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EVAPORATION AND THE PRINCIPLES OF ITS MEASUREMENT

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1. Introduction

The term 'evaporation' is intended to cover direct water loss from soil and water surfaces, as well as transpiration loss from vegetation. All these processes are essentially similar, and both soil and plants may be thought of as alternative paths through which water flows on its way to the surface from which it finally diffuses into the air. In the case of vegetation, the heterogeneity of the effective surface, and the biologically controlled variability of the resistance to internal flow, are merely additional complications.

2. Factors influencing evaporation

Evaporation is best thought of in terms of three interdependent influences. The first of these is the availability of energy at the evaporating surface, to supply the latent heat demand. Possible sources are radiation from the sun, sky and clouds, and sensible heat transferred from the adjacent air and soil. The ability of the air to transfer water vapour away from the surface. This depends on the vertical gradient of humidity and the effectiveness of turbulent mixing in the lower air layers. The third influence is the availability of water at the effective evaporating surface, which in the case of a land surface depends on the permeabilities of soil and vegetation, and on the prevailing soil moisture status.

3. Determination of evaporation

a) Empirical formulae: A variety of attempts have been made to arrive at estimates of evaporation from standard meteorological data. Of the many empirical formulae derived for evaporation or potential evaporation (Lowry and Johnson, 1941; Thornthwaite, 1948; Blaney and Criddle, 1950; Hargreaves, 1956), most are based on average temperature, often with an allowance for day length, generally via latitude and time of year. In addition some, such as that of Blaney and Criddle, allow for differences between the responses of different crops, while Hargreaves goes a step further and allows for atmospheric humidity as well. Basically all these formulae attempt to estimate evaporation from wet surfaces, and their application to even partially dry surfaces would lead to considerable error. Formulae such as those of Penman (1948) and Ferguson (1952) have a better basis of physical reasoning but also involve a number of assumptions and approximations, and hence restrictions.

b) Actual measurement: Under most agricultural conditions, the main interest lies in the need of knowing precisely the value of water that is being evaporated and in the above approaches the usual simplifying assumptions break down, and it is most important that measurements of actual evaporation should be used whenever possible.

There are four main approaches available, each of which will be dealt with inturn. These depend on —

— (1) The water balance of the evaporating system;

— (2) The upward flow of water vapour in the air layers near the ground;

— (3) The energy consumed in the conversion of liquid water into vapour

— (4) Combinations of (2) and (3).

Water balance

The basic relationship employed here is the hydrological equation,

$$E = P - RO - UD - \Delta W$$

Where E is evaporation, P precipitation (including irrigation); RO net surface runoff; UD underground drainage below a chosen depth and ΔW the change in water content; W, of the soil down to that depth - all for the same time interval.

This method has been applied most successfully to lakes and dams, although under favourable conditions it also appears satisfactory for land surfaces, from large river valleys down to experimental plots. However, as sampling variability is high, even with considerable effort and elaborate equipment and sampling procedures, good results can rarely be expected over periods less than a week.

EVAPORATION FROM FREE WATER

(a) Lakes and ponds: The evaporation rate from a large water surface is itself affected by such local characteristics as the depth and turbidity of the water, and also by its rate of mixing. These together determine the vertical extent of water which shares in the incoming energy supply during the day (or in summer) and which supplies energy to the surface at night (or in winter). In other words they determine the effective heat storage capacity of the lake and hence the degree of 'lagging' of its daily and annual cycles of temperatures and vapour pressure.

(b) Standard evaporimeters: Because of its simplicity of construction and operation a standardised tank of water (for which $E = P - \Delta W$) has become widely used for 'evaporation' measurement. In actual fact it measures nothing more than a

[illegible] local value of the power of the atmosphere to extract water from a tank of that size, shape and colour, in that type of surround. This power is by now well known to depend considerably on the construction and exposure of the tank. It is generally greater for above-ground than for sunken tanks, and for small ones compared with large tanks. Adoption of U.S.W.B. class A pan was suggested by W.M.O. for international comparisons. There are also other types of evaporimeters such as Piche atmometer, shallow tanks, black-painted tanks etc.

LYSIMETRY

With land surfaces a much closer approach to natural conditions can be obtained by the use of lysimeters containing the appropriate soil and vegetation. A properly designed lysimeter, suitably filled and exposed and correctly operated, undoubtedly constitutes the most direct, acurate and reliable means for determining true evaporation from almost any type of land surface.

a) Weighed lysimeters: For shorter periods, or with natural as against 'potential' evaporation, some means of determining ΔW becomes absolutely necessary. For the best method is to weigh the soil block, container and all, necessitating a separate outer container and a balance. Although this can add considerably to the cost and complexity of a lysimeter installation, it is justified for teaching and research purposes and can provide an excellent means of calibrating other methods of evaporation measurements.

b) Floating lysimeters: An alternative approach is to replace balances altogether with a hydraulic system; i.e., to float the soil container (together with any necessary buoyancy chambers) in a suitable liquid (e.g. zinc chloride) held in the outer container, and then detect the changes in level as the unit rises out of the liquid with loss of weight by evaporation, or falls back with gain of weight. These lysimeters are comparatively cheaper and simpler. These can easily be adopted to continuous remote recording. On the other hand, these are liable to temperature error and errors due to dynamic effects of wind disturbance near the air-vents. However, corrections are generally possible.

Potential evaporimeters:

These are really a specialised form of lysimeter, in which a single container is set directly in the soil, and both it and its immediate surroundings are kept plentfully supplied with water, either by overhead irrigation or maintenance of a fairly high water table. The water consumption in the container is determined by metering the irrigation or water table supply (plus rainfall) and substracting the amount of percolate, collected from a vent underneath. The quantity measured by such units is called commonly as potential evaporation. This concept was introduced by Thornthwaite (1948). It is the hypothetical rate of water loss from a large homogeneous area of continuous green crop, under the given meteorological conditions, when there is no resistance to water supply at the evaporating surface.

Local measurement of soil water content:

In applying the hydrologic equation, the parameter soil water storage, $\triangle W$ requires either a weighing technique or else in situ measurements of soil water content at various depths and various locations within the area being investigated.

The various techniques are (a) Direct gravimetric sampling using tubes or augers, (b) electrical conductivity of porous blocks, and (c) neutron dispersion, etc.

VAPOUR FLOW METHODS

The flow of water vapour upward into the atmosphere is governed by turbulent diffusion caused by essentially random movements of the air which give rise at any point to a constantly fluctuating vertical component of motion.

The mean rate of this transfer, per unit time and per unit area, is known as the vertical eddy flux of water vapour. Letting a bar over a quantity denote its mean value with respect to time, and a dash the difference between an instantaneous value of a quantity and its mean value, the eddy flux is given by

$$E_z = \overline{(\rho W)' q'}$$

where E_z is the eddy flux at height z, ρ, W and q are simultaneous values of air density, vertical wind speed and specific humidity at the measuring point, also at height z.

There are three different ways of arriving at atmospheric fluxes.

(a) Direct eddy flux determination

It is obviously impossible to measure instantaneous values of W and q, but with sufficiently rapid response instruments like hot wire anemometers and fine wet & dry bulb thermometers. The actual rapidity of response required decreases with height above ground but increases with wind speed. It also varies with the roughness of the surface and with certain weather factors.

THE DALTON EQUATION

$$E = a(b + u)\,(e_o - e_a)$$

where E is evaporation (as a rate rather than a quantity), a and b are empirical constants and $(e_o - e_a)$ is the vapour pressure difference between the surface and the bulk air. a(b+u) represents the conductance of the layer from surface to measuring height; with a essentially a function of surface aerodynamic roughness. b arises from the contribution to the total transfer of free or thermal convection, which becomes important under calm conditions. Generally b will be small and often taken as zero.

For large bodies of water, particularly when averaged over long periods and for the whole area the Dalton formula is now fully stabilised.

$$E = \frac{123}{p} u_2 (e_o - e_2) \text{ mm day}^{-1}$$

where u_2 is 2 meter wind speed in m $\sec^{-1}$

p is the mean barometric pressure

with modified constants the Dalton equation should also hold for natural surfaces other than water.

THE AERODYNAMIC OR PROFILE METHOD

This considers the turbulent transfer of water vapour at a small distance above ground, rather than the overall process of ground-to-air transfer. The equation is

$$E_z = \rho K_w \frac{\partial q}{\partial z}$$

where K_w is the eddy transfer coefficient for water vapour at height z. The gradient $\partial q / \partial z$ can be determined from measurements of the changes in humidity with height. Strictly speaking the above theory only holds under natural conditions, i.e., when the lapse rate or rate of temperature fall with height, is small. However, under non-neutral conditions, which are often occasions of strong evaporation, modifications are suggested by workers like Deacon, Priestly, Webb, Swinbank etc.

To sumup, the profile approach seems at present suited mainly for special purpose short-period measurements, involving only a limited amount of instrumentation - although for accuracy great care in measurement and analysis is necessary.

ENERGY BALANCE METHOD

The basic equation applying here is

$$R = H + LE + G$$

where R is the net radiative flux received by the surface;
G is the heat flux into the ground;
H is the sensible heat flux into the air;
LE is the latent heat flux into the air;
L being the latent heat of evaporation and
E the upward flux of water vapour.

G will generally be small relative to other quantities and over long periods can be neglected or estimated as a fraction of R. For most water bodies of economic interest, the energy balance method must be regarded as only suitable for annual, seasonal or perhaps monthly evaporation.

Where humidity measurements are made by other than psychrometers such as hair element, dew point, lithium chloride or other divices, the formulae appropriate to the instrument used is to be employed, where $B = \frac{C_p \Delta T}{L \Delta q}$; ΔQ is difference in specific humidity.

$$LE = \frac{R - G}{1 + B}$$

B (Bowen ratio) = H/LE.

Where vapour pressure differences can be obtained as directly and easily as wet bulb temperature differences, then the following equation can be used:

$$E = \frac{R - G}{L}\left[1 - \frac{\gamma}{S + \gamma}\left(\frac{\Delta T}{\Delta T_w}\right)\right]$$

where ΔT and ΔT_w are measured differences of dry bulb and wet bulb temperatures respectively and $\gamma/S+\gamma$ is a slowly varying temperature dependent weighting function, obtained from tables or graphs.

COMBINATION METHOD

Combining the Dalton and energy balance approaches, three alternative procedures have been put forward, independently by Penman (1948), Ferguson (1952) and Slatyer and McIlroy (1961). The first and second of these provide, essentially, estimates of the free water evaporation, while the third aim at estimating potential or even actual crop evaporation. The basic combination formula is

$$LE = \frac{S}{S + \gamma}(R - G) + h\,(D - D_o)$$

where $h = a(b + u)$ is a Dalton-type coefficient and the subscript 'o' denotes a value at the zero height.

Often for more demanding purposes provided data adequately representative of the crop concerned are used; the best formulae are

1. Where the surface is effectively saturated and therefore D_o is nearly equal to 0, we speak of potential evaporation E_p given by

$$LE_p = \frac{S}{S + \gamma}(R - G) + hD$$

Potential evaporimeter or a lysimeter measures this quantity. It varies with vegetation type, reflectivity, roughness etc., and also location in particular according to the strength of advection and hence on the nature of the upwind fetch.

2. When a large area surrounding the measuring point with freely transpiring vegetation as often mentioned in literature what is actually meant is not potential but equilibrium evaporation. In a indefinitely large, uniform, thoroughly moist surface, the air near the ground tends towards saturation i.e., D tends to 0 when equilibrium evaporation is given by

$$E_e = \frac{S}{S + \gamma}\,\frac{R - G}{L} \sim \frac{S}{S + \gamma}\,\frac{R}{L} \quad \text{(on average)}$$

3. Actual evaporation is unlikely to approach E_e. In non-saturated surfaces it is generally difficult to measure D_o directly or even to estimate with accuracy. However, it can be estimated by using the expression

$$E = \frac{E_p}{1 + \frac{\gamma}{S + \gamma}\,\frac{h}{h_i}}$$

where h_i is the internal conductance analogous to h but applicable to molecular diffusion of water vapour through the stomata of leaves or through dry top layers of soil as the case may be.

$$h_i = C\,(M - M_c)$$

where M = soil moisture content; C and M_c are empirically determined constants.

THE PENMAN APPROACH

The Penman formula designed by Penman in 1948 for the estimation of evaporation from a free water surface and of potential evaporatinspiration from a vegetative cover, has been widely used throughout the world for the last 30 years with generally satisfactory results. The formula has been modified by Penman (1963) and others. A simple method for field calculations of the Penman formula was given by Frere and Popov in 1979. It is as follows:

a) Potential evapotranspiration

$$ET=\frac{\frac{p_o}{p}\cdot\frac{\Delta}{\gamma}\left[0.75R_A(a+b\frac{n}{N}-\sigma T_k^4(0.56-0.079\sqrt{e_d})(0.10+0.90\frac{n}{N}\right]+0.26(e_a-e_d)(1.00+0.54\text{[illegible]}}{\frac{p_o}{p}\cdot\frac{\Delta}{\gamma}+1.00}$$

b) Evaporation of a free water surface

$$E_o=\frac{\frac{p_o}{p}\cdot\frac{\Delta}{\gamma}\left[0.95R_A(a+b\frac{n}{N})-\sigma T_k^4(0.56-0.079\sqrt{e_d})(0.10+0.90\frac{n}{N})\right]+0.26(e_a-e_d)(0.50+\text{[illegible]})}{\frac{p_o}{p}\cdot\frac{\Delta}{\gamma}+1.00}$$

Explanation of the units used in the formula

The terms intervening in the formulae and in the working sheets are defined hereunder and expressed in the following units:

E_T = estimation of the potential evapotranspiration for a given period, expressed in mm;

E_o = estimation of the evaporation from a free water surface for a given period, expressed in mm;

P_o = mean atmospheric pressure expressed in millibars at sea level;

p = mean atmospheric pressure expressed in millibars as a function of altitude, for the station where the estimate is calculated;

Δ rate of change with temperature of the saturation vapour pressure expressed in millibars per degree °C;

γ the psychometric coefficient for the psychometer with forced ventilation = 0.66;

0.75 and 0.95: factors expressing the reduction in the incoming short wave radiation on the evaporating surfaces and corresponding respectively to an albedo of 0.25 and 0.05;

R_A short wave radiation received at the limit of the atmosphere expressed in mm of evaporable water (1 mm = 59 calories) and taking for the solar constant the value of 2.00 $cal.cm^{-2}.min^{-1}$;

a and b = coefficients for the estimation of total radiation from the sunshine duration (

n = sunshine duration for the period considered in hours and tenths;

N = sunshine duration astronomically possible for the given period;

σT_k^4 = Blackbody radiation expressed in mm of evaporable water for the prevailing air temperature;

e_a = saturation vapour pressure expressed in millibars;

e_d = vapour pressure for the period under consideration expressed in millibars;

T^oC = air temperature measured in the meteorological shelter and expressed in degrees Celsius;

$T^o{}_k$ = air temperature expressed in degrees Kelvin where $T_k^o=T_C^o+273$;

$U_{m/s}$ = mean wind speed at an elevation of 2 m for the given period and expressed in m/sec.

4. SURFACE ATMOSPHERE INTERACTIONS

In considering the influence of evaporation on individual surface or atmospheric properties, it is convenient to do so by examining their effect on the various terms appearing in one or the other of the expressions for evaporation. Examples of studies on different surfaces, deep and shallow water moist and dry grass, swamp etc., which effect parameters such as albedo cause change in the expression utilised for the study of evaporation. A good number of examples of studies with different surfaces can be referred in the works reported by researchers at Aspendale in Australia and other places.

5. TECHNIQUES OF EVAPORATION CONTROL

In some instances it becomes necessary to arrest evaporation to retain moisture in the soil. In such cases reduction of evaporation is restored to, with the use of chemicals spread as a thin layer over the free water or land surface. Much work in this direction has been in progress in many countries. A work in India was reported by Gangopadhyaya and Venkataraman (Evapotranspiration Reduction, Agri.Met., Vol.6, 1969,p.339-45). It was seen that initial missing of the top layer of the soil with OED green (The water emulsion of alkoxy ethanol obtained commercially by combining docosanol and octadecanol with ethylene oxyde) gave economically worthwhile reduction in evaporation from a freely drained periodically irrigated soil surface. Spraying of OED green initially on a wet soil surface appeared to be promising.

6. CONCLUSIONS

With most methods of evaporation measurement, development is still going on. In individual cases major performance improvements may well result. Hydrologic balance applications, while sound in many ways, have one basic drawback. To keep the error reliably

below 10% for periods less than a day a sensitive weighed or floated lysimeter is required. Use of neutron probe or resistance blocks etc., requires an elaborate network of careful measurements to ensure the same tolerance over several days. On the otherhand, until suitable integrating equipment becomes available, the eddy correlation and profile methods can only be utilised for short runs, generally of the order of an hour. The energy balance, Dalton and combination methods fall somewhere in between the above extremes. As regards accuracy and reliability, weighed lysimeters emerge as clearly superior at present to all other techniques, provided they are suitably designed and sited.

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