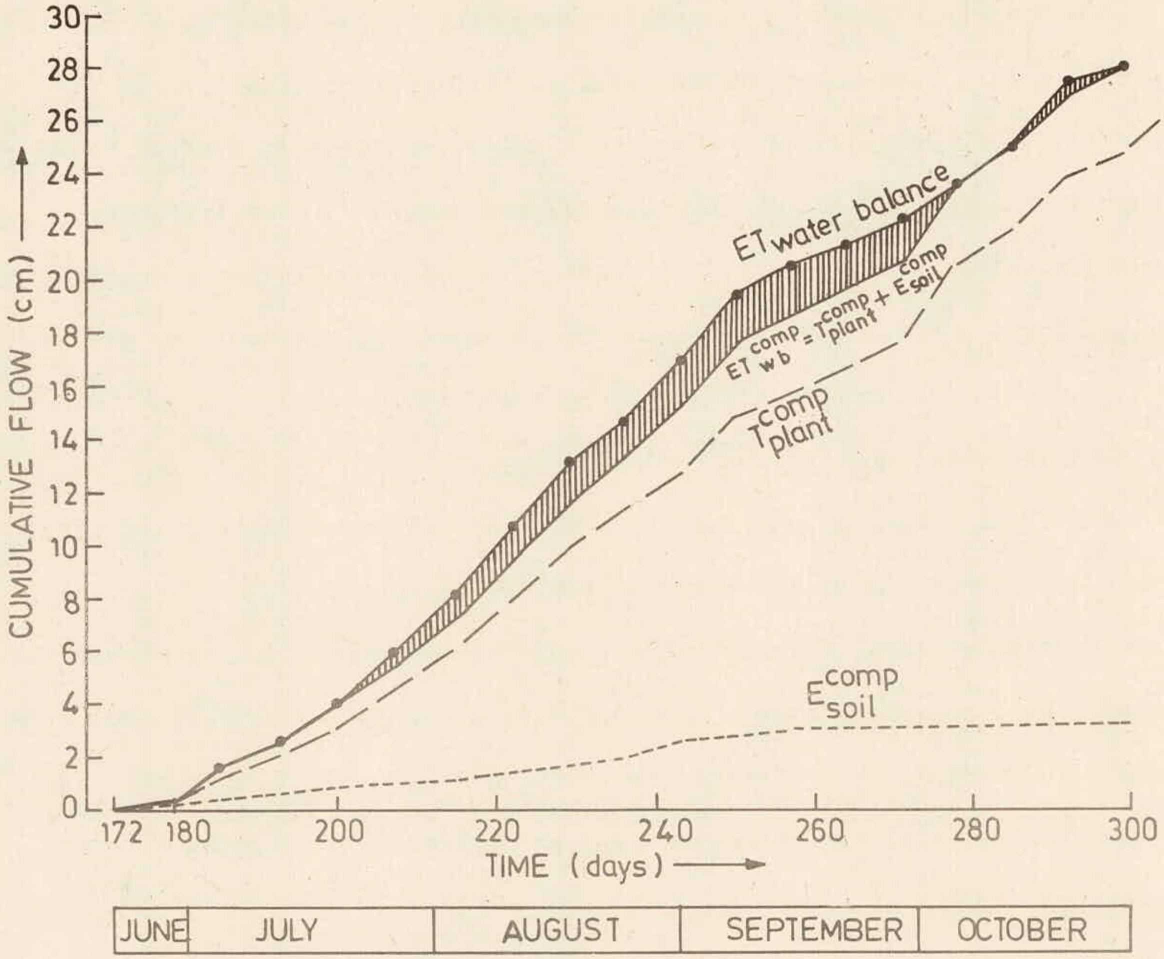


FIGURE 8 - COMPUTED CUMULATIVE EVAPOTRANSPIRATION E_{wb}^{comp}, TRANSPIRATION T_{plant}^{comp}, SOIL EVAPORATION E_{soil}^{comp} AND MEASURED CUMULATIVE EVAPOTRANSPIRATION $ET_{water\ balance}$

[Source : Feddes et al.(1978)]

4.0 METHODS OF ESTIMATING EVAPOTRANSPIRATION

4.1 General

Evapotranspiration is a major component of the hydrologic cycle and is involved to some degree in nearly all hydrologic studies. It is important in the planning and development of water resources in a river basin. It forms the foundation for the planning and design of most irrigation projects, since it is usually the starting point in determining surface and subsurface storage requirements, the capacity of the delivery system, and general operation practices. Evapotranspiration plays a major role in managing water quality and soil salinity.

The tremendous and continuing need for ET data has resulted in numerous methods for estimating ET. Many of these methods consider only a few variables since estimates often were needed where limited meteorological data were available. However, regardless of how quickly the estimates are needed or how limited the input data are, engineers should apply basic science, physics and engineering principles in developing ET estimates.

4.1.1 Seasonal evapotranspiration

Numerous studies over the past half-century have shown that if soil water does not significantly limit crop growth and yield of the product, the total water evaporated from planting to harvest for a given crop is about the same on all soils under similar climatic conditions. This quantity of water, usually expressed as volume per unit area is called seasonal evapotranspiration, or seasonal consumptive use. In contrast, the total amount of water applied to a crop per unit land area can vary greatly because it includes both ET and deep percolation.

The magnitude of daily ET from a green crop is controlled primarily by meteorological conditions when leaf area and soil water are not limiting factor. In contrast, the magnitude of deep percolation is influenced by the amount of irrigation water applied at each irrigation, rainfall, soil characteristics, and the soil water level maintained. The major impact of ET and deep percolation on reported water requirements is that ET is limited at any time by the heat energy available for evaporation and tends to fall within predictable limits. In contrast, deep percolation is not so limited, but may vary widely depending on irrigation practices and the management of the system. As a result, ET data for a given crop will be similar in areas of similar climate and growing seasons, but if water requirement data represent ET plus deep percolation and other losses, much greater variations are encountered. Also, since the magnitude of deep percolation and other losses are influenced by man, they are less predictable.

4.1.2 Annual evapotranspiration

Annual ET includes seasonal and off-season evapotranspiration. The off-season period is from the harvest of one crop to the planting of the next crop where only one crop is grown each year. This parameter has received much less attention in irrigation studies in the past, but it can be very important in optimizing the use of a limited supply of water for irrigation since stored soil water may be used by a deep-rooted crop and excess water applied to a shallow-rooted crop may replace deep soil water.

The total quantity of water evaporated from a cropped field over a given time period, w_{et}, can be expressed as

$$w_{et} = \int_o^t (ET)\,d\tau = \int_o^t K_c(PET)\,d\tau \qquad \ldots(60)$$

where K_c is the crop coefficient.

4.1.3 Distribution system

Water can not be applied with perfect uniformity nor without some unavoidable losses with most irrigation systems. Surface runoff may be inherent with some irrigation methods. Level or nonuniformly sloped fields, diked at the lower end may require little or no runoff to adequately irrigate the lower end and may enable reducing deep percolation near the upper end. Similarly additional water may be required with sprinkler irrigation systems in addition to direct evaporation from the spray to compensate for non-uniform applications.

In addition to allowances for non-uniform water distribution, there is evaporation and leakage from the storage and distribution systems. A substantial portion of the leakages may be recovered for use within a project or for projects downstream. Some of these leakages are illustrated in figure 9. The magnitudes of evaporation and nonbeneficial ET from leakage sources are greatly dependent on local subsurface soil and geological conditions.

4.2 Factors Affecting Evapotranspiration

In as much as evapotranspiration includes the sum of volumes of water used by both evaporation and transpiration processes, it is obvious that many of the factors, primarily climatic factors, which influence the amount of evaporation from a free water surface, also affect the amount of evapotranspiration, for example, solar radiation intensity and duration, wind conditions, relative humidity, cloud cover, atmospheric pressure and others. In addition to the climatic factors, however both soil and vegetative factors govern the evapotranspiration from an area.

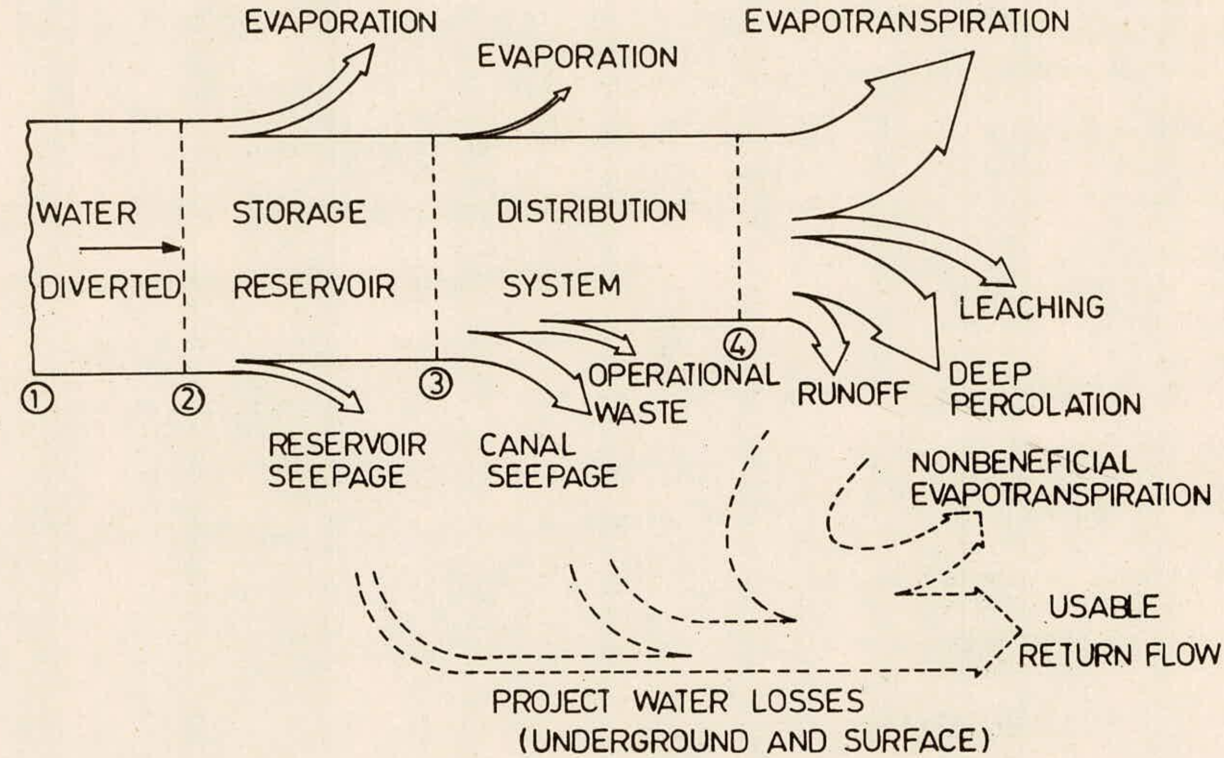


FIGURE 9 - DISPOSITION OF WATER DIVERTED FOR IRRIGATION

[Source : Jensen (1967)]

4.2.1 Climate

(a) Variation with time

It is common practice to use mean climatic data for determining mean evapotranspiration. However, due to weather changes, ET will vary from year to year and for each period within a year. Annual ET will vary some 10 percent for humid tropics upto some 25 percent for mid-continental climates. From year to year, the monthly values show greater variation. For instance, in mid-latitude climates radiation for a given month can show extreme variations. In areas having distinct dry and wet seasons, the transition month shows significant differences from year to year depending on rains arriving early or late. Monthly ET values can vary from one year to the next by 50 percent or more. Daily values can vary drastically, with low values on days that are rainy, cloudy, humid and calm and with high values on dry, sunny and windy days. This variation will obviously be obscured when using mean climatic data to obtain mean ET.

(b) Variation with distance

For calculating ET, climatic data are sometimes used from stations located some distance away from the area under study. This is permissible in areas where the same weather extends for long distances. Zones with rapid changes in climate over short distances frequently occur, for instance in arid areas inland from large lakes and where an airmass is forced upward by moutain ranges. With the change in weather over distance consequently ET may change markedly over small distances. A check needs to be made on whether climatic data used from distant stations are representative for the area of study. No generalized guidance can be given on use of data from distant stations.

(c) Variation with size of irrigation development, advection

Meteorological data used,are often collected prior to irrigation development in stations located in rainfed or uncultivated areas, or even on rooftops and airports. Irrigated fields will produce a different micro climate and ET may not be equal to predicted values based on these data. This is more pronounced for large schemes in arid, windy climates. In arid and semi-arid climates, irrigated fields surrounded by extensive dry fallow areas are subject to advection. Airmass moving into the irrigated fields gives up heat as it passes over. This results in a 'clothesline' effect at the upwind edge and an 'oasis' effect inside the irrigated field. With warm, dry winds, appreciably higher ET can be expected at the upwind edge of the field. With increased distance the air becomes cooler and more humid. The 'clothesline' effect become negligible with distance from the border which may extend in hot, dry climates for 100 to 400 m for wind speed greater than 5 m/sec.

Due to the 'oasis' effect, ET will be higher in fields surrounded by dry fallow land as compared to surrounded by extensive vegetated area. Using climatic data collected outside or prior to irrigation development, figure 10 suggests the correction factors needed to obtain ET for irrigated fields of different sizes located in dry fallow surrounds in arid, hot conditions with moderate wind. Factors should not be applied to very small fields (< 0.05 ha) since the correction on ET could be large enough to result in wilting of the crop and stunting of growth.

(d) Variation with altitude

In a given climatic zone, ET will vary with altitude. This is not caused by difference in altitude as such but mainly by associated changes in temperature, humidity and wind speed. Also radiation at high altitudes

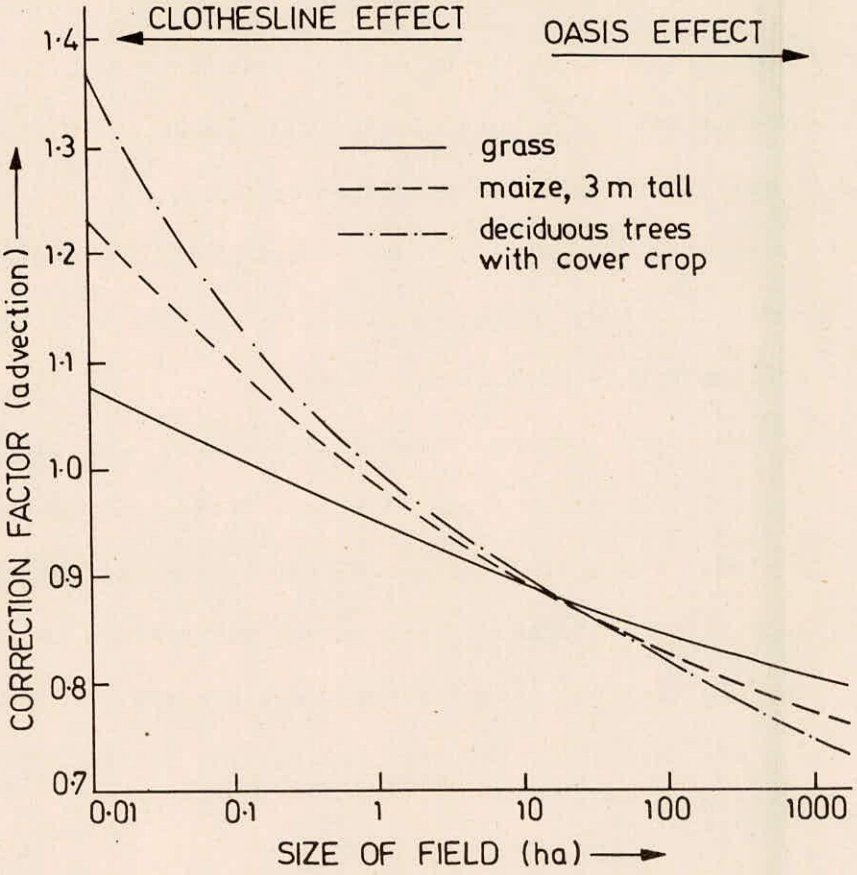


FIGURE 10- CORRECTION FACTOR FOR EVAPOTRANSPIRATION WHEN DETERMINED USING CLIMATIC DATA COLLECTED OUTSIDE OR PRIOR TO IRRIGATION DEVELOPMENT, FOR DIFFERENT SIZES OF IRRIGATED FIELDS UNDER ARID AND MODERATE WIND CONDITIONS

[Source : Doorenbos and Pruitt (1977)]

may be different from that in low lying areas.

4.2.2 Soil water

Published data on depth over which the crop extracts most of its water,show great differences. With salt-free soil water in ample supply, water uptake for most field crops has been expressed as 40 percent of total water uptake over the first one-fourth of total rooting depth, 30 percent over the second one-fourth, 20 percent over the third and 10 percent over the last. However, movement of soil water will take place inside and to the root zone when portions become dry. Also water can be supplied to the roots from shallow groundwater.

(a) Level of available soil water

After irrigation or rain, the soil water content is reduced primarily by evapotranspiration. As the soil dries, the rate of water transmitted through the soil will reduce. The effect of soil water content on evapotranspiration varies with crop and is conditioned primarily by type of soils and water holding characteristics, crop rooting characteristics and the meteorological factors determining the level of transpiration. When evaporative conditions are lower the crop may transpire at the predicted ET rate even though available soil water depletion is greater.When evaporative conditions are higher, ET will be reduced if the rate of water supply to the roots is unable to cope with transpiration losses. This will be more pronounced in heavy textured than in light textured soils. Following an irrigation the crop will transpire at the predicted rate during the days immediately following irrigation. With time the soils become drier and the rate will decrease, more so under high as compared to low evaporative conditions.

(b) Ground water

For most crops, growth and consequently ET will be affected when ground water is shallow or the soil is waterlogged. In spring in cooler climates, wet soils warm up slowly, causing delay in seed germination and plant development; land preparation may be delayed, resulting in later planting. Consequently, different ET values apply during the remainder of the season. The tolerance of some crops to shallow ground water tables and waterlogging is given in table 2.

Table 2 - Tolerance Levels of Crops to High Groundwater Tables and Waterlogging

	Ground water at 50 cm	Waterlogging
High tolerance	sugarcane,potatoes, broad beans	rice,willow,strawberries, various grasses,plums
Medium tolerance	sugarbeet,wheat,barley, oats,peas,cotton	citrus,bananas,apples, pears,blackberries, onions
Sensitive	maize,tobacco	peaches,cherries, date palms,olives, peas,beans

[Source:Irrigation, Drainage and Salinity, An International Source Book, FAO/UNESCO, 1973]

Higher ground water tables are generally permitted in sandy rather than loam and clay soils due to the difference in capillary fringe above the ground water table. For most crops minimum depth of ground water table required for maximum yield has been expressed as : for sand,rooting depth +20 cm; for clay, rooting depth +40 cm; for loam, rooting depth +80 cm.

(c) Salinity

ET can be affected by soil salinity since the soil water uptake by the plant can be drastically reduced due to higher osmotic potential of the saline ground water. Poor crop growth may be due to adverse physical characteristics of some saline soils. Reduced water uptake under saline conditions is shown by symptoms similar to those caused by drought, such as early wilting, leaf burning, a bluish-green colour in some plants, reduced growth and small leaves. The negative effect of soil salinity can be partly offset by maintaining a high soil water level in the root zone.

4.2.3 Method of irrigation

ET is affected little by the method of irrigation if the system is properly designed, installed and operated. The advantages of one method over another are therefore not determined by differences in total irrigation water supplied but by the adequacy and effectiveness with which crop requirements can be met.

Different methods imply different rates of water application. When comparing the various methods in terms of water efficiency in meeting crop demand , such differences should be recognised, the apparent superiority of one method over another may be merely the result of too much or too little water being applied.

4.3 Methods of Measuring Evapotranspiration

4.3.1 Soil water depletion

Evapotranspiration under field conditions can be determined by observing changes in soil water over a period of time. This method is used primarily by soil sampling and gravimetric analyses. Now the neutron soil moisture probe has essentially replaced the gravimetrical procedure except for evaluating soil water in the surface 10 to 20 cm. The major potential source of error in evapotranspiration determined by this method is drainage from the zone sampled or upward movement from a saturated zone. Both are difficult to detect, but can be minimized with proper precautions. The soil is usually sampled two to four days after an irrigation and again 7 to 15 days later or just before the next irrigation. Since the change in soil water over the time interval is desired, the holes from which the soil cores are removed (gravimetric procedure) are usually filled with soil, marked, and the next sample taken 30 to 40 cm from the first core in order to minimize error due to soil variability. Access tubes are used with the neutron probe and the same site is measured each time.

The average rate of ET between sampling dates is calculated using the following equation:

$$ET = \frac{w_{et}}{\Delta t} = \frac{\sum_{i=1}^{n_r} (\theta_1 - \theta_2)_i \; \Delta S_i + R_e - w_d}{\Delta t} \qquad \ldots(61)$$

where n_r is the number of layers to the depth of the effective root zone, ΔS is the thickness of each layer, mm, θ_1 and θ_2 are the volumetric water content on the first and second date of sampling respectivley, R_e is rainfall that does not runoff the area, mm, and w_d is drainage from the zone sampled, mm, (negative for upward flow to the sampling zone).

Reliable ET rates determined by soil sampling require adequate precautions such as -

(1) at least 6 sampling sites representative of general field conditions are used, a minimum of 4 may be adequate when using neutron techniques;

(2) depth to the water table should be much greater than the root zone depth;

(3) only those sampling periods when rainfall is light, are used, all others are questionable because drainage (w_d) may be excessive;

(4) drainage is minimized by:

(a) giving the preplant irrigation at least 10 days before planting,

(b) applying less water at each irrigation than the amount that could be retained,

(c) waiting at least 2 days after a normal light irrigation before taking the first sample, and longer if excessive irrigations were involved and when evapotranspiration is small, and

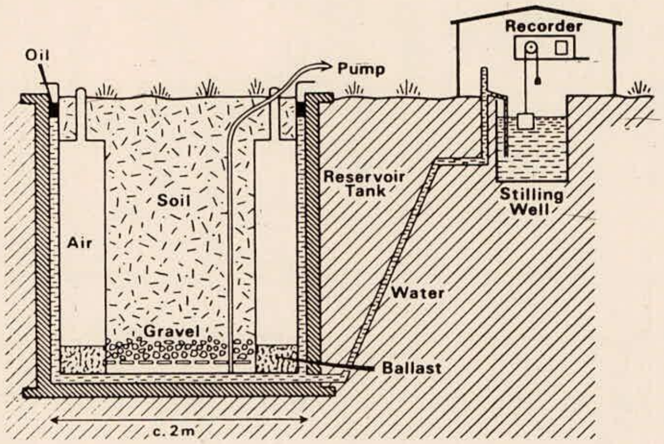
(d) only the active root zone depth is used for ET computations.

4.3.2 Tanks and lysimeters

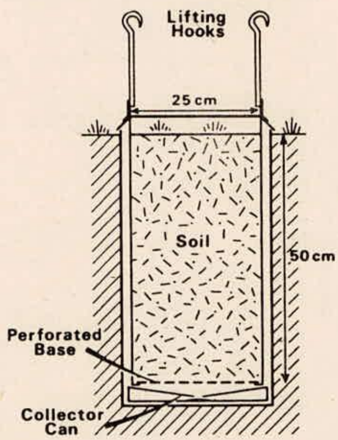
Lysimeters (Evapotranspirimeters) are tanks filled with soil (figure 11) in which crops are grown under natural conditions to measure the amount of water lost by evaporation and transpiration. This method provides the only direct measurement of ET and is frequently used to study climatic effects on ET and to evaluate estimating procedures. However, all lysimeter data are not representative of field conditions. Soil conditions inside the lysimeters must be essentially the same as those outside. The lysimeter must be surrounded by the same crop that is growing in the lysimeter, located within a field of the same crop, and at least 100 m from the edge of the field.

Lysimeters can be grouped into three categories:

(1) nonweighing, constant water-table type, which provides reliable data in areas where a high water table normally exists and where the water table level is essentially the same inside and outside the lysimeter;



HYDRAULIC LYSIMETER



SIMPLE WEIGHING LYSIMETER

FIGURE 11 - LYSIMETERS

(2) nonweighing percolation type, in which changes in water stored in the soil are determined by sampling or neutron methods and the rainfall and percolation are measured. These units are often used in areas of high precipitation; and

(3) weighing types, in which changes in soil water are determined either by weighing the entire unit with a mechanical scale, counter-balanced load cell, or by supporting the lysimeter hydraulically. Weighing lysimeters have recently become more popular and generally provide the most accurate data for short time periods. Evapotranspiration can be determined accurately over periods as short as one hour with a mechanical scale or load cell system. Hydraulically weighed lysimeters generally are not accurate for periods less than 24 hours

4.3.3 <u>Percolation gauges</u>

These are instruments specially designed for measuring evaporation and transpiration from a vegetated surface, ET, and are comparable with the tanks and pans used for measuring evaporation. Similarly, there are very many different designs and, in general, these are regarded as research tools rather than standard instruments to be installed at every climatological station (figure 12). A cylindrical or rectangular tank about 1 m deep is filled with a representative soil sample supporting a vegetated surface and is then set in the ground. A pipe from the bottom of the tank leads surplus percolating water to a collecting container. The surface of the gauge should be indistinguishable from the surrounding grass or crop covered ground. A raingauge is sited nearby and the evaporation plus transpiration is given by the following equation:

$$ET = \text{Rainfall} - \text{Percolation} \qquad \ldots(62)$$

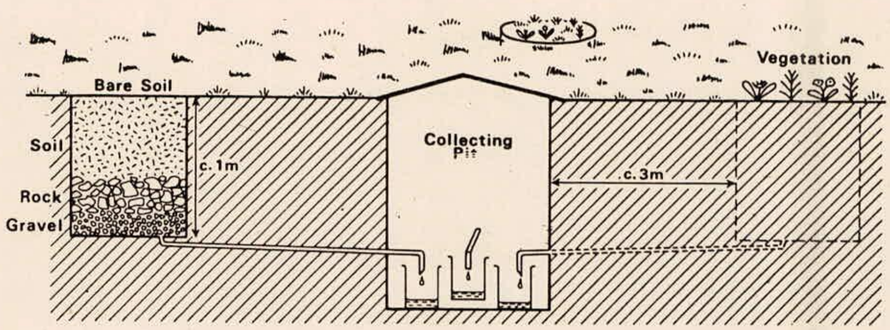


FIGURE 12 - PERCOLATION GAUGE

Percolation gauges do not take into account changes in the soil moisture storage and thus measurements should be made over a time period defined by instances when the gauge is saturated so that any difference in the soil moisture storage is small. Records are generally compiled on a monthly basis in climates with rainfall all the year round.

4.3.4 Water balance approach

This has been applied at a number of different scales including lysimeters, water table fluctuations, river basins and small experimental watersheds, and moisture fluctuations within the soil profile.

(i) An approximate measure of long-term evapotranspiration may be determined by considering the major water balance components of a river basin. Assuming that over a period of one year sub-surface storage changes are negligible, a simplified water balance equation may be written:

$$\text{evapotranspiration} = \text{precipitation} - \text{runoff} \quad \ldots(63)$$

For any but very small basins, however, there are difficulties in obtaining reliable values of precipitation from existing gauge networks and also storage/lag problems at each year end when late precipitation may not appear as runoff until the early months of the following year.

(ii) Some of the problems inherent in river basin determinations of actual evapotranspiration are avoided in small experimental watersheds in which detailed measurements are made of a wide range of parameters. In this case the main concept is that if the water balance equation $P-Q-ET-\Delta S-\Delta G=0$ (where P is precipitation, Q is streamflow, ET is actual evapotranspiration and ΔS and ΔG are changes in soil moisture and ground water storage respectively) can be solved, then it is likely that the measurements or estimations of the individual components of that balance are satisfactory. Evidently, this may be an erroneous supposition if discrepancies in the assessment of individual parameters are fortuitously complementary, although this source of error can usually be guarded against by making specimen water balance calculations for different time periods within the same run of data. Since precipitation, streamflow, soil moisture and ground water measurements can almost certainly be made with a greater accuracy than the corresponding measurements or estimations of evaporative losses, it can be argued that the value of ET which consistently gives the best result in the water balance equation is the most "suitable".

(iii) A further check on the small watershed water balance can be provided by calculating the partial water balance for the soil profile only. The important work done by Thornthwaite and his colleagues on the climatic water balance has substantiated the close relationships and inter-relationships between P, ET and Δ S. In the final analysis an approach through soil moisture measurement and particularly the determination of successive soil moisture profiles, may well hold out most hope for accurate determination of actual evapotranspiration since it is only within the soil profile that one can obtain a direct measure of the amount of water withdrawn by vegetation cover. The principal disadvantage of this approach is the need for considerable replication of moisture measurements, although this problem is eased by using a neutron probe, and by the careful choice of measuring sites where dominantly lateral (as opposed to vertical) movements of moisture within the profile are not present. In many forested areas where lysimeter studies are not practicable, neutron soil moisture measurements may represent the only realistic approach to the measurement of evapotranspiration.

An extension of the soil profile water balance approach concerns the determination of salt movement within the soil profile to use it as an index of evapotranspiration.

4.3.5 Moisture flux measurements

Determinations of actual evapotranspiration from measurements of the flux of moisture above the evaporating surface have been investigated in a number of different situations.

(i) A theoretical approach was developed in detail by Thornthwaite and Holzman (1939). This approach is based upon the fact that evapotranspiration into the lower layers of the atmosphere will tend to establish a moisture gradient from the evaporating surface into the overlying air, and that turbulent motion will tend to break that gradient down and so establish uniform moisture conditions above the evaporating surface. If then both the moisture gradient and the turbulent motion of the air can be accurately measured, the contribution of water vapour from the evapotranspiration process can be deduced.

Instrumental limitations and the non-generality of the formulae were at first decisive in preventing the wider development of this approach. Technological developments have reduced many of the instrumental limitations, at least in relation to research investigations. At the present time direct measurement of the moisture flux is being investigated in two main ways: namely, leaf chamber studies and the eddy correlation method.

(ii) In leaf chamber studies, independent, continuous and simultaneous measurements are possible of the water vapour and carbon dioxide exchanges between each surface of a leaf and the surrounding atmosphere under controlled conditions of visible and total radiation, air and leaf temperatures, and carbon dioxide and water vapour concentrations. Field measurements from an entire plant or collection of plants are possible in larger plant chambers. This is a clear plastic chamber of 6 cubic decimetres volume which may be placed over a small living plant in the field and within which changes in relative humidity are measured by a hygrosensor and registered on a sensitive galvanometer. Such measurements will undoubtedly help towards a better understanding of the transpiration process although in terms of a meaningful measurement of actual water loss from a natural surface,those from individual leaves may be of comparatively little value. Thus, within a vegetation cover, leaves do not exist in isolation but in such large numbers that mutual interference in terms of incoming and outgoing radiation, humidity and sensible heat must occur: again, leaf and in some cases plant chamber studies are concerned primarily with the final phase of the transpiration process only, i.e. the conversion of liquid water within the plant to water vapour and its subsequent transfer into the atmosphere, thereby ignoring many of the preceding stages of moisture movement and moisture stress within, say, the soil and root systems.

(iii) Many of the weaknesses of the leaf and plant chamber approach are avoided when direct measurements are made of the moisture flux over an existing, natural vegetated surface. Fluxes could be determined from the correlation of temperature and humidity fluctuations with the vertical component of wind velocity. In order to avoid laborious data computations it was necessary to develop an instrument in which the relevant calculations were carried out instantaneously. It was desirable to design a small, portable instrument, subsequently called the "Evapotron", which could be used in the field. Preliminary tests were carried out in 1961 and the instrument was later used in extensive investigations as reported by Dyer and Pruitt (1962). By means of delicate sensing devices simultaneous measurements are made of the minute eddy fluctuations of humidity, wind and temperature above the evaporating surface. This information may be fed directly into a computer and an output of net upward movement of water vapour from the evaporating surface obtained.

Although such instruments are still clearly in the process of development, they will undoubtedly eventually come into widespread use as standard measuring devices. Ironically, this is likely to raise almost as many problems as it solves, particularly in connection with the representativeness of the measuring sites and the degree of replication required, since the more sophisticated the measured data are, the less likely are they to be broadly representative of surrounding conditions. However, despite the instrumental and theoretical progress which has been made in the field of evapotranspiration measurement, little work has been done on the problem of applying the point measurement procedures to watershed size areas.

4.4 Estimating Evapotranspiration

Engineers are required to "estimate" ET using historical meteorological and cropping conditions, or predict future ET. Both involve meteorological data, but predictions are based on expected meteorological and cropping data. The accuracy of ET estimates depends primarily on the ability of the equations being used to describe the physical laws governing the processes and the accuracy of the meteorological and cropping data. This also applies to predicted ET.

The expected reliability of predicted data can be evaluated by considering daily evapotranspiration to be a random variable for a given day, week or month. The best estimate of the expected value for any future time period is the historical mean for the corresponding time period if this is known. Therefore, expected ET can be expressed as a function of time if historical data are available. However, most engineers are concerned with predicting ET in areas where historical data are not available. The usual procedure is to correlate observed ET with a parameter calculated from meteorological data such as air temperature, humidity, wind, percent of sunshine, etc. For example, two jointly and normally distributed random variables can be considered. The first random variable can be considered as the potential evapotranspiration in an area for a given time period and the second random variable a single or composite meteorological parameter. Equations describing these random variables are

$$m_1 = E[X_1] \qquad \text{...(64)}$$

$$m_2 = E[X_2] \qquad \text{...(65)}$$

$$\sigma_1^2 = Var[X_1] \qquad \text{...(66)}$$

$$\sigma_2^2 = Var[X_2] \qquad \text{...(67)}$$

where m_1 and m_2 are the mean of the first variable which we will consider potential evapotranspiration and the mean of the meteorological parameter, respectively. These are defined as being equal to the expected value of the random variables and σ_1^2 and σ_2^2 are the variance of the first and second random variables, respectively.

Equation (64) indicates that if historical data are available to adequately describe evapotranspiration, then the best estimate of predicted ET is the mean of historical ET for the respective time period. Similarly, the best estimates of expected meteorological conditions are the historical means.

When no historical ET data are available, basically "conditional expectation" is utilized to predict ET. This is described by the equation

$$E[X_1|X_2=x_2] = m_1+r\,\frac{\sigma_1}{\sigma_2}(x_2-m_2) \qquad \text{...(68)}$$

This equation states that the best estimate of the first random variable, X_1, given the value of the second random variable, X_2, as x_2(in this case x_2 represents the meteorological variable parameter), is equal to the mean of the first random variable plus the correlation coefficient, r, times the ratio of the standard deviations and the difference between the actual second variable and mean of the meteorological variable. In simple terms, this equation states that the best estimate of evapotranspiration in some areas will be equal to mean evapotranspiration in a known area, m_1, plus an adjustment for meteorological conditions, x_2 , that are different from that of the first site, m_2. Another important factor that must be considered in both predicting and estimating evapotranspiration , is that the standard deviation of the mean decreases with the number of values represented by the mean.

In the case of evapotranspiration, the standard deviation of the mean for a given time period will vary with the number of days represented in the period as indicated below.

$$\sigma_{\bar{x}} = \frac{\sigma_x}{\sqrt{n}} = \frac{\sigma_x}{\sqrt{\Delta t}} \qquad \ldots(69)$$

where $\sigma_{\bar{x}}$ is the standard deviation of the mean of variable x, n is the number of values used (in this case daily values) and Δt equals the number of days in the time period. Equation (69) indicates that the natural variation in the mean value for a given time period will become less as the number of days in the time period increases. Therefore, as the length of the time period being considered increases, the correlation coefficient relating ET to a meteorological parameter also increases significantly. For example, seasonal ET may correlate closely with the sum of temperatures since both increase with time, but weekly or daily values may be poorly correlated.

The potential for ET (PET) is most often defined as an atmospheric determined quantity. This assumes that the ET flux will not exceed the available energy from both radiant and convective sources. This is generally a workable assumption for most predictive methods and allows considering the atmospheric variables apart from the plant and soil effects. However, interactions and feedback from the plant and soil to the atmosphere do not allow complete isolation. The definition of PET for predicting ET from irrigated fields has often been that from a well-watered reference crop.

In this approach, some crop and soil variables become partially integrated into the PET values which make them less generally applicable than the PET values derived from atmospheric variables.

For most hydrologic applications, estimates of PET are based on daily, weekly, or sometimes even monthly measurements. These may include one or more atmospheric variables, such as solar or net radiation , air temperature and humidity, or some related measurement such as pan evaporation. Direct measurement or estimation of variables such as vapour or heat flux is difficult and not yet practical for most applications. Others, like radiation, have only been routinely measured for a relatively few years, and even now are not commonly available. As a result, most procedures for estimating PET are empirically based on atmospheric-related variables or methods. Fortunately, the primary causative atmospheric variables are relatively conservative in space where strong topographic or cultural changes are absent, thus PET estimates can often be transferred some distance with minimal error. For hydrologic applications, transfer is often necessary because data are not available on the area where needed.

There are numerous reported methods for estimating evapotranspiration, each one of them being unique because of its basis, data requirements, area of application, and accuracy. The following are a few of the methods which are often applied in hydrologic predictions.

4.4.1 Evaporation methods

Many studies over several decades have suggested the use of pan evaporation data to estimate potential evapotranspiration using a simple proportional relationship,

$$PET = C_{ET} \times E_{pan} \qquad \ldots(70)$$

where C_{ET} is evaporation pan coefficient, dependent on the reference crop and type of pan involved. Such studies have led to the use of pans in programs designed to aid farmers in scheduling irrigations.

The extreme care is needed in interpreting pan evaporation data to obtain reliable estimates of PET. The value of C_{ET} in equation (70) is very much a function of the kind of pan involved, the pan environment in relation to nearby surfaces, obstructions, etc., and the climate itself. Briefly, the size of the pan may affect C_{ET} very significantly in dry climates when pans are surrounded by several meters or more of dry-surface areas. Such is not the case in humid climates, or even in dry climates when pans are located within large cropped fields. Of greater importance is the fact that even for a given pan, significantly lower values of C_{ET} have been noted for drier, windy climates as compared to humid, calmer climates. The value of C_{ET} increases substantially in going from very low relative humidity areas to highly humid areas for any given wind level. A larger change occurs when going from calm to very windy conditions(C_{ET} decreases as wind speed increases).

In addition to the variation of coefficients with wind and humidity levels, there is also an interaction with the level of radiation. Since smaller coefficients under drier and windier conditions are largely the result of greater response of pans to advection as compared to crops (short, smooth ones at least), it is obvious that the relative effect (and hence, coefficients) should be greater under the lower radiation conditions of fall, winter and spring than under higher radiation, mid-summer conditions. Where careful standardization of pan environment is maintained, and where strong, dry-wind conditions occur only occasionally, the use of pans may be very reliable. When careful attention is paid to pan environment, mean monthly PET should be predictable with $\pm$ 10%

or better for most climates.

If differences in the factor C_{ET} resulting from physiological variation of the crop during the growing season are disregarded, table 3 can be used as a first approximation.

Table 3 - Values of C_{ET}, in $PET = C_{ET} \times E_{pan}$

Crop	Values of the factor C_{ET}, in $PET=C_{ET} \times E_{pan}$					
	Humid			Arid or semi-arid		
	Temperate winter	Temperate summer	Tropical	large area winter	large area summer	small area (less than 1 ha) summer
Wet, after rain or irrigation	0.9	1.0	1.0	1.0	1.2	1.5
Short grass	0.7	0.8	0.8	0.8	1.0	1.2
Tall crop (wheat, sugarcane)	0.8	1.0	1.0	1.0	1.2	1.5
Rice	1.0	1.0	1.2	1.0	1.3	1.6

[Source: van der Molen (1971)]

4.4.2 Energy budget

Methods of estimating PET based on the vertical energy budget of a vegetated surface have a physical basis because energy limits ET where moisture is readily available and the necessary vapour transport occurs. Figure 13 shows the major components of the energy budget which form the basis for the several methods that use this approach. Except for cases of significant advection, like field edges and oasis effects, the horizontal components are usually negligible. The budget (in $cal/cm^2/min$, except as noted)of the major vertical components may be expressed as

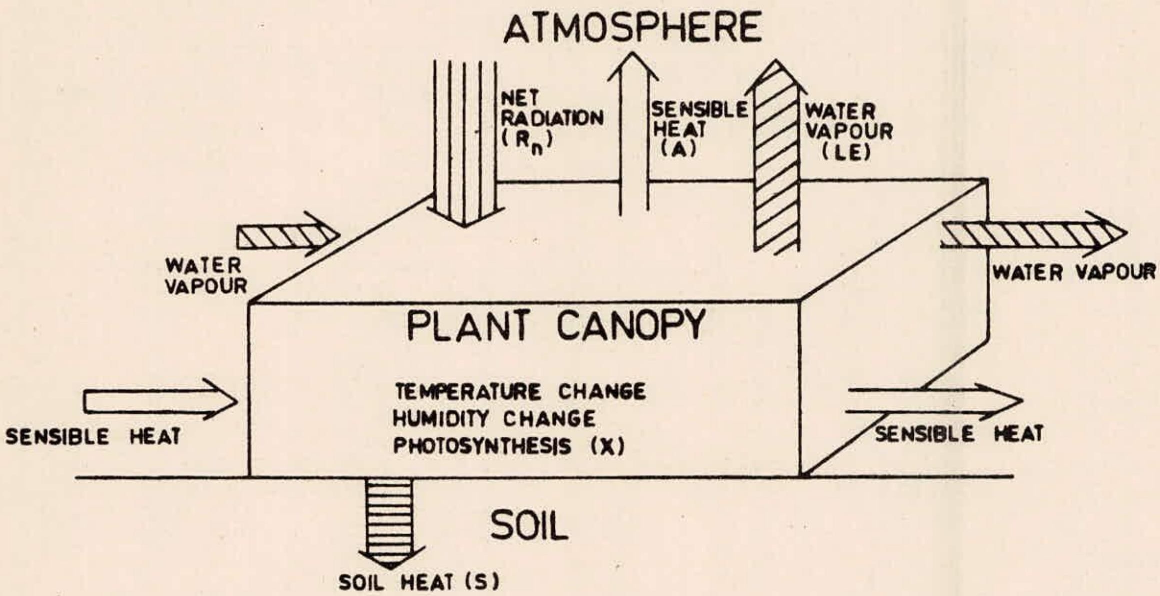


FIGURE 13 - ENERGY BUDGET OF A VEGETATED SURFACE

$$R_n = A+LE+S+X \qquad \ldots(71)$$

and $$R_n = R_s - aR_s + R_l - R_{lr} \qquad \ldots(72)$$

where,

R_n = net radiation,

R_s = incoming solar radiation (short-wave),

aR_s = solar radiation reflected,

R_l = incoming solar radiation (long-wave),

R_{lr} = emitted long wave radiation,

A = sensible heat of air,

LE = latent heat of water vapour,

L = latent heat of vapourization, cal/cm^3 or cal/g,

E = depth of evaporative water, cm^3/cm^2/min,

S = soil heat,

X = miscellaneous heat sinks, like plant and air heat storage, photosynthesis etc.

Recognizing that A and LE terms are of much greater magnitude than S and X, Bowen (1926) proposed using the ratio of sensible to latent heat

$$\beta = A/(LE) \qquad \ldots(73)$$

where β is commonly referred to as the Bowen ratio. Equation (72) with S and X neglected and equation (73) substituted, becomes

$$LE = R_n/(1+\beta) \qquad \ldots(74)$$

The value of β can be calculated from gradients of air temperature and vapour pressure above the evaporating surface, which is a relatively difficult measurement usually made only at research installations. Typical β values range from 0.1 to 0.3 for moist conditions (Priestley and Taylor, 1972).

4.4.3 Radiation methods

Because solar radiation provides the required energy for the phase change of water and often limits the ET process where water is readily available, a number of methods have been developed to estimate PET using a radiation data base. Most often, direct solar radiation is used because this is frequently measured by national weather networks, and air temperature has been found through correlation to be a useful second variable. A few such methods will be discussed.

(a) Makkink

Makkink (1957) presented the following equation for estimating ET for grass over 10-day periods under cool climatic conditions of The Netherlands.

$$ET_g = 0.61 \frac{\Delta}{\Delta+\gamma} \frac{R_S}{58.5} - 0.12 \qquad \ldots(75)$$

where,

ET_g = evapotranspiration for grass, mm/day;

Δ = slope of the saturation vapour pressure - temperature curve, de/dT , mb/C;

γ = Psychrometric constant, mb/C;

R_S = solar radiation, cal/cm^2/day; ($R_n \cong 0.6\ R_S$)

R_n = net radiation, cal/cm^2/day.

(b) Turc

Turc (1961) simplified earlier versions of an equation for potential evapotranspiration for 10-day periods under general climatic conditions of Western Europe. When expressed on a daily basis in mm/day, the equations are:

For relative humidity > 50%

$$PET = 0.013 \frac{T}{T+15} (R_S+50) \qquad ...(76)$$

For relative humidity < 50%

$$PET = 0.013 \frac{T}{T+15} (R_S+50)(1+ \frac{50\text{-relative humidity}}{70}) \qquad ...(77)$$

where T is the average temperature in ^oC and R_S is in langleys per day (calories per cm^2 per day).

(c) Jensen and Haise

Jensen and Haise (1963) evaluated 3,000 observations of ET as determined by soil sampling procedures over a 35-year period. From about 100 values for well-watered crops with full cover in the Western USA, the constants for the following linear equation were C_T=0.014 and T_X=26.4 for temperature in ^oF, and 0.025 and -3 for temperature in ^oC.

$$PET = C_T(T-T_X)R_S \qquad ...(78)$$

where, C_T is a temperature coefficient and T_X is the intercept of the temperature axis. These coefficients are considered as constants for an area.

Obtaining reliable radiation values by either measurement or estimation often becomes the key to successful application of energy budget methods. Direct radiation measurements are usually one of two types:

(a) total incoming solar, R_S(e.g., by Epply pyrheliometer) or

(b) all wave net radiation, R_n(e.g., by Fritschen type net radiometer). Net radiation can be used directly to predict PET. Solar radiation, which is the most common measurement at meteorologic stations, can be used to estimate R_n taking into account the albedo and heating coefficients of the plant and soil surfaces. Albedo and emittance vary with stage of plant growth, soil colour, degree of wetness, crop and soil temperature, sun angle, and other factors. Davies and Buttimor (1969) pooled data for many crops and world locations and concluded that many surfaces are similar enough that mean values can be quite useful. The mean of their relationships is

$$R_n = 0.63\ R_S - 48 \qquad \text{...(79)}$$

Jensen (1973) also summarizes radiation measurements from 28 diverse locations and sites. Almost all results had a correlation coefficient greater than 0.90, many above 0.95. The means for these data provide the relationship

$$R_n = 0.65\ R_S - 45 \qquad \text{...(80)}$$

which is almost identical with equation (79). Saxton (1972) reported a relationship very similar to those of equations (79) and (80); however, he further showed a seasonal trend in the R_n/R_S ratio, which ranged from about 0.40 at mid-March and November to about 0.55 in mid-July.

The slope and aspect of a plane at the earth's surface may introduce considerable variation to the incident radiation as compared with a horizontal plane. A watershed is composed of a multitude of individual facets, but the average slope-aspect effect can be determined from the view that all incoming radiation must pass through a plane defined by points on the watershed boundary, thus a single plane can be used for some objectives. The slope and orientation will be particularly important for relatively small, steep watersheds.

(d) Hargreaves et al.

Hargreaves et al. (1985) presented an equation for PET that requires only maximum and minimum temperatures and extraterrestrial radiation (RA). Hargreaves and Samani (1985) compared the monthly PET values estimates by the

Hargreaves equation with those from the modified Penman method and with cool-season grass lysimeter evapotranspiration. They concluded that the temperature-based equation provided a better fit with the measured monthly values in most cases. The equation is

$$PET = 0.0023\ RA\ (T^{o}C+17.8)\ TD^{0.50} \qquad \ldots(81)$$

where,

RA = extraterrestrial radiation;

$T^{o}C$ = mean maximum temperature (T_{mx}) plus mean minimum temperature (T_{mi}) divided by two, $(T_{mx}+T_{mi})/2$;

and TD = $T_{mx}-T_{mi}$.

PET and RA are in the same units of equivalent depth of water evaporation, and values of T_{mx} and T_{mi} are in degrees Celsius.

4.4.4 Aerodynamic profile methods

The measurement of water vapour, as it is transported away from an evaporating surface, offers the potential of the most direct measurement of ET. The approach usually involves measuring temperature and vapour pressure of the air at two or more heights above the evaporating crop and a profile of wind velocities to define moisture and temperature gradients and fluctuations of wind velocity and humidity at a single height. The measurements are all quite sensitive and the amount of required data is voluminous. Considerable research has been conducted with sophisticated instrumentation, and good results have been obtained as compared with lysimeters, but instrumentation and techniques for hydrologic measurements or predictions are not yet developed to the point that these methods can be routinely applied.

4.4.5 Combination methods

Neither the vertical energy budget nor the aerodynamic methods are capable of predicting PET without assumptions and limitations. Penman (1948,1956) developed a method to combine these two theories which removed some of the limitations, and his equation is widely used. With refinement and testing it now represents one of the more reliable techniques for predicting PET from climatic data.

The complete derivation of the combination equation is quite lengthy and involves many micrometeorologic concepts. The derivation can be divided into the following steps:

(a) define the vertical energy budget of the soil or plant surface,

(b) apply the Dalton-type transport function to obtain Bowen's ratio,

(c) apply Penman's psychrometric simplification to eliminate the need for surface temperature, and

(d) apply the vertical transport equation obtained from turbulent transport theory.

(i) A complete development for each of these steps is given by Saxton (1972). Several assumptions are made in the course of the derivation, like air thermal stability and equal transport coefficients of momentum and vapour, but these seem to have negligible effects for most applications.

The combination equation may be written:

$$LE = \frac{(\frac{\Delta}{\gamma})R_n + (KLd_a u_a)/[\ln(Z_a - \frac{d}{Z_o})]^2}{1+(\frac{\Delta}{\gamma})} \qquad \ldots(82)$$

and

$$K = \frac{\rho k^2 \varepsilon}{P} \qquad \ldots(83)$$

where,

E = potential evapotranspiration rate (cm/day),

Δ = slope of psychrometric saturation line (mbars/^{0}C),

γ = psychrometric constant (mbars/^{0}C),

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

L = latent heat of vaporization (cal/g),

d_a = saturation vapour pressure deficit of air (mbars),

u_a = windspeed at elevation Z_a (m/day),

Z_a = anemometer height above soil (cm),

d = wind profile displacement height (cm),

Z_o = wind profile roughness height (cm),

ρ = air density (g/cm^3),

k = von Karman coefficient (0.41),

ε = water/air molecular ratio (0.622), and

p = ambient air pressure (mbars).

All terms of K and the value of L are treated as constants in most applications. Application of the combination equation requires measurements or estimates of four variables - net radiation, air temperature, air humidity, and horizontal wind movement - plus appropriate values for the other parameters which can usually be treated as constants for a given site. Net radiation can be assessed the same as for the energy budget approach. The (Δ/γ) term is a function of air temperature and tabled values are available (van Bavel, 1966).

(ii) The Penman-Monteith evapotranspiration equation

The Penman-Monteith equation is a one-dimensional single-source model of the evapotranspiration process. The assumptions on which it is based are described in the development by Monteith (1965). Within the limitations of these assumptions actual evapotranspiration (ET) is predicted by:

$$ET = \frac{\Delta_s A + \rho C_p (q_{w,Td} - q)/r_a}{\lambda[\Delta_s + (C_p/\lambda)(1 + r_c/r_a)]} \quad (Kgm^{-2}s^{-1} \approx mms^{-1}) \qquad \ldots(84)$$

where,

λ = latent heat of vaporization of water ($=2.47\times10^6 J\ Kg^{-1}$)

C_P = specific heat of air at constant pressure ($=1.01\times10^3 J\ Kg^{-1}{}^oC^{-1}$)

ρ = density of air ($=1.2 Kgm^{-2}$)

A = available energy given by $A = R_N - G$ (Wm^{-2})

R_N = net radiation measured at the reference height, Z (Wm^{-2})

G denotes the sum of energy fluxes into the ground, to adsorption by photosynthesis and respiration and to storage between ground level and Z (Wm^{-2})

$q_{w,Td}$ = saturated specific humidity at dry-bulb temperature, Td ($KgKg^{-1}$)

SHD = derived from measurements of Td and wet-bulb temperature depression, DEP

Δ_s = slope of the specific humidity/temperature curve between the air temperature Td and the surface temperature of the vegetation Ts ($KgKg^{-1}{}^oC^{-1}$)

r_a = aerodynamic resistance to the transport of water vapour from the surface to the reference level Z (sm^{-1})

r_c = (Monteith) canopy resistance to the transport of water from some region within or below the evaporating surface to the surface itself, and is expected to be a function of the stomatal resistance of individual leaves. Under wet-canopy conditions $r_c=0$ (sm^{-1})

$(q_{w,Td} - q)$ = specific humidity deficit (SHD) ($kg\ kg^{-1}$)

Equation (84) assumes that all evapotranspiration within the complex soil-vegetation canopy system takes place from a single representative source layer. The meteorological data required for the model are values of A(or R_N if G is small), T_d and DEP ($=T_d-T_w$) where T_w is wet-bulb temperature. The parametric data required are values of r_a and r_c.

4.4.6 Temperature methods

One of the earliest methods of estimating ET involved the use of air temperature. Studies initiated in the 1920's related ET to air temperature.

(a) Lowry and Johnson

Lowry and Johnson (1942) developed a procedure for estimating water requirements for irrigation projects which has been used by the United States Bureau of Reclamation. This method applies to a valley, not to an individual farm, and has been widely used with good results by the Bureau in the arid western portions of the United States. It is essentially an empirical procedure.

A linear relationship is assumed between 'effective heat' and consumptive use. Effective heat is defined as the accumulation, in day-degrees, of maximum daily growing season temperatures above 32^oF.

The approximate relationship

$$CU = 0.8 + 0.156\ I_F \qquad ...(85)$$

is used in estimating the valley consumptive use by the Lowry-Johnson method,

where CU = consumptive use in acre-feet per acre, and

I_F = effective heat in thousands of day-degrees.

(b) Thornthwaite

Thornthwaite (1948) correlated mean monthly air temperature with ET as determined by water balance studies in valley of east central USA. The resulting equation carried the following requirements(Thornthwaite and Mather, 1955)

(i) The albedo of the evaporating surface must be a standard.

(ii) The rate of evapotranspiration must not be influenced by advection of moist or dry air.

(iii) The ratio of energy utilized in evaporation to that of heating the air must remain essentially constant.

Since these conditions essentially do not exist in arid and semiarid areas, the equation could not be expected to provide reliable data in these areas.

If t_n=average monthly temperature of the consecutive months of the year in ^oC (where n=1,2,3,.....,12) and j=monthly 'heat index', then

$$j = \left(\frac{t_n}{5}\right)^{1.514} \qquad \ldots(86)$$

and the yearly 'heat index',J, is given by

$$J = \sum_{1}^{12} j \text{(for the 12 months)}$$

The potential evapotranspiration for any month with average temperature $t(^oC)$ is then given, as PET_x, by

$$PET_x = 16\left(\frac{10t}{J}\right)^a \text{ mm per month} \qquad \ldots(87)$$

where

$$a = (675\text{x}10^{-9})J^3-(771\text{x}10^{-7})J^2+(179\text{x}10^{-4})J+0.492 \qquad \ldots(88)$$

However, PET_x is a theoretical standard monthly value based on 30 days and 12 hours of sunshine per day. The actual PET for the particular month with average temperature $t(^{o}C)$ is given by

$$PET = PET_x \frac{DT}{360} \text{ mm} \qquad \ldots(89)$$

where,

D = number of days in the month, and

T = average number of hours between sunrise and sunset in the month.

The method has been tested by Serra, who suggested that equations (86) and (88) may be simplified as follows

$$j = 0.09\ t_n^{3/2} \qquad \ldots(90)$$

$$a = 0.016J+0.5 \qquad \ldots(91)$$

This method of estimating potential evapotranspiration is empirical and complicated. It has been found that the method gives reasonably good results whatever the vegetation cover, though different types of vegetation will affect a particular locality's true value. The formula is based on temperature, which does not necessarily correspond to incoming solar radiation immediately, because of the 'heat inertia' of land and water. Transpiration, however, responds directly to solar radiation. Accordingly, care should be exercised when using the method to ensure that conditions do not change abruptly in a particular month though if figures for many consecutive months are being used, the cumulative differences are probably negligible. For project studies the method is a useful complement to the Penman approach.

(c) Blaney and Criddle

The Blaney-Criddle procedure for estimating evapotranspiration is used extensively throughout the world. The basic procedures were based on measurements of evapotranspiration in the 1920's and 1930's, using primarily soil sampling techniques. Blaney and Morin (1942) first developed an empirical relationship between evapotranspiration and mean air temperature, average relative humidity, and mean percentage of day time hours. The relationship may be expressed as

$$PET = p(0.45T+8)(H-R^{m})/100 \qquad \ldots(92)$$

in which PET is potential evapotranspiration (mm/day) when T and R are daily values; p is ratio of maximum sunshine hours for period of interest to the annual maximum; T is temperature (^{o}C); R is relative humidity (%); H and m are constants.

This relationship was later modified by Blaney and Criddle (1945,1950, and 1962) and Blaney et al.(1952) excluding the humidity term. The method is suggested for areas where available climatic data cover air temperature data only. The original Blaney-Criddle equation (1950) involves the calculation of the consumptive use factor(f) from mean temperature (T) and percentage (p) of total annual daylight hours occurring during the period being considered. An empirically determined consumptive use crop coefficient (K) is then applied to establish the consumptive water requirements (CU) or

$$CU = K.f = K(\frac{p.T}{100}) \text{ with T in } ^{o}F \qquad \ldots(93)$$

CU is defined as'the amount of water potentially required to meet the evapotranspiration needs of vegetative areas so that plant production is not limited by lack of water'. Mean daily percentage (p) of annual daytime hours for different latitudes are given in table 4 (India lies between $8^{o}4'$ and $37^{o}6'$ North latitude). In the absence of a more reliable information based on local experimental data, the values of consumptive use crop coefficient(K) for common irrigated crops given in table 5 may be used in computations.

Table 4 - Mean Daily Percentage (p) of Annual **Daytime Hours** for Different Latitudes

Latitude North / South	Jan. / Jul.	Feb. / Aug.	Mar. / Sept.	Apr. / Oct.	May / Nov.	Jun. / Dec.	Jul. / Jan.	Aug. / Feb.	Sept. / Mar.	Oct. / Apr.	Nov. / May	Dec. / Jun.
60°	0.15	0.20	0.26	0.32	0.38	0.41	0.40	0.34	0.28	0.22	0.17	0.13
58	0.16	0.21	0.26	0.32	0.37	0.40	0.39	0.34	0.28	0.23	0.18	0.15
56	0.17	0.21	0.26	0.32	0.36	0.39	0.38	0.33	0.28	0.23	0.18	0.16
54	0.18	0.22	0.26	0.31	0.36	0.38	0.37	0.33	0.28	0.23	0.19	0.17
52	0.19	0.22	0.27	0.31	0.35	0.37	0.36	0.33	0.28	0.24	0.20	0.17
50	0.19	0.23	0.27	0.31	0.34	0.36	0.35	0.32	0.28	0.24	0.20	0.18
48	0.20	0.23	0.27	0.31	0.34	0.36	0.35	0.32	0.28	0.24	0.21	0.19
46	0.20	0.23	0.27	0.30	0.34	0.35	0.34	0.32	0.28	0.24	0.21	0.20
44	0.21	0.24	0.27	0.30	0.33	0.35	0.34	0.31	0.28	0.25	0.22	0.20
42	0.21	0.24	0.27	0.30	0.33	0.34	0.33	0.31	0.28	0.25	0.22	0.21
40	0.22	0.24	0.27	0.30	0.32	0.34	0.33	0.31	0.28	0.25	0.22	0.21
35	0.23	0.25	0.27	0.29	0.31	0.32	0.32	0.30	0.28	0.25	0.23	0.22
30	0.24	0.25	0.27	0.29	0.31	0.32	0.31	0.30	0.28	0.26	0.24	0.23
25	0.24	0.26	0.27	0.29	0.30	0.31	0.31	0.29	0.28	0.26	0.25	0.24
20	0.25	0.26	0.27	0.28	0.29	0.30	0.30	0.29	0.28	0.26	0.25	0.25
15	0.26	0.26	0.27	0.28	0.29	0.29	0.29	0.28	0.28	0.27	0.26	0.25
10	0.26	0.27	0.27	0.28	0.28	0.29	0.29	0.28	0.28	0.27	0.26	0.26
5	0.27	0.27	0.27	0.28	0.28	0.28	0.28	0.28	0.28	0.27	0.27	0.27
0	0.27	0.27	0.27	0.27	0.27	0.27	0.27	0.27	0.27	0.27	0.27	0.27

[Source:Doorenbos and Pruitt (1977)]

Table 5 - Monthly Consumptive Use Crop Coefficients (K) for Use in Blaney - Criddle Formula

Crop	Months											
	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sept.	Oct.	Nov.	Dec.
Rice				0.85	1.00	1.15	1.30	1.25	1.10	0.90		
Maize				0.50	0.60	0.70	0.80	0.80	0.60	0.50		
Wheat	0.50	0.70	0.75	0.70								
Sugar cane	0.75	0.80	0.85	0.85	0.90	0.95	1.00	1.00	0.95	0.90	0.85	0.75
Cotton				0.50	0.60	0.75	0.90	0.85	0.75	0.55	0.50	0.50
Vegetables	0.50	0.55	0.60	0.65	0.70	0.75	0.80	0.80	0.70	0.60	0.55	0.50
Berseem	0.50	0.70	0.80	0.90	1.00				0.60	0.65	0.70	0.60
Citrus	0.50	0.55	0.55	0.60	0.60	0.65	0.70	0.70	0.65	0.60	0.55	0.55

[Source : **Dastane (1972)**]

The effect of climate on crop water requirements is, however, insufficiently defined by temperature and day length ; crop water requirements will vary widely between climates having similar values of T and p. Consequently the consumptive use crop coefficient (K) will need to vary not only with the crop but also very much with climatic conditions. For a better definition of the effect of climate on crop water requirements, but still employing the Blaney-Criddle temperature and day length related f factor, Doorenbos and Pruitt (1977) presented a method to calculate crop evapotranspiration (ET_{crop}). Using measured temperature data as well as general levels of humidity, sunshine and wind, an improved prediction of the effect of climate on evapotranspiration should be obtainable. The crop coefficients are considered to be less dependent on climate. The relationship recommended, representing mean value over the given month, is expressed as:

$$ET_{crop} = K_c . C[p(0.46T+8)] \text{ mm/day} \qquad \ldots(94)$$

where,

ET_{crop} = crop evapotranspiration in mm/day for the month considered,

K_c = crop coefficient (table 6 to 14), as given by Doorenbos and Pruitt (1977),

C = adjustment factor which depends on minimum relative humidity, sunshine hours and day time wind estimates,

p = mean daily percentage of total annual day time hours obtained from table 4 for a given month and latitude, and

T = mean daily temperature in ^{o}C over the month considered.

Since the empiricism involved in any ET prediction method using a single weather factor is inevitably high, this method should only be used when temperature data are the only measured weather data available. It should be used with scepticism (i) in equatorial regions where temperatures remain fairly constant but other weather parameters will change;(ii) for small islands and coastal areas where air temperature is affected by the sea temperature having little response to seasonal change in radiation; (iii) at high altitudes due to the fairly low mean daily temperatures (cold nights)even though day time radiation levels are high; and (iv) in climates with a wide variability in sunshine hours during transition periods (e.g., monsoon climates, mid-latitude climates during spring and autumn). The Radiation Method is preferable under these conditions even when the sunshine or radiation data need to be obtained from regional or global maps in the absence of any actual measured data.

At high latitudes (55^{o} or more)the days are relatively long but radiation is lower as compared to low and medium latitude areas having the same day length values. This results in an undue weight being given to the day length related p factor. Calculated evapotranspiration values should be reduced by upto 15 percent for areas at latitudes of 55^{o} or more. Concerning altitude, in semi-arid and arid areas evapotranspiration values can be adjusted downwards some 10 percent for each 1000 m altitude change above sea level.

Calculation of mean daily evapotranspiration should be made for periods no shorter than one month. Since for a given location climatic conditions and consequently evapotranspiration may vary greatly from year to year, evapotranspiration should preferably be calculated for each calendar month for each year of record rather than by using mean temperatures based on several years' record.

For ease of reference, approximate ranges of seasonal ET_{crop} for different crops are given in table 15. The magnitudes shown will change according to climate, crop characteristics, length of growing season, time of planting etc.

Table 6 - Crop Coefficient (K_c) for Field and Vegetable Crops for Different Stages of Crop Growth and Prevailing Climatic Conditions

Crop	Humidity		RHmin >70%		RHmin <20%	
	Wind m/sec		0-5	5-8	0-5	5-8
	Crop stage*					
Artichokes (perennial - clean cultivated)	mid-season	3	.95	.95	1.0	1.05
	at harvest or maturity	4	.9	.9	.95	1.0
Barley		3	1.05	1.1	1.15	1.2
		4	.25	.25	.2	.2
Beans (green)		3	.95	.95	1.0	1.05
		4	.85	.85	.9	.9
Beans (dry)		3	1.05	1.1	1.15	1.2
Pulses		4	.3	.3	.25	.25
Beets (table)		3	1.0	1.0	1.05	1.1
		4	.9	.9	.95	1.0
Carrots		3	1.0	1.05	1.1	1.15
		4	.7	.75	.8	.85
Castorbeans		3	1.05	1.1	1.15	1.2
		4	.5	.5	.5	.5
Celery		3	1.0	1.05	1.1	1.15
		4	.9	.95	1.0	1.05
Corn (sweet)		3	1.05	1.1	1.15	1.2
(maize)		4	.95	1.0	1.05	1.1
Corn (grain)		3	1.05	1.1	1.15	1.2
(maize)		4	.55	.55	.6	.6
Cotton		3	1.05	1.15	1.2	1.25
		4	.65	.65	.65	.7
Crucifers (cabbage, cauliflower, broccoli, Brussels sprout)		3	.95	1.0	1.05	1.1
		4	.80	.85	.9	.95
Cucumber		3	.9	.9	.95	1.0
Fresh market		4	.7	.7	.75	.8
Machine harvest		4	.85	.85	.95	1.0
Egg plant		3	.95	1.0	1.05	1.1
(aubergine)		4	.8	.85	.85	.9
Flax		3	1.0	1.05	1.1	1.15
		4	.25	.25	.2	.2
Grain		3	1.05	1.1	1.15	1.2
		4	.3	.3	.25	.25
Lentil		3	1.05	1.1	1.15	1.2
		4	.3	.3	.25	.25

Crop	Humidity / Wind m/sec		RHmin >70% 0-5	RHmin >70% 5-8	RHmin <20% 0-5	RHmin <20% 5-8
Lettuce	Crop stage*	3	.95	.95	1.0	1.05
		4	.9	.9	.9	1.0
Melons		3	.95	.95	1.0	1.05
		4	.65	.65	.75	.75
Millet		3	1.0	1.05	1.1	1.15
		4	.3	.3	.25	.25
Oats	mid-season	3	1.05	1.1	1.15	1.2
	harvest/maturity	4	.25	.25	.2	.2
Onion (dry)		3	.95	.95	1.05	1.1
		4	.75	.75	.8	.85
(green)		3	.95	.95	1.0	1.05
		4	.95	.95	1.0	1.05
Peanuts (Groundnuts)		3	.95	1.0	1.05	1.1
		4	.55	.55	.6	.6
Peas		3	1.05	1.1	1.15	1.2
		4	.95	1.0	1.05	1.1
Peppers (fresh)		3	.95	1.0	1.05	1.1
		4	.8	.85	.85	.9
Potato		3	1.05	1.1	1.15	1.2
		4	.7	.7	.75	.75
Radishes		3	.8	.8	.85	.9
		4	.75	.75	.8	.85
Safflower		3	1.05	1.1	1.15	1.2
		4	.25	.25	.2	.2
Sorghum		3	1.0	1.05	1.1	1.15
		4	.5	.5	.55	.55
Soybeans		3	1.0	1.05	1.1	1.15
		4	.45	.45	.45	.45
Spinach		3	.95	.95	1.0	1.05
		4	.9	.9	.95	1.0
Squash		3	.9	.9	.95	1.0
		4	.7	.7	.75	.8
Sugarbeet		3	1.05	1.1	1.15	1.2
		4	.9	.95	1.0	1.0
	no irrigation last month	4	.6	.6	.6	.6
Sunflower		3	1.05	1.1	1.15	1.2
		4	.4	.4	.35	.35
Tomato		3	1.05	1.1	1.2	1.25
		4	.6	.6	.65	.65
Wheat		3	1.05	1.1	1.15	1.2
		4	.25	.25	.2	.2

NB: Many cool season crops cannot grow in dry, hot climates. Values of K_c are given for latter conditions since they may occur occasionally, and result in the need for higher K_c values, especially for tall rough crops.

* The crop growing season has been divided into four stages :

Stage 1 (initial stage):	germination and early growth when the soil surface is not or is hardly covered by the crop (ground cover < 10%)
Stage 2 (crop development stage):	from end of initial stage to attainment of effective full ground cover (ground cover $\simeq$ 70-80%)
Stage 3 (mid-season stage):	from attainment of effective full ground cover to time of start of maturing as indicated by discolouring of leaves (beans) or leaves falling off (cotton). For some crops this may extend to very near harvest (sugarbeets) unless irrigation is not applied at late season and reduction in crop evapotranspiration is induced to increase yield and/or quality (sugarcane, cotton, some grains); normally well past the flowering stage of annual crops
Stage 4 (late season stage):	from end of mid-season stage until full maturity or harvest.

Table 7 - K_c Values for Alfalfa, Clover, Grass-legumes and Pasture

		Alfalfa	Grass for hay	Clover, Grass-legumes	Pasture
Humid Light to moderate wind	K_c mean	0.85	0.8	1.0	0.95
	K_c peak	1.05	1.05	1.05	1.05
	K_c low 1/	0.5	0.6	0.55	0.55
Dry Light to moderate wind	K_c mean	0.95	0.9	1.05	1.0
	K_c peak	1.15	1.1	1.15	1.1
	K_c low 1/	0.4	0.55	0.55	0.5
Strong wind	K_c mean	1.05	1.0	1.1	1.05
	K_c peak	1.25	1.15	1.2	1.15
	K_c low 1/	0.3	0.5	0.55	0.5

K_c(mean) represents mean value between cutting, K_c(low) just after cutting, K_c(peak) just before harvesting.

1/ Under dry soil conditions; under wet conditions increase values by 30%.

Table 8 - K_c Values for Bananas

	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
Mediterranean climate												
First-year crop, based on March planting with crop height 3.5m by August:												
Humid, light to mod. wind	-	-	.65	.6	.55	.6	.7	.85	.95	1.0	1.0	1.0
Humid, strong wind	-	-	.65	.6	.55	.6	.75	.9	1.0	1.05	1.05	1.05
Dry, light to mod. wind	-	-	.5	.45	.5	.6	.75	.95	1.1	1.15	1.1	1.1
Dry, strong wind	-	-	.5	.45	.5	.65	.8	1.0	1.15	1.2	1.15	1.15
Second season with removal of original plants in Feb. and 80% ground cover by August:												
Humid, light to mod. wind	1.0	.8	.75	.7	.7	.75	.9	1.05	1.05	1.05	1.0	1.0
Humid, strong wind	1.05	.8	.75	.7	.7	.8	.95	1.1	1.1	1.1	1.05	1.05
Dry, light to mod. wind	1.1	.7	.75	.7	.75	.85	1.05	1.2	1.2	1.2	1.15	1.15
Dry, strong wind	1.15	.7	.75	.7	.75	.9	1.1	1.25	1.25	1.25	1.2	1.2

continued table 8........

Tropical climates

months following planting:	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
	.4	.4	.45	.5	.6	.7	.85	1.0	1.1	1.1	.9	.8	.8	.95	1.05

suckering | shooting | harvesting

Table 9 - K_c Values for Citrus (Grown in Predominantly Dry Areas with Light to Moderate Wind)

	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
Large mature trees providing ≃ 70% tree ground cover-Clean cultivated	.75	.75	.7	.7	.7	.65	.65	.65	.65	.7	.7	.7
No weed control	.9	.9	.85	.85	.85	.85	.85	.85	.85	.85	.85	.85
Trees providing ≃ 50% tree ground cover-Clean cultivated	.65	.65	.6	.6	.6	.55	.55	.55	.55	.55	.6	.6
No weed control	.9	.9	.85	.85	.85	.85	.85	.85	.85	.85	.85	.85
Trees providing ≃ 20% tree ground cover-Clean cultivated	.55	.55	.5	.5	.5	.45	.45	.45	.45	.45	.5	.5
No weed control	1.0	1.0	.95	.95	.95	.95	.95	.95	.95	.95	.95	.95

Table 10 - K_c Values for Full Grown Deciduous Fruit and Nut Trees

	With ground-cover crop [1/]									Without ground cover crop [2/] (clean cultivated, weed free)								
	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov
COLD WINTER WITH KILLING FROST : GROUND COVER STARTING IN APRIL																		
Apples, cherries																		
humid, light to mod. wind	-	.5	.75	1.0	1.1	1.1	1.1	.85	-	-	.45	.55	.75	.85	.85	.8	.6	-
humid, strong wind	-	.5	.75	1.1	1.2	1.2	1.15	.9	-	-	.45	.55	.8	.9	.9	.85	.65	-
dry, light to mod. wind	-	.45	.85	1.15	1.25	1.25	1.2	.95	-	-	.4	.6	.85	1.0	1.0	.95	.7	-
dry, strong wind	-	.45	.85	1.2	1.35	1.35	1.25	1.0	-	-	.4	.65	.9	1.05	1.05	1.0	.75	-
Peaches, apricots, pears, plums																		
humid, light to mod. wind	-	.5	.7	.9	1.0	1.0	.95	.75	-	-	.45	.5	.65	.75	.75	.7	.55	-
humid, strong wind	-	.5	.7	1.0	1.05	1.1	1.0	.8	-	-	.45	.55	.7	.8	.8	.75	.6	-
dry, light to mod. wind	-	.45	.8	1.05	1.15	1.15	1.1	.85	-	-	.4	.55	.75	.9	.9	.7	.65	-
dry, strong wind	-	.45	.8	1.1	1.2	1.2	1.15	.9	-	-	.4	.6	.8	.95	.95	.9	.65	-
COLD WINTER WITH LIGHT FROST : NO DORMANCY IN GRASS COVER CROPS																		
Apples, cherries, walnuts [3/]																		
humid, light to mod. wind	.8	.9	1.0	1.1	1.1	1.1	1.05	.85	.8	.6	.7	.8	.85	.85	.8	.8	.75	.65
humid, strong wind	.8	.95	1.1	1.15	1.2	1.2	1.15	.9	.8	.6	.75	.85	.9	.9	.85	.8	.8	.7
dry, light to mod. wind	.85	1.0	1.15	1.25	1.25	1.25	1.2	.95	.85	.5	.75	.95	1.0	1.0	.95	.9	.85	.7
dry, strong wind	.85	1.05	1.2	1.35	1.35	1.35	1.25	1.0	.85	.5	.8	1.0	1.05	1.05	1.0	.95	.9	.75

continued table 10.....

	With ground cover crop[1/]									Without ground cover crop[2/] (clean cultivated, weed free)								
	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov
	COLD WINTER WITH LIGHT FROST:NO DORMANCY IN GRASS COVER CROPS																	
Peaches, apricots, pears, plums, almonds, pecans																		
humid, light to mod. wind	.8	.85	.9	1.0	1.0	1.0	.95	.8	.8	.55	7	.75	.8	.8	.7	.7	.65	.55
humid, strong wind	.8	.9	.95	1.0	1.1	1.1	1.0	.85	.8	.55	.7	.75	.8	.8	.8	.75	.7	.6
dry, light to mod. wind	.85	.95	1.05	1.15	1.15	1.15	1.1	.9	.85	.5	.7	.85	.9	.9	.9	.8	.75	.65
dry, strong wind	.85	1.0	1.1	1.2	1.2	1.2	1.15	.95	.85	.5	.75	.9	.95	.95	.95	.85	.8	.7

1/ K_c values need to be increased if frequent rain occurs. For young orchards with tree ground cover of 20 and 50%, reduce mid-season K_c values by 10 to 15% and 5 to 10% respectively.

2/ K_c values assume infrequent wetting by irrigation or rain (every 2 to 4 weeks). In the case of frequent irrigation for May to October use K_c values of table "with ground cover crop". For young orchards with tree ground cover of 20 and 50% reduce mid-season K_c values by 10 to 15%.

3/ For walnuts March-May possibly 10 to 20% lower values due to slower leaf growth.

Table 11 - K_c Values for Grapes (Clean Cultivated, Infrequent Irrigation, Soil Surface Dry Most of the Time)

	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
Mature grapes grown in areas with killing frost; initial leaves early May, harvest mid-September; ground cover 40-50% at mid-season												
humid, light to mod. wind	-	-	-	-	.5	.65	.75	.8	.75	.65	-	-
humid, strong wind	-	-	-	-	.5	.7	.8	.85	.8	.7	-	-
dry, light to mod. wind	-	-	-	-	.45	.7	.85	.9	.85	.7	-	-
dry, strong wind	-	-	-	-	.5	.75	.9	.95	.9	.75	-	-
Mature grapes in areas of only light frosts; initial leaves early April, harvest late August to early September; ground cover 30-35% at mid-season												
humid, light to mod. wind	-	-	-	.5	.55	.6	.6	.6	.6	.5	.4	-
humid, strong wind	-	-	-	.5	.55	.65	.65	.65	.65	.55	.4	-
dry, light to mod. wind	-	-	-	.45	.6	.7	.7	.7	.7	.6	.35	-
dry, strong wind	-	-	-	.45	.65	.75	.75	.75	.75	.65	.35	-
Mature grapes grown in hot dry areas; initial leaves late February-early March, harvest late half of July; ground cover 30-35% at mid-season												
dry, light to mod. wind	-	-	.25	.45	.6	.7	.7	.65	.55	.45	.35	-
dry, strong wind	-	-	.25	.45	.65	.75	.75	.7	.55	.45	.35	-

Table 12 - K_c Values for Rice

	Planting	Harvest	First & Second month	Mid-season	Last 4 weeks
<u>Humid Asia</u>					
wet season (monsoon)	June-July	Nov-Dec			
light to mod. wind			1.1	1.05	.95
strong wind			1.15	1.1	1.0
dry season 1/	Dec-Jan	mid-May			
light to mod. wind			1.1	1.25	1.0
strong wind			1.15	1.35	1.05
<u>North Australia</u>					
wet season	Dec-Jan	Apr-May			
light to mod. wind			1.1	1.05	.95
strong wind			1.15	1.1	1.0
<u>South Australia</u>					
dry summer	Oct	March			
light to mod. wind			1.1	1.25	1.0
strong wind			1.15	1.35	1.05
<u>Humid S.America</u>					
wet season	Nov-Dec	Apr-May			
light to mod. wind			1.1	1.05	.95
strong wind			1.15	1.1	1.0
<u>Europe (Spain, S. France and Italy)</u>					
dry season	May-June	Sept-Oct			
light to mod. wind			1.1	1.2	.95
strong wind			1.15	1.3	1.0
<u>U.S.A.</u>					
wet summer (South)	May	Sept-Oct			
light to mod. wind			1.1	1.1	.95
strong wind			1.15	1.15	1.0
dry summer (Calif.)	early May	early Oct			
light to mod. wind			1.1	1.25	1.0
strong wind			1.15	1.35	1.05

1/ Only when RHmin > 70%, K_c values for wet season are to be used.

Table 13 - K_c Values for Sugarcane

Crop age		Growth stages	RHmin > 70%		RHmin < 20%	
12 month	24 month		light to mod.wind	strong wind	light to mod.wind	strong wind
0-1	0-2.5	planting to 0.25 full canopy	.55	.6	.4	.45
1-2	2.5-3.5	0.25-0.5 full canopy	.8	.85	.75	.8
2-2.5	3.5-4.5	0.5-0.75 full canopy	.9	.95	.95	1.0
2.5-4	4.5-6	0.75 to full canopy	1.0	1.1	1.1	1.2
4-10	6-17	peak use	1.05	1.15	1.25	1.3
10-11	17-22	early senescence	.8	.85	.95	1.05
11-12	22-24	ripening	.6	.65	.7	.75

Table 14 - K_c Values for Aquatic Weeds

Type of vegetation	Humid		Dry	
	light to mod. wind	strong wind	light to mod. wind	strong wind
Submerged (crassipes)	1.1	1.15	1.15	1.2
Floating (duckweed)	1.05	1.05	1.05	1.05
Flat leaf (water lilies)	1.05	1.1	1.05	1.1
Protruding (water hyacinth)	1.1	1.15	1.15	1.2
Reed swamp (papyrus, cattails)				
standing water	.85	.85	.9	.95
moist soil	.65	.65	.75	.8

Table 15 - Approximate Range of Seasonal Crop Evapotranspiration

S.No.	Crop	Seasonal ET_{crop} (mm)
1.	Alfalfa	600-1500
2.	Avocado	650-1000
3.	Bananas	700-1700
4.	Beans	250-500
5.	Cocoa	800-1200
6.	Coffee	800-1200
7.	Cotton	550-950
8.	Dates	900-1300
9.	Deciduous trees	700-1050
10.	Flax	450-900
11.	Grains (small)	300-450
12.	Grape fruit	650-1000
13.	Maize	400-750
14.	Oil seeds	300-600
15.	Onions	350-600
16.	Orange	600-950
17.	Potatoes	350-625
18.	Rice	500-950
19.	Sisal	550-800
20.	Sorghum	300-650
21.	Soybeans	450-825
22.	Sugarbeets	450-850
23.	Sugarcane	1000-1500
24.	Sweet potatoes	400-675
25.	Tobacco	300-500
26.	Tomatoes	300-600
27.	Vegetables	250-500
28.	Vineyards	450-900
29.	Walnuts	700-1000

[Source: Doorenbos and Pruitt (1977)]

(a) Turc

Turc, in Europe, developed formulae, based on a statistical study of data collected from 254 watersheds located in all parts of the world, to relate evaporation, rainfall and temperature over watersheds. He suggests that annual evaporation or evapotranspiration may be estimated as

$$E = \frac{P}{[0.90+(\frac{P}{I_T})^2]^{\frac{1}{2}}} \qquad \ldots(95)$$

in which E = annual evaporation (mm),

P = annual precipitation (mm),

$I_T = 300+25T+0.05T^3$, and

T = mean air temperature (^oC).

Turc also suggests a more complex formula to give evaporation over short periods of time. In this formula he attempts to take into account the effect on evapotranspiration of different levels of soil moisture by different crops:

$$E = \frac{P+E_{10}+K}{[1+(\frac{P+E}{I_T} + \frac{K}{2I_T})^2]^{\frac{1}{2}}} \qquad \ldots(96)$$

where,

E = evaporation in 10 day period (mm),

P = precipitation in 10 day period (mm),

E_{10} = estimated evaporation (in 10 day period) from bare soil assuming no precipitation, and is not greater than 10 mm,

K = a crop factor expressible in two ways, one of which is

$$K = 25(MC/G)^{\frac{1}{2}} \qquad \ldots(97)$$

where, 100 M = the final yield of dry matter (kg/ha),

10 G = the length of the growing season (days), and

C = a crop constant,

I_T = the "evaporation capacity" of the air, calculated from

$$I_T = \frac{(T+2)R_{si}^{\frac{1}{2}}}{16} \qquad \ldots(98)$$

where, T = mean air temperature over the 10 day period (^{o}C), and

$\dot{R}_{si}$ = incoming radiant energy (cal/cm^2/day).

Values of C, given by Turc, are

Maize and beet	0.67
Potatoes	0.83
Cereals, flax, carrots	1.00
Peas, clover, legumes except lucerne	1.17
Lucerne, meadow grasses, and mustard	1.33

When soil moisture is not limiting, in summer weather when $I_T > 10$, Turc reduces the equation to

$$E = \frac{P+E_{10}+70}{[1+(\frac{P+E}{I_T} + \frac{70}{2I_T})^2]^{\frac{1}{2}}} \text{ mm /10 days} \qquad \ldots(99)$$

(e) Hamon

Methods of computing potential evapotranspiration by analytical procedures have been based on the application of the turbulent-transport and energy-balance concepts. Empirical formulas, correlating some temperature function and adjusting for day-time hours, have proved valuable in practical utilization. Hamon (1961) used this latter approach to formulate a simple computational procedure whereby average daily potential evapotranspiration is represented as proportional to the product of day-time hours squared and the saturated water vapour concentration (absolute humidity) at the mean temperature. The day-time factor was determined from a consideration of the disparity between net radiation and temperature, latitudinally, and the fact that transpiration is restricted during darkness since the leaf stomata are closed. Computed values of potential evapotranspiration obtained by the new procedure are compared with those obtained by the more complex Thornthwaite method and other methods. General applicability was found justified from comparisons between observed and computed values of potential evapotranspiration, both on a yearly and seasonal basis. The final simplified expression for potential evapotranspiration was formulated as

$$PET = CD^2P_t \qquad ...(100)$$

in which PET represents the average potential evapotranspiration in inches per day, D is the possible hours of sunshine in units of 12 hour, P_t is the saturated water vapour density (absolute humidity) at the daily mean temperature in grams per cubic meter, times 10^{-2}, C=0.55, chosen to give appropriate yearly values of potential evapotranspiration.

4.4.7 Humidity methods

Saturation deficit is one of the oldest parameters used to estimate potential ET.David (1936) has proposed that[vide Vitkevich (1958)]

$$PET = (\bar{e}_Z^0 - e_Z)/2 \qquad ...(101)$$

where PET is in mm/day and $\bar{e}_Z^0$ is the saturation vapour pressure at mean air temperature.

Halstead (1951) proposed

$$PET = Cd_L(q_{max} - q_{min}) \qquad ...(102)$$

where, q_{max} and q_{min} is saturation absolute humidity corresponding to maximum and minimum air temperature, d_L is the fraction of annual day-light hours, and C=1 when PET is in mm/month.

Ostromecki (1965) described a common general equation that has been used in east Europe which apparently was developed by Alpat'ev (1954),

$$ET = \beta_H \, d_a \qquad \ldots(103)$$

where, ET is in mm/day, d_a is average daily vapour pressure deficit in mb, and β_H is a "hygrometric" coefficient (β_H=0.56 for clover).

Pepadakis (1966) proposed

$$PET = 0.5625 \; (e^o_{max} - e_Z) \qquad \ldots(104)$$

where, PET is monthly potential ET in cm, e^o_{max} is saturation vapour pressure corresponding to average daily maximum temperature, mb, and e_Z is the average vapour pressure for the month. When e_Z is not known, then e_{min-2} is used where it represents the saturation vapour pressure at 2^oC below minimum temperature.

4.4.8 Multiple correlation methods

Christiansen (1968) and Christiansen and Hargreaves (1969) developed an equation for estimating USWB Class A pan evaporation from which potential ET can be estimated. The equations for pan evaporation and coefficients are reproduced below using Christiansen's notation.

$$PET = 0.755 \; E_V C_{T2} C_{W2} C_{H2} C_{S2} \qquad \ldots(105)$$

where, E_V is measured Class A pan evaporation.

$$C_{T2} = 0.670 + 0.476\left(\frac{T}{T_0}\right) - 0.146\left(\frac{T}{T_0}\right)^2 \qquad \ldots(106)$$

where, T is the mean temperature, ^oF, and $T_0 = 68^oF$.

In metric units,

$$C_{T2} = 0.862 + 0.179\left(\frac{T_C}{T_{CO}}\right) - 0.041\left(\frac{T_C}{T_{CO}}\right)^2 \qquad \ldots(107)$$

where, T_C is the mean temperature, ^oC, and $T_{C0} = 20^oC$.

$$C_{W2} = 1.189 - 0.240(\frac{w}{w_0}) + 0.051(\frac{w}{w_0})^2 \qquad \ldots(108)$$

where, w is the mean wind velocity 2 metres above ground level in miles per day or Km per hour, and w_0=100 miles per day or 6.7 Km per hour.

$$C_{H2} = 0.499 + 0.620(\frac{H_m}{H_{m0}}) - 0.119(\frac{H_m}{H_{m0}})^2 \qquad \ldots(109)$$

where, H_m is the mean relative humidity, expressed decimally, and H_{m0}=0.60.

$$C_{S2} = 0.904 + 0.0080(\frac{S}{S_0}) + 0.088(\frac{S}{S_0})^2 \qquad \ldots(110)$$

where, S is the percentage of possible sunshine, expressed decimally and S_0=0.80.

Using measured incoming radiation, R_S, as a base, the formula relating potential evapotranspiration, PET to measured incoming radiation can be written

$$PET = 0.492\ R_S C_{TT} C_{WT} C_{HT} \qquad \ldots(111)$$

where, R_S is the measured incoming solar radiation expressed as equivalent depth of evaporation.

$$C_{TT} = 0.174 + 0.428(\frac{T}{T_0}) + 0.398(\frac{T}{T_0})^2 \qquad \ldots(112)$$

where, T is the mean temperature, ^oF and T_0=68^oF.

In metric units this becomes

$$C_{TT} = 0.463 + 0.425(\frac{T_C}{T_{C0}}) + 0.112(\frac{T_C}{T_{C0}})^2 \qquad \ldots(113)$$

where, T_C is the mean temperature, ^oC, and T_{C0}=20^oC.

$$C_{WT} = 0.672+0.406(\frac{w}{w_0})-0.0780(\frac{w}{w_0})^2 \qquad ...(114)$$

where, w is the mean wind velocity 2 metres above ground level, and w_0=100 miles per day or 6.7 Km per hour.

$$C_{HT} = 1.035+0.240(\frac{H_m}{H_{m0}})^2-0.275(\frac{H_m}{H_{m0}})^3 \qquad ...(115)$$

where, H_m is the mean relative humidity expressed decimally and H_{m0}=0.60.

The measured incoming solar radiation is generally reported as langleys per day (calories per cm^2 per day). The equivalent depth of evaporation per day is obtained by dividing langleys per day, R_{ly}, by the latent heat of vaporization, L.

$$R_S = \frac{R_{ly}}{L} \qquad ...(116)$$

At a temperature of 20^oC (68^oF), L has a value of 584.9 calories per gram. For other temperatures in ^oC, the relation is

$$L = 595.9 - 0.55T_C \qquad ...(117)$$

and for temperatures in ^oF,

$$L = 595.9 - 0.305(T-32) \qquad ...(118)$$

To obtain R_S for the month, one must multiply the mean daily value of R_S by the number of days in the month, or use the total langleys per month in equation (111). Because of the small variation of L with temperature, the value for 20^oC is generally used.

4.4.9 The complementary relationship

The concept of a complementary relationship between areal and potential evapotranspiration is based on the interaction between evaporating surfaces and overpassing air. By incorporating this crucial feedback mechanism, the relationship avoids the complexities of the soil -

plant system so that areal evapotranspiration can be estimated from its effects on the routinely observed temperatures and humidities used in computing potential evapotranspiration with no need for locally optimized coefficients. The complementary relationship can be represented by:

$$ET + PET = 2\ WET \qquad \ldots(119$$

or by : $ET = 2\ WET - PET$

in which ET is the areal evapotranspiration, the actual evapotranspiratio from an area so large that the effects of upwind boundary transitions are negligible; PET is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy-balance equations, representin the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and WET is the wet environmen areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

Figure 14 provides a schematic representation of equation (119) under conditions of constant radiant-energy supply. The ordinate represents evapotranspiration and the abscissa represents the water supply to the soil-plant surfaces of the area, a quantity that is usually unknow When there is no water available for areal evapotranspiration (extreme left of figure 14) it follows that ET = 0, that the air is very hot and dry and that PET is at its maximum rate of 2WET (the dry environment potential evapotranspiration). As the water supply to the soil-plant surfaces of the area increases (moving to the right in figure 14) the resultar equivalent increase in ET causes the overpassing air to become cooler and more humid which in turn produces an equivalent decrease in PET. Finally, when the supply of water to the soil-plant surfaces of the area has increased sufficiently, the values of ET and PET converge to that of WET.

The conventional definition for potential evapotranspiration is the same as the definition for the wet environment areal evapotranspiration. However, the potential evapotranspiration that is estimated from a solution of the vapour transfer and energy-balance equations by analytical, graphical or iterative techniques has reactions to changes in the water supply to the soil-plant surfaces similar to those shown for E_{TP} in figure 14 so that what is being estimated can exceed what is being defined by as much as 100%. By taking into account such reactions the complementary relationship is analogous to the Bernouilli equation for open-channel flow in which the potential energy responds in a complementary way to changes in kinetic energy.

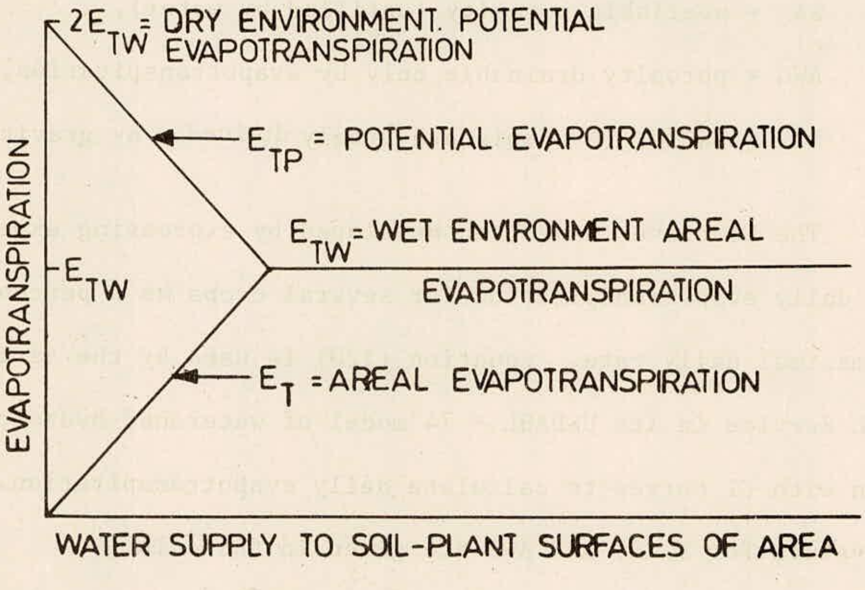


FIGURE 14 - SCHEMATIC REPRESENTATION OF COMPLEMENTARY RELATIONSHIP BETWEEN AREAL AND POTENTIAL EVAPOTRANSPIRATION WITH CONSTANT RADIANT - ENERGY SUPPLY

4.4.10 Agricultural Research Service

An equation for estimating potential evapotranspiration developed by the Agricultural Research Service (ARS) illustrates efforts to include vegetal characteristics and soil moisture in such a calculation. The evapotranspiration potential for any given day is determined as follows :

$$PET = GI.k.E_p.\left(\frac{S-SA}{AWC}\right)^x \qquad \ldots(120)$$

where,

PET = evapotranspiration potential (in./day),

GI = growth index of crop in % of maturity,

k = ratio of GI to pan evaporation, usually 1.0 - 1.2 for short grasses, 1.2 - 1.6 for crops up to shoulder height, and 1.6 - 2.0 for forest,

E_p = pan evaporation (in./day),

S = total porosity,

SA = available porosity (unfilled by water),

AWC = porosity drainable only by evapotranspiration, and

x = AWC/G (G = moisture freely drained by gravity).

The GI curves have been developed by expressing experimental data on daily evapotranspiration for several crops as a percentage of the annual maximal daily rate. Equation (120) is used by the Agricultural Research Service in its USDAHL - 74 model of watershed hydrology in combination with GI curves to calculate daily evapotranspiration. Representative values for S, G, and AWC are given in table 16.

4.4.11 Evapotranspiration from shallow water table areas

Evapotranspiration is the combined effect of the evaporation of water from moist soil and the transpiration of water by natural vegetation

Table 16 - Hydrologic Capacities of Soil Texture Classes

Texture Class	S (%)	G (%)	AWC (%)	x (AWC/G)
Coarse sand	24.4	17.7	6.7	0.38
Coarse sandy loam	24.5	15.8	8.7	0.55
Sand	32.3	19.0	13.3	0.70
Loamy sand	37.0	26.9	10.1	0.38
Loamy fine sand	32.6	27.2	5.4	0.20
Sandv loam	30.9	18.6	12.3	0.66
Fine sandy loam	36.6	23.5	13.1	0.56
Very fine sandy loam	32.7	21.0	11.7	0.56
Loam	30.0	14.4	15.6	1.08
Silt loam	31.3	11.4	19.9	1.74
Sandy clay loam	25.3	13.4	11.9	0.89
Clay loam	25.7	13.0	12.7	0.98
Silty clay loam	23.3	8.4	14.9	1.77
Sandy clay	19.4	11.6	7.8	0.67
Silty clay	21.4	9.1	12.3	1.34
Clay	18.8	7.3	11.5	1.58

S = total porosity - 15 bar moisture %,

G = total porosity - 0.3 bar moisture %, and

AWC = S minus G.

[Source : Holtan et al.(1975)]

and cultivated crops. Many factors play a role in the evapotranspiration from a groundwater basin: climate, soils, soil water availability, soil fertility, crops, cropping pattern and intensity, environment and exposure, cultivation practices and irrigation methods. One must therefore compile a land use map indicating the areas covered with natural vegetation and cultivated crops, waste areas, bare soils, urban areas, surface water bodies, etc. Since the water consumption of various crops differ, a crop survey must be made and a map compiled of the cropping pattern. A soils map and depth-to-watertable maps for the growing season should also be made. Existing climatological data should be collected and evaluated.

In shallow watertable areas, the groundwater contributes to evapotranspiration through capillary rise. This discharge is determined by the depth of the groundwater below the root zone, the capillary and conductive properties of the soil, and the soil water content or soil water tension in the root zone. At certain depths below the root zone, depending on the type of soil and provided that impermeable layers do not occur, the groundwater will contribute less than 1 mm/day to the root zone. These depths may be taken as approximately 50 to 90 cm for coarse and heavy textured soils and about 120 to 200 cm for most medium textured soils. Figure 15 shows the upward rate of groundwater flow in mm/day for different depths of groundwater below the root zone and for different soil types, under the assumption that the soil in the root zone is relatively moist, i.e. soil water tension equals about 0.5 atm.

4.4.12 Study of ground water fluctuations

The calculation of evapotranspiration from water table fluctuations represents one of the oldest approaches to this problem. In situations where vegetation obtains most of its moisture for growth from the capillary fringe above a water table, the amount of use is reflected in changes in the position of the water table. For shallow water table conditions, White (1932) suggests that consumptive use may be estimated by the expression :

$$CU = S_y(24a + b) \qquad \ldots(121)$$

in which,

CU = consumptive use (in./day),

S_y = specific yield of the soil.

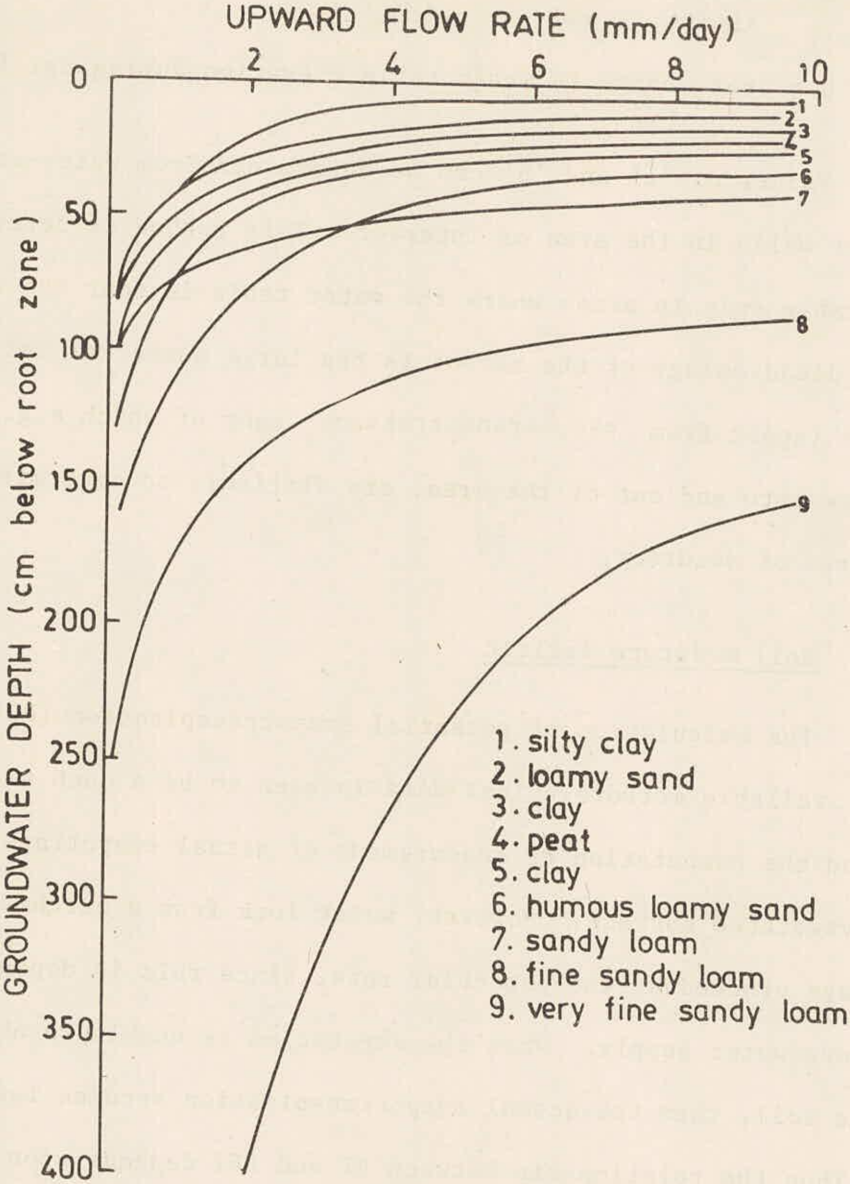


FIGURE 15 - CONTRIBUTION OF GROUNDWATER TO THE ROOT ZONE IN mm/d FOR DIFFERENT DEPTHS OF THE WATERTABLE BELOW THE ROOT ZONE AND FOR DIFFERENT SOIL TYPES UNDER MOIST CONDITIONS

(soil water tension of root zone approximately 0.5 atm.)

[Source : Doorenbos and Pruitt (1977)]

a = rate of rise of the water table from midnight to 4 A.M. (in./hr.), and

b = net change in water table elevation during day (in.).

Values of 'a' and 'b' can be determined from water-stage recorder charts for wells in the area of interest. This method of determining CU is applicable only in areas where the water table is near the surface. The main disadvantage of the method is the large number of influencing variables (apart from evapotranspiration), many of which e.g. ground water flow into and out of the area, are difficult to determine with a high degree of accuracy.

4.4.13 Soil moisture deficit

The calculation of potential evapotranspiration (PET) from readily available meteorological data is seen to be a much simple operation than the computation or measurement of actual evapotranspiration (ET) from a vegetated surface. However, water loss from a catchment area does not always proceed at the potential rate, since this is dependent on a continuous water supply. When the vegetation is unable to abstract water from the soil, then the actual evapotranspiration becomes less than potential. Thus the relationship between ET and PET depends upon the soil moisture content.

When the soil is saturated, it will hold no more water. After rainfall ceases, saturated soil relinquishes water and becomes unsaturated until it can just hold a certain amount against the forces of gravity; it is then said to be at 'field capacity'. In this range of conditions, ET = PET; evapotranspiration occurs at the maximum possible rate determined by the meteorological conditions. If there is no rain to replenish the water supply, the soil moisture gradually becomes depleted by the demands of the vegetation to produce a soil moisture deficit (SMD), viz. the amount of water required to restore the soil to field capacity. As SMD increases, ET becomes increasingly less than PET. The values of SMD and ET vary with soil type and vegetation, and the relative changes in ET with increasing SMD have been the subject of considerable study by botanists and soil physicists. Penman (1950) introduced the concept of a 'root constant' (RC) that defines the amount of soil moisture (mm depth) that can be extracted from a soil without difficulty by a given vegetation. Table 17 lists some typical root constants.

Table 17 - Root Constants in Soil Moisture Depths

S.No.	Vegetation Type	Root Constant (mm)
1.	Permanent grassland	75
2.	Root crops, e.g. potatoes	100
3.	Cereals, e.g. wheat, oats	140
4.	Woodland	200

It is assumed that ET = PET for a particular type of vegetation until the SMD reaches the appropriate root constant plus a further 25 mm approximately, which is added to allow for extraction from the soil immediately below the root zone. Thereafter, ET becomes less than PET as moisture is extracted with greater difficulty. As the SMD increases further, the vegetation wilts and ET becomes very small or negligible. Before the onset of wilting, vegetation will recover if the soil moisture is replenished, but there is a maximum SMD for each plant type known as 'permanent wilting point' from which the vegetation can not recover and dies.

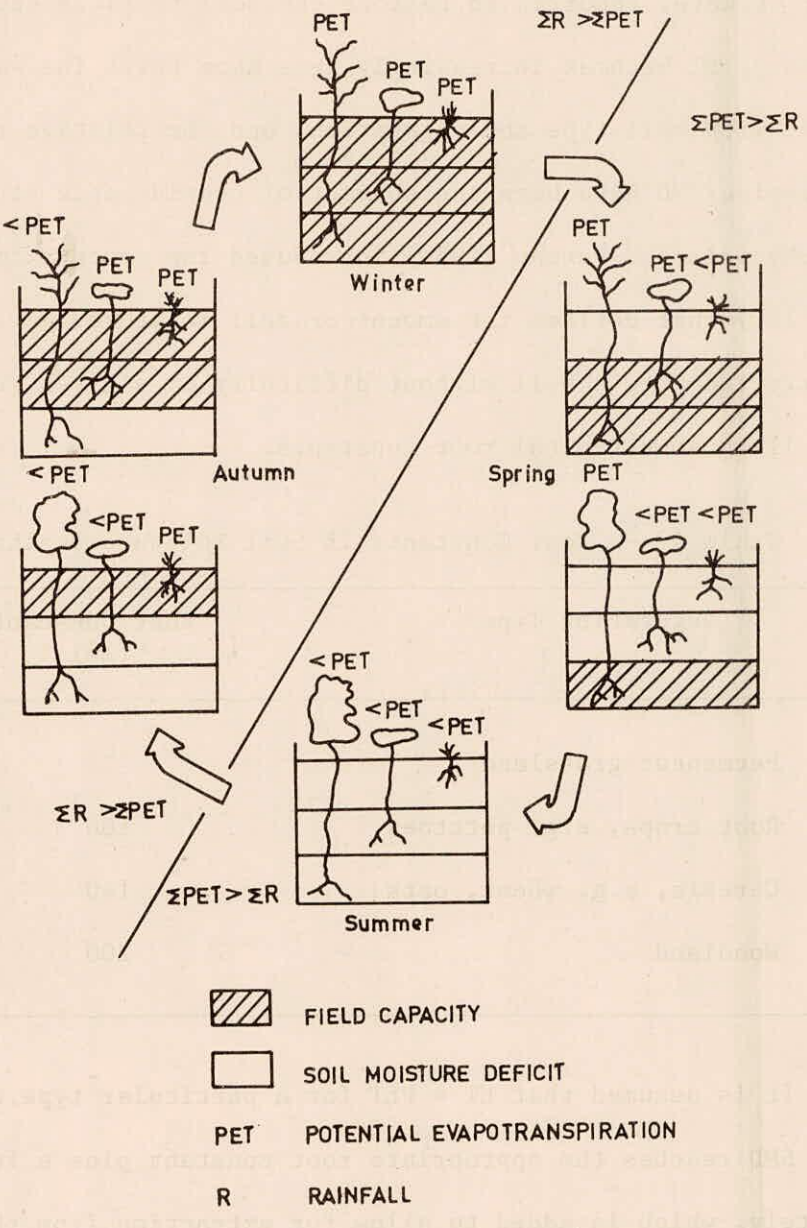


FIGURE 16 - IDEALIZED ANNUAL SOIL MOISTURE CYCLE FOR THREE VEGETATIONAL TYPES

Figure 16 shows diagrammatically the average sequences of soil moisture content related to the rainfall and potential evapotranspiration through the annual seasonal cycle. The soil moisture state is given at each stage for three types of vegetation: short-rooted grass, medium-rooted shrubs and deep-rooted trees. The actual evapotranspiration (ET) is given in terms of PET for each vegetational type at each stage. During the spring, when potential evapotranspiration exceeds rainfall, the soil moisture deficit begins first in the surface soil layers and then moves downwards into the lower layers as the water in the soil is used up, until in the summer months there could be soil moisture deficits in all the rooting zones of the soil. When rainfall totals begin to exceed potential evapotranspiration in the autumn, the soil moisture stores are gradually replenished from the top soil layer downwards, until all the soil layers reach field capacity again in the winter.

A soil moisture budget can be made on a monthly basis for various types of vegetation classified according to their root constants. To evaluate soil moisture deficit and actual evapotranspiration over a catchment area, the proportions of the different types of vegetation covering the catchment must be known. This entails a land-use survey and a classification of the vegetation for allocation of RCs before water budgeting may be carried out.

4.5 Choice of Method for Estimating Evapotranspiration

The method selected for a particular use may depend on the accuracy of available meteorological data, the training and experience of the user, and the general acceptance of previous estimates. No single existing method using meteorological data is universally adequate under all climatic regimes, especially for tropical areas and for high elevations, without some local or regional calibration. Availability of meteorological data alone should not be the sole criterion in selecting a method since some of the needed data can be estimated with sufficient accuracy by trained and experienced engineers and scientists to permit using one of the better estimating methods.

Solar radiation is the major driving force behind the evapotranspiration process, and any reasonably accurate method for computing evapotranspiration should account for it. It is essential to realize that methods that avoid the measurement of radiation can not be expected to produce consistent results. Two other factors - the soil heat flux and the controlling effects of the evaporating and transpiring surface - also need to be estimated for accurate results. The absence or presence of these factors in the estimation process leads to many of the differences in accuracy of the methods described.

The accurate methods for the estimation of evapotranspiration are lysimetry, measurement of the energy budget, and physically based combination approaches. Even for short time intervals, such as an hour, these approaches can estimate evapotranspiration with less than a 10% error, and cumulative error is reduced with increase in time. In the absence of these most accurate techniques, detailed soil-moisture measurement can provide reliable estimates and can be used for either large or small study areas. On the opposite side of the spectrum of estimation techniques are the simpler empirical approaches.

In the physically based approaches and the empirical approaches, trade-offs exist between accuracy and ease of application. The physically based approaches are the most accurate and can be applied to any location, but they require a considerable number of measurements and thus increase commitments of time and cost. In contrast, the simple empirical approaches require neither a great deal of data nor extensive computation time. However, the simple approaches have limited accuracy, especially for time intervals of less than one month, and they should be used only for areas for which they are calibrated.

5.0 RECENT STUDIES

In the water balance, evapotranspiration (ET) is normally of the same order of magnitude as rainfall or run-off. Nevertheless the number of studies on rainfall and runoff far exceed the number on the process of evapotranspiration in hydrological literature. Hydrologists have always found it hard to come to grips with transpiration. Perhaps it is because of the background of many hydrologists that makes them unfamiliar with the subject. Or may be, the process of ET seems less attractive and less urgent to be studied. A third reason may be the complexity of the process of evapotranspiration. For example, measurements of runoff produce spatially integrated values. Rainfall measurements normally represent an estimate for an area. But for actual evapotranspiration the representativity is always open to question, because of the spatial variation in vegetation and soil moisture regime over a catchment. Besides, careful measurements of ET need more equipment and daily management than rainfall or runoff measurements, for which many experimental and representative basins are well equipped, but are poorly equipped with meteorological instruments, even to determine potential ET. More testing is needed of existing equipment and approaches and more development of new tools for estimating actual and potential ET.

To compute an estimate of the actual ET emanating from a particular agricultural field or vegetated surface, requires a procedure that considers and incorporates the several major processes. The objective for each ET estimate will suggest the parameters of space and time incrementation and expected accuracy. Even then, considerable averaging of inputs and effects will be needed depending on information and data available.

There has been a wide variety of actual ET prediction methods developed and reported in recent years. These range from short single equations to sets of equations in a decision logic to quite complex computer models. Each is unique in its objective, accuracy, and data required. The following are a few examples across this range of complexity.

Agricultural hydrologists have traditionally estimated the antecedent soil moisture status for events by simple water budget - ET equations such as antecedent retention index reported by Saxton and Lenz (1967). More recently, slightly enhanced water budget methods have been applied to hydrology analyses and predictive models (Haan,1972; Holtan et al.,1975). Actual ET estimates for irrigation situations have become more complex as reported by Jensen (1973), and with the exception that these methods usually do not consider water stress, they have good application potential to a variety of other agricultural situations.

Ritchie (1972) composed a model to incorporate most basic considerations through a series of interactive equations. Potential for ET was based on the Penman method, soil evaporation and transpiration were computed separately, a leaf-area index (LAI) described plant growth, and the soil water for the entire profile was budgeted, all on a daily time sequence.

A comprehensive model to compute daily actual ET from small watersheds was developed and reported by Saxton et al.(1974). This model, shown schematically in figure 17 separates the major climatic, crop, and

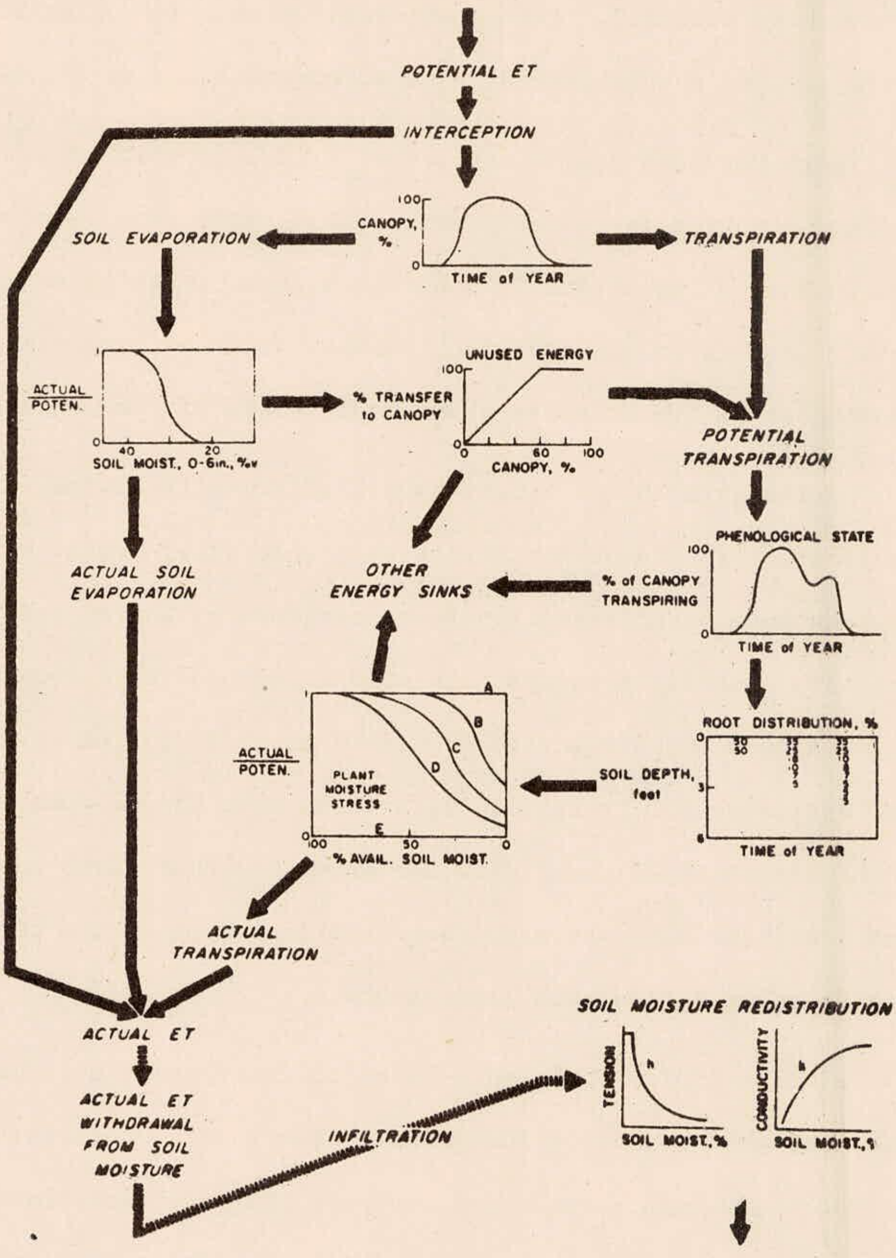


FIGURE 17 - CALCULATION FLOW CHART FOR COMPUTING DAILY ACTUAL EVAPOTRANSPIRATION AND SOIL WATER PROFILES BY THE SPAW MODEL

[Source : Saxton et al.(1974)]

soil effects into a calculation procedure with emphasis on graphical representation of principle relationships. Calculated amounts of interception evaporation, soil evaporation, and plant transpiration are combined to provide daily actual ET estimates.

Beginning at the top of figure 17, a daily potential ET is computed by any one of the several methods. Intercepted water at the plant and soil surfaces is then considered to have first use of the potential ET energy, and no limits are imposed. Remaining potential ET is divided between soil water evaporation or plant transpiration according to plant canopy present. 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 canopy, a percent of the unused soil evaporation potential is returned to the plant transpiration potential to account for radiated and convected energy from the heated soil and air. Actual transpiration is computed through sequential consideration of :

(1) plant phenology to describe the transpirability of the existing canopy,

(2) a root distribution to reflect where in the soil profile the plant is attempting to obtain water, and

(3) a water stress relationship which is applied to each soil layer and is a function of the 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 soil layer with roots, 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.

Several methods have been reported which apply best to basin hydrology or specific land uses. The ET method by Crawford and Linsley (1966) in their Stanford Watershed Model IV considers the effect of areal variation through a coefficient which estimates percent of area which attains varying percentages of an evaporative opportunity, which is in turn a function of a time-dependent coefficient to represent crop growth. Using pan evaporation as a potential ET, this model combines estimated actual ET from interception, two soil zones, and ground water to estimate a total daily actual ET. No attempt is made to separate soil evaporation from transpiration, and calibration parameters are obtained by fitting to observed watershed data.

Morton (1976) presented a method based on regional climatic data and showed its application to many large basins in the United States and Canada. Hanson (1976) and Aase et al. (1973) have developed ET prediction equations for native rangeland of the western U.S. Federer (1979) described a detailed model of transpiration that simulates both diurnal and growing season behaviour of a hard wood forest.

Other methods have restricted their data inputs to readily available climatological data which often makes them more practical than more sophisticated methods. Eagleman (1967) described a method based on air temperature and humidity. Kanemansu et al. (1976) and Rosenthal et al.(1977) described methods using air temperature, R_n, and LAI.

Several methods have been developed which describe the ET processes within the soil-plant-atmosphere system. The soil-plant-atmosphere model (SPAM) described by Shawcroft et al. (1974) treats the ET and plant growth characteristics in detail.

A similar model reported by Hanks et al.(1969) concentrates more on the soil water and its plant interaction.

Even more sophisticated models are being developed as improved computing capacity and ease of programming become available. Kristensen and Jensen (1975) applied a detailed ET model to the crops of barley, sugarbeets, and grass over a 4-year period with reported accuracy of 10 percent, and Wright and Jensen (1978) reported similar success for irrigated conditions with a different model. Some of the recent studies are given below:

Morton

Morton (1978) reported that the concept of a complementary relationship between the evapotranspiration from an area and the potential evaporation at some point in the area is diametrically opposed to the conventional wisdom. However, pan and dish evaporation data from irrigated areas and adjoining deserts provide evidence that it reflects reality. Furthermore it provides the basis for a model that permits evapotranspiration to be estimated from the routine observations of temperature, humidity and sunshine duration that are used in computing potential evaporation, thereby eliminating the need to represent the complexities of the soil-vegetation system by locally optimized fudge factors. This means that the model is verifiable so that errors in the associated assumptions and empirical relationships can be detected and corrected over an ever-widening range of environments. The current model represents the culmination of such a methodology. It is calibrated at climatological stations in desert areas - where the monthly precipitation approximates evapotranspiration - and applied over a wide range of environments without local optimization of coefficients. The practicality of this unorthodox approach is demonstrated by comparing model estimates of evapotranspiration with the corresponding water budget estimates for 122 river basins in Canada, Ireland, Kenya and the southern U.S.A.

Stricker and Brutsaert

Stricker and Brutsaert (1978) determined actual evapotranspiration in a rural catchment "Hupselse Beek", situated in a sandy region in the eastern part of the province of Gelderland, in The Netherlands; the experiment covered a period of nearly 80 days in the summer of 1976, during which a severe drought was experienced in western Europe. Actual evapotranspiration was obtained indirectly by the energy budget on an hourly basis, in which the sensible heat flux was calculated from the mean temperature profile and the wind speed. The sensitivity of the results was analyzed for the choice of the roughness and the zero-plane displacement height in the wind profile, and for errors in the measured temperatures. The calculated results were quite insensitive to the exact formulation of the Monin-Obukhov functions for the mean wind and temperature profiles. However, it was found that the effect of non-neutral atmospheric conditions can not be neglected; the effect of the buoyancy due to water-vapour stratification on the overall stability of the atmosphere was also noticeable. A comparison with a measure of potential evapotranspiration showed that the actual evapotranspiration was far from the potential.

Black

Black (1979) measured the rate of evapotranspiration from thinned and unthinned stands of Douglas fir using energy and water balance methods. At high values of soil water storage in the root zone the evapotranspiration rate was approximately 80% of the equilibrium evaporation rate. Below a critical value of soil water storage the ratio of the evapotranspiration rate to the equilibrium evaporation rate (ET/E_{eq}) tended to decrease linearly with decreasing soil water storage. The critical values of soil water storage in the root zone were 11.8 and 8.3 cm for the thinned and unthinned stand, respectively. Below these critical storage values, there was approximately 3.5 cm of water remaining in both root zones that was extractable by the trees. The relationship between ET/E_{eq} and the fraction of extractable water in the root zone for both stands was very similar for sunny days. In this relationship, ET/E_{eq} began to decrease when there was approximately 40% of the extractable water remaining in the root zones of both stands.

Brutsaert and Stricker

Brutsaert and Stricker (1979) calculated actual regional evapotranspiration by means of a procedure requiring only meteorological data, which are those commonly used in the various versions of the combination approache for potential evaporation. The approach is based on a conceptual model involving, first, the effect of regional advection on potential evaporation, and second, an assumed symmetry between potential and actual evaporation with respect to the evaporative power of the air in the absence of advection. Thus the degree of nonavailability of water for evapotranspiration, that is the aridity of the region, is deduced from the regional advection of drying power of the air, as implied by the atmospheric conditions. The approach was found to give good agreement with daily data of evapotranspiration obtained by means of an energy budget method for a period of severe drought in a rural watershed in a sandy region. One of the advantages is that no soil moisture data, no stomatal resistance properties of the vegetation, nor any other additional aridity parameters are required to determine actual evapotranspiration.

Beven

Beven (1979) briefly described the evapotranspiration component, based on the Penman - Monteith equation, of the distributed Systeme Hydrologique Europeen (SHE) model of catchment hydrology. The importance of evapotranspiration predictions to the operation of the model is stressed. However, such predictions may be expected to be in error and ideally it would be desirable to predict the error variance associated with estimates of evapotranspiration. This is not yet possible but the sensitivity of such estimates to errors in input data and estimated model parameter values can be investigated. It is shown that for dry canopy conditions at three sites within a broadly humid temperate region, the sensitivity of Penman - Monteith estimates of evapotranspiration to different input data and parameters is very dependent on the values of the aerodynamic and canopy resistance parameters that introduce the influence of vegetation type into the predictions. For forest surfaces in particular, the evapotranspiration predictions are highly sensitive to values of the canopy resistance, so that accurate estimation of this parameter is important.

Morton

Morton (1983) formulated and documented the most recent version of the complementary relationship areal evapotranspiration (CRAE) models. The reliability of the independent operational estimates of areal evapotranspiration is tested with comparable long-term water-budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand. The practicality, and potential impact of such estimates are demonstrated with examples which show how the availability of such estimates can revitalize the science and practice of hydrology by providing a reliable basis for detailed water-balance studies; for further research on the development of causal models; for hydrological, agricultural and fire hazard forecasts; for detecting the development of errors in hydrometeorological records; for detecting and monitoring the effects of land-use changes; for explaining hydrologic anomalies; and for other better known applications. It is suggested that the collection of the required climatological data by hydrometric agencies could be justified on the grounds that the agencies would gain a technique for quality control and the users would gain by a significant expansion in the information content of the hydrometric data, all at minimal additional expense.

Roberts

Roberts (1983) reported that transpiration from forests is difficult to measure adequately by using sampling techniques within tree crowns. Transpiration from European forests estimated using either soil-moisture abstraction or micrometeorological techniques revealed very little variability between studies and the losses were much less than suggested by calculations of potential transpiration. The influences of forest understories, negative feedback effects of surface resistance with climatic demand, and an insensitivity to changes in available soil moisture are all considered important in explaining the observed results. In addition selection, by nature or forestry enterprise, of species known to show good growth and survival at a given site may also limit the variability of transpiration.

Morton et al.

Morton et al.(1984) evolved a program WREVAP based on the complementary relationship between areal and potential evapotranspiration. It comprises the only technique yet available that can use routine climatological observations to provide reliable independent estimates of areal evapotranspiration, wet surface evaporation or lake evaporation anywhere in the world with no need for locally calibrated coefficients. It includes the CRAE (complementary relationship areal evapotranspiration), CRWE (complementary relationship wet-surface evaporation and CRLE (complementary relationship lake evaporation) options. The program can be run for time periods ranging from one day to a month. It has been demonstrated [Morton (1983)] now independent estimates evaporation and transpiration, which have been until now the largest and most intractable unknown in the hydrological cycle, can do much to overcome the current stagnation in the science and practice of hydrology by permitting solutions to the water balance equation, by providing a realistic basis for hydrological forecasts, by detecting and monitoring the effects of land-use changes, by detecting the development of errors in hydrometeorological records and by other more obvious applications. Program WREVAP expedites the exploitation of these potentialities by the flexibility of its input, time period and output requirements.

Duru

Duru (1984) developed an evapotranspiration model for application in Nigeria, which parallels that proposed earlier by Blaney and Morin. The model,designated as the Blaney-Morin-Nigerial evapotranspiration model,. predicts potential evapotranspiration with accuracy and consistency that are better than the Penman model, under Nigerian conditions. It is suggested that the Blaney-Morin evapotranspiration concept may have similar potential elsewhere when given specific form with appropriate constants derived to reflect climatic peculiarities. The empirical formulation of the presented model is -

$$PET = r_f (0.45T + 8)(520 - R^{1.31})/100 \quad \ldots(122)$$

where, PET is potential evapotranspiration (mm/day); r_f is the radiation term defined as the ratio of maximum possible radiation to the annual maximum; T is the summation of the mean daily temperature (°C) over a month divided by the number of days in that month, where mean daily temperature was obtained by averaging the daily maximum and minimum temperature; R is relative humidity (%) obtained by summing the daily means of relative humidity at 09^h00^m GMT and 15^h00^m GMT over a month and dividing by the number of days in that month.

Halldin

Halldin et al.(1984) conceived a model to work with routine meteorological data while still retaining the basic physics. Prediction of forest water balance for estimation of, e.g. stress periods, is always subjected to the conflict between data availability and the desire to approach physical reality. Values of all nine parameters were estimated independently for a 120 year old oak stand in Fontainebleau forest near Paris, France. Weekly measurements of soil-water content and throughfall served as validation for the wet 1981 and the dry 1982. Model sensitivity was assessed to changes in the values of the surface and aerodynamic resistances, the interception capacity, the field capacity and the soil-water content at onset of soil-water stress. A comparison of different formulae to predict potential transpiration/evaporation clearly demonstrated the necessity to include a surface resistance term in a forest transpiration formula. The aerodynamic resistance was less important and the simple threshold formulation for interception did not always work well. The overall behaviour of the model was primarily limited by the availability of precipitation data very near the study site.

Ali and Mawdsley

Ali and Mawdsley (1987) tested the advection-aridity model [Brutsaert and Stricker (1979)] for estimating actual evapotranspiration, ET with over 700 days of lysimeter evapotranspiration and meteorological data from barley, turf and rye-grass from three sites in the U.K. The performance of the model is also compared with the API model [proposed by Mawdsley and Ali (1985)]. It is observed from the test that the advection-aridity model overestimates nonpotential ET and tends to underestimate potential ET, but when tested with potential and nonpotential data together, the tendencies appear to cancel each other. On a daily basis the performance level of this model is found to be of the same order as the API model; correlation coefficients were obtained between the model estimates and lysimeter data of 0.62 and 0.68 respectively. For periods greater than one day, generally the performance of the models are improved.

Bringfelt and Lindroth

Bringfelt and Lindroth (1987) developed and tested a model at one site and applied it to another site where validation data such as transpiration and interception were available. The model was based on the Penman combination equation for transpiration and on Rutter's equation for interception evaporation. The transpiration was estimated from energy balance/ Bowen ratio measurements at both sites and the interception evaporation was estimated from measurements of through fall and gross precipitation. Total evapotranspiration calculated by the model was within 10% of the evapotranspiration of the application site when using parameters estimated from the test site data. However, this agreement was apparent in the sense that the model underestimated the transpiration by about 20% and overestimated the interception evaporation by about 60%. An attempt was made to derive an independent parameter set on basis of the test site values. The parameter determining the magnitude of the surface resistance was scaled with respect to the difference in leaf area index between the forests and with respect to the relative differences in stomatal resistance between the different species. Using this modified parameter set resulted in a 50% underestimation of the transpiration. It was concluded that factors other than forest density and species were responsible for the scaling of the surface resistance between different forests.

Kovacs

The information of actual areal evapotranspiration has paramount importance in hydrology, but a method easily applicable in the practice and providing acceptable accuracy for its estimation is still missing. The concept based on the complementary character of actual and potential evapotranspiration offers a promising way. The practical application of this approach requires the precise interpretation of the independent variables considered in the model. Kovacs (1987) analysed and stated that the complementary approach of potential and actual evapotranspiration provides a promising method of estimating the areal average of ET without requiring the detailed investigation of plant-soil-water systems. The model proposed by Morton can be further simplified by introducing a slight modification in the starting hypothesis. An important characteristic of the model derived in this way is the different interpretation of the space-scale used for the calculation of the radiation and the ventilation terms respectively. ET_R is an average variable describing the actual amount of energy (depending also on surface conditions) available within the whole area, while ET_V gives the integrated effects of the varying surface at the down-wind edge of the area. When selecting the data to be substituted into the model, this difference should be considered.

Hargreaves and Samani

Hargreaves and Samani (1988) presented relationships for use in estimating the standard deviation of PET. Some weather simulation procedures require mean values of potential evapotranspiration (PET) and the standard deviations in PET. For purposes of irrigation planning and design it may be desirable to know the probable deviations from mean values of PET. Various crop production or crop yield models have been developed that require input of daily measurements of climate. A weather simulation procedure utilizing a monthly climatic data base can be substituted for the daily climatic data to produce very comparable results. The weather simulation procedure recommended requires the standard deviation of PET. A series of monthly mean values of maximum and minimum temperatures provides the required data for estimating mean PET and the standard deviation. If only long-term mean maximum and minimum temperatures and the mean temperatures for a series of years are available, the standard deviation of mean temperature provides a means for making an estimate of the standard deviation in PET. The relationship is influenced by latitude and elevation.

Not all recent methods of modelling ET have been cited. Those that are presented vary widely in their objectives, the amount of details required, computation time, and required data. Comparison of the several methods shows that the best results are obtained when a potential ET is derived from an energy balance - aerodynamic method and the soil-plant-atmosphere system is represented by dynamic simulations of the major processes which determine actual ET. Thus, each ET application must match the prediction method to the objectives.

6.0 POTENTIAL EVAPOTRANSPIRATION OVER INDIA

Evapotranspiration studies in India, particularly based on rigorous formula of Penman, are extremely few. Most of the earlier studies were based on Thornthwaite formula. Using the Penman's formula, Rao et al. (1971) computed normal monthly and annual potential evapotranspiration of about 300 stations in and near India. Meteorological data for computations were taken from 1931-60 climatological tables of observatories in India published by the India Meteorological Department. Some of the important features noted in potential evapotranspiration (PET) distribution are described below. All PET values are given in unit of cm per month.

Monthly PET

Winter (Jan. - Feb.)

January

PET decreases latitudinally north of Lat. 20°N and becomes less than 4 cm over extreme northern part of Punjab. At Srinagar in Kashmir, PET is less than 2 cm. At Leh it is only 1.6 cm. This is due to both elevation and the western disturbances which affect the area. Over Assam and Sub. Himalayan Bengal also, PET is low (4-6 cm). In the Peninsula, PET is high (13-14 cm) along the west coast and south of Lat.15°N. Extreme southeast of Tamil Nadu (Tuticorin to Kanya Kumari) has the highest PET of more than 14 cm per month.

February

A general rise of 2 cm over January values is noted north of Lat. 20°N. The Peninsula is nearly a flat area, values ranging from 12 to 14 cm. High PET (13-14 cm) continues along the west coast and the extreme southeast of Tamil Nadu.

Summarising, in winter PET is high over the Peninsula where it ranges from 12 to 14 cm per month. PET decreases gradually northwards to less than 4 cm in Kashmir.

Summer (March-May)

March

PET has risen rapidly over the whole country. The rise is of the order of 4 cm. High values along the west coast are maintained. The highest PET is now over the northern half of the Peninsula between Lat. 15° to 20°N and Long. 77° to 80°E, covering the states of Karnataka and Andhra Pradesh. Raichur, Gulbarga and Anantpur in these states have each PET of 19 cm.

April

PET is high all over the country. It ranges from 16 to 20 cm over most of the area north of 15°N and west of 85°E. A small area surrounding Jalgaon in north Maharashtra near Lat. 20°N and Rajkot have PET greater than 22 cm.

May

This is the hottest month of the year over most of the country. PET has increased further and the gradient is generally steep. It is more than 20 cm over the country north of Lat. 15°N and west of Long. 88°E. Elevated plateau, mountainous and coastal areas are excluded. The highest PET is over Saurashtra, West Rajasthan and the areas around Jalgaon. A few stations in these areas have very high values of the order of 30 cm. The regions of low PET (14 cm) are over (i) west coast of the Peninsula south of Lat. 15°N, and (ii) Assam, east of Long. 90°E.

With the advance of the summer season, there is (i) increase of PET north of Lat. 15°N, and (ii) the gradient becomes more steep. The region of high PET shifts to north Maharashtra, Saurashtra and west - Rajasthan. The only region where PET has nearly the same value with the advance of the season is the west-coast south of Lat.15°N.

Monsoon (June - September)

June

This is the month when the southwest monsoon sets in over the Peninsula and other parts of the country. PET has generally decreased over the country except in Tamil Nadu and northwest India. Over west Rajasthan PET has risen to more than 28 cm, Jaisalmer showing 31 cm. Values have become less than 10 cm along the west coast of the Peninsula south of Lat. 15°N. In Assam also, PET is low (11-12 cm).

July

The monsoon is established over the whole country. PET has generally decreased in all areas under the influence of the monsoon. Values range from 10 to 16 cm over most parts of the country. The zone of highest PET is west Rajasthan with more than 24 cm. Extreme south-east Tamil Nadu has a small pocket of high values round Tuticorin.

August

The general features are similar to July, though a decreasing tendency is noted except in south-east Tamil Nadu. A pocket of low PET exists near Jabalpur.

September

There is general decrease of 2 cm all over the country. East of Long. 80°E, there is little variation, the general value being 10 cm. The region of highest PET is west Rajasthan and the extreme south-east of the Peninsula.

Post Monsoon (October - December)

October

Excepting over Saurashtra and adjoining areas of west Rajasthan and small areas in the extreme south-east of the Peninsula, PET is generally 10-14 cm. The region of highest PET is over Saurashtra and Kutch where it is 16 cm.

November

With the setting of the winter conditions, PET has decreased further over the northern parts of the country to 4 cm in Punjab and the variation upto 20°N is almost latitudinal. High PET (12-13 cm) is over west coast of the Peninsula between 16° and 21°N Latitude and adjoining parts of Saurashtra, and Raichur-Sholapur area. Peninsula is mostly a flat area with very little of gradient.

December

PET increases from less than 4 cm in Punjab almost latitudinally from north to south except for the west coast where the variation is little. The highest PET is along the west coast and the Tuticorin-Kanya Kumari area.

Seasonal PET

Over most of the country outside Rajasthan, PET for the south-west monsoon season (June - September) ranges between 40 to 60 cm. It is less than 40 cm over Assam. Areas where it exceeds 60 cm are (1) Rajasthan,

(2) East Coast south of Lat.15°N, (3) Gulbarga-Anantpur area, (4) Jalgaon-Dohad area, and (5) Lucknow-Orai area. Value exceeds 80 cm over west Rajasthan and in the extreme southeast of Tamil Nadu.

Annual PET

Annual PET ranges between 140 to 180 cm over most parts of the country. Annual PET is highest over extreme west Rajasthan in Jaisalmer area. Extreme south-east of Tamil Nadu in Tuticorin area shows high values of the order of 180 cm. PET is less than 140 cm in Coastal Mysore, parts of Bihar Plateau and east Madhya Pradesh and Assam.

7.0 CONCLUSIONS

In the previous chapters a number of methods and models have been discussed. In practice there are difficulties in choosing the right method or model and translating the one-dimensional model results obtained at one point to larger areas. The complex reality has to be schematized into a proper way, i.e. one has to distinguish the soil-plant-atmosphere system in a limited number of characteristic situations. Important properties are :

(a) soil use as far as it influences evapotranspiration through its roughness, rooting depth, etc;

(b) soil physical parameters as soil water characteristic, capillary conductivity and derived properties such as infiltration capacity, water availability for the plant and storage coefficient;

(c) topographical data such as altitude, slope, etc.

In the first place the choice of method or model depends on the objective of the study. Secondly, it depends on the availability of the above mentioned data. Reversely, collection of data depends on the type of model chosen. Most problems are encountered with regard to the following items :

(i) estimation of rooting depth, also as function of time. A good estimation of this depth is very important for the calculated evapotranspiration;

(ii) influence of very wet conditions upon plant growth and transpiration;

(iii) reaction and adaptation of an active growing plant system on changes in external circumstances. The history of water stress can be very important for the subsequent reaction of crops on water shortages;

(iv) soil loosening and compaction have distinct effects on soil physical properties. Mostly, however, they are taken as invariant with time;

(v) translation of the soil map into a map of soil physical characteristics. Testing has to be done to extrapolate data measured in a particular soil to unmeasured soils with identical classification elsehwhere;

(vi) spatial variability of soil physical properties. Even seemingly uniform soil areas manifest large variations in hydraulic conductivity values;

(vii) fast and simple determination of the soil physical properties;

(viii) field determination of the drainage respectively infiltration resistance of open water courses;

(ix) in simulation models topographical variation of the soil surface within the area of investigation, is seldomly taken into account;

(x) representativity of meteorological data;

(xi) coupling of the meteorological system and the soil plant system. Micro-meteorological approaches are in fact bounded by the soil water pressure head in the root zone, while on the other hand the soil physical models are bounded by the atmospheric conditions. This is a drastic simplification of reality;

(xii) accessibility of data. The present-day models need much data which should be computer accessible. Especially data on crops, soil physical and hydrological properties for larger areas that can be applied on a routine basis are scarce and difficult to obtain;

(xiii) choice of the method or the model that is best suitable to apply to a certain situation; and

(xiv) formulation of the problem in hydrological term.

Future Outlook

The principal objective of evapotranspiration research is to find methods for calculating the loss of water under varying conditions of climate, soil and vegetation.Various groups need evapotranspiration data in their work, for example agriculturists, horticulturists, hydrologists, environmentalists, drinking-water suppliers etc.

The data mostly wanted are reliable areal evapotranspiration data. It is still questionable whether remote sensing techniques from satellites will give an exact answer to this. It is to be expected that in the future remote sensing techniques will be more reliable than nowadays. A drawback of using satellite data is, however, that polar orbiting satellites pass only once a day. So the diurnal variation in evapotranspiration can only be estimated by mathematical models that are verified with satellite data. Moreover remote sensing nowadays gives only relative surface temperatures, so that on the other hand satellite data have to be calibrated with additional measurements on the spot. It can be expected that in the future the research on catchment areas will go on, to improve our knowledge of evapotranspiration as a function of atmospherical and soil physical circumstances.

Some improvements can be expected in measuring meteorological data for evapotranspiration calculation, in data-handling. Some research can be done on the reliability and comparability of lysimeters. Due to the computer, integrated mathematical-hydrological models are expected to be developed which can take into account the influence of human activities and changes in land use on the evapotranspiration. Further more the computer has the advantage that lots of data can be handled and computations can be done on a real-time basis. This means that the user may have the disposal of evaporation data at short notice, in every kind of output he wants (total values, mean values, frequency distributions). In the future the evaporation data might be combined with precipitation data so that real-time groundwater levels can be calculated, dependent on the type of soil.

Although evapotranspiration is essentially a physical process it is not to be expected that in the future models will be developed that can take into account all atmospheric, crop and soil properties influencing the process. As the complexity of the models increases, the data requirements to drive the equations often make the model useless for routine applications. This means that also in the future the calculation of evapotranspiration will be a compromise between empirical and totally physically justified models.

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