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FORECASTING MANUAL

PART IV

COMPREHENSIVE ARTICLES ON SELECTED TOPICS

20 : EVAPORATION

BY

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FORECASTING MANUAL

Part IV. Comprehensive Articles on Selected Topics

20. Evaporation

by

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1. Introduction

1.1 The problem of evaporation from lakes, lands and oceans has received wide attention since this is of great interest to hydrological engineers (Linsley, Kohler and Paulhus, 1949), meteorologists, irrigationists (Penman, 1949), foresters (Wright, 1939), plant pathologists (Canadian Dept. of Agriculture, 1940) plant physiologists and botanists (Curtis and Chazka, 1960) and also to workers in the other branches of science such as medicine, geography, climatology and certain branches of engineering, each of whom has a specific problem in evaporation. In meteorology the knowledge of the distribution of moisture in the atmosphere is important, without which an attempt to forecast weather cannot be made. The means being the principal source of supply of water vapour to the atmosphere, the study of evaporation over the oceans is of basic interest in meteorology.

1.2 Evaporation is the process of conversion of water or ice into aqueous vapour. For most practical uses evaporation implies a rate. Thus measurements of evaporation are expressed as cm, m³ or inches of water per hour, day, month or year.

1.3 In the atmosphere, evaporation proceeds from free liquid surfaces, such as seas, lakes and rivers, from solids such as soil, vegetation, snow fields and glaciers and from water drops, snow-flakes and ice crystals in the upper air. C.W. Thornthwaite (1937) has called the evaporation from liquid

water plus transpiration from plants as evapotranspiration.

1.4 The mechanism of evaporation is explained with the help of kinetic theory of gases. When a substance and its vapour are in equilibrium, molecules of the substance are continuously passing into vapour phase, and at the same time an equal number of molecules are passing from vapour to liquid phase. If the vapour pressure in the space surrounding the substance is reduced by the removal of some molecules to separate space, there is a net passage of the liquid into the vapour phase. If, on the other hand, the vapour pressure is increased, there is net flow of molecules from vapour to liquid. When the flow is from vapour to liquid, condensation is occurring. It is a difficult matter to predict precisely how much substance will evaporate under specified conditions in laboratory. In the field of meteorology, where there is no control of the factors and where it is much more difficult to measure the existing conditions accurately, the prediction can only be approximate. Some experimental statements regarding evaporation can, however, be made. They are—

- (i) Other things remaining constant, evaporation is proportional to the difference between the saturation vapour pressure at the temperature of water and the actual vapour pressure of air.
- (ii) Evaporation will be continuous only if energy is continuously supplied from some outside source (e.g. insolation).
- (iii) Evaporation will be proportional to the rate at which water vapour is removed from the immediate vicinity of the evaporating substance. For example, the evaporation rate into still air is much smaller than evaporation into a wind.
- (iv) Evaporation is more rapid from fresh than from salt water.

2. Evaporation from Soil

2.1 The amount of evaporation from bare soil depends on the depth of water table. When the soil is wet or saturated at the surface, evaporation is rapid and when the water table goes below a depth exceeding limits of capillary action, evaporation ceases. Ordinarily when the water table is below 4 ft. there is no evaporation. Randa and Malik (1939) have shown that as the distance between water table and evaporation surface is increased, the rate of

evaporation decreases rapidly. It has been observed that evaporation from sand saturated at the surface is slightly in excess than the evaporation from a free water surface. Evaporation from the average loamy soil is about 90% of that from a free water surface. For whole of the land surface of the earth, the following quantitative estimate has been derived by Wüst and Sverdrup.

Total precipitation on land	=	99,000 km ³ per year.
Evaporation from land surfaces and inland water	=	62,000 km ³ per year.
Amount supplied to oceans by run-off	=	37,000 km ³ per year.

2.2 It must be emphasised here that it is not possible to determine actual evaporation from soil or from free water surfaces by measuring the rate of loss of water from an exposed pan. When the soil surface is moist the evaporation exceeds pan measurements because the soil with its minute irregularities presents a greater evaporating surface and because its surface temperatures during the part of the day when most of the evaporation occurs are higher than water temperatures. However, when the soil surface becomes dry or partially dry, less evaporation occurs from the soil than from a pan. Even though the subsoil is moist, capillary action cannot supply the surface with water at a rate at all comparable to the evaporation from the surface of a body of water. Hence, water molecules can escape to the outer air only by a very slow diffusion process which takes place from the lower soil levels through the soil air and results in a deceleration in the rate of moisture loss.

3. Evaporation from Oceans.

3.1 As early as 1686, Halley set out to determine "the quantity of vapour raised out of the sea by the warmth of the sun". The first reasonable answer to Halley's problem appears to have been given by Brückner (1908). He extrapolated values from available observations on sea coasts and estimated that an average thickness of 106 cm was evaporated in a year from all the oceans.

3.2 Of the total solar energy absorbed at the sea surface during the course of a year, approximately fifty percent is used for evaporating sea water. Some quantitative results of Wüst and Sverdrup for the oceans are given below:

Total evaporation from oceans	=	334,000 km ³ per year.
Total precipitation on oceans	=	297,000 km ³ per year.

Table I.

Average values of evaporation(E) and precipitation(P), cm per year (After Wüst)

Latitude	Atlantic Ocean		Indian Ocean		Pacific Ocean	
	E	P	E	P	E	P
40°N	94	76			94	93
30°	121	54			116	65
20°	149	40	125	74	130	62
10°	132	101	125	88	123	127
0°	116	96	125	131	116	98
10°S	143	22	99	156	131	96
20°	132	30	143	59	121	70
30°	116	45	134	58	110	64
40°	81	72	83	71	81	84
50°	43	72	41	79	43	84

If ocean area between 70°N and 70°S is taken into consideration the average annual evaporation is found to be 99 ± 10 cm per year.

3.3 Figs. 1 to 5 show the annual and seasonal variation of evaporation over the Indian Ocean (Venkateswaran, 1956). A centre of maximum evaporation is found between the latitudes 15°S and 20°S. The lowest evaporation occurs in the equatorial zone or in the higher latitude regions of the southern hemisphere.

3.4 It has been known for a long time that evaporation is controlled by atmospheric humidity, surface water temperature and wind speed. Many empirical methods have been developed to relate evaporation to the above factors but none of them gives satisfactory results. Equations developed on theoretical basis to determine evaporation give divergent results. So far there is no single theoretical method which has a general application when use is made of temperature, humidity and wind data. However, mention can be made of the following equation which has been used to compute average evaporation over oceans, using available marine climatic data:

$$E = K(e_w - e_a) w_a \quad \dots(3.1)$$

where K is an empirical "evaporation factor" arrived at by comparing the long term annual ocean evaporation, e_w is the vapour pressure at the sea surface, e_a is the vapour pressure at height 'a' above the sea surface and w_a is the wind speed at height 'a'.

3.5 The first measurement of evaporation from pans on board ship was made by Mohn (1883) in 1876-78. He used a pan that floated in a large container and determined the evaporation in 12 hours. In later experiments, mostly between the years 1908 and 1914, evaporation from oceans has been determined by measuring the increase in salinity during the period of exposure.

3.6 Wüst in 1920 showed that evaporation from pans on board ship depends basically on wind velocity w and evaporation potential ϕ . The latter is defined as

$$\phi = \frac{T}{273} (e_s - e) \quad \dots(3.2)$$

where T is absolute temperature of water in the pan, e_s the vapour pressure at water surface and e the vapour pressure in the air measured on board ship. The vapour pressure e_s is given by

$$e_s = e_d(1 - 0.0053S) \quad \dots(3.3)$$

where e_d is vapour pressure over distilled water and S the salinity in parts per thousand. It should be noted that e_s is somewhat lower than e_d . The evaporation E in cm per 24 hours could then be represented by

$$E = 0.40 \phi (1 + 0.40w) \quad \dots(3.4)$$

e , which comes in term ϕ is measured in millibars and w in metres per second. Wüst claimed that values of evaporation thus derived have an accuracy of $\pm 8\%$. But later, Schmidt (1916) showed that the evaporation from sea surface was probably only half of that from the pans. This matter has been discussed in great detail by Wüst (1920, 1935) and Cherubim (1931).

3.7 From pan observations at sea Wüst has derived average values of the evaporation from the different oceans in different latitudes (Table I). Similar annual values can also be computed by means of energy equation assuming that the net transport of heat by ocean current can be neglected. Comparison of these computed values with the observations shows that the compu-

ted values are higher at the lower latitudes and lower at the higher latitudes. This indicates that in low latitudes, part of the radiation surplus is stored in water, is carried north by ocean currents and is used for evaporation in higher latitudes. According to Jacobs the energy transported by ocean currents across 30°N is 1.4×10^{16} cal min⁻¹.

4. Determination of Evaporation

There are several approaches to the problem of determination of evaporation in the atmosphere. Methods based on the following have been tried so far:--

(i) Formulae based on meteorological elements

(ii) Theoretical methods

This is approached in different ways by different workers in the field. The first method concentrates attention on the mechanism of the removal of vapour by diffusion and is primarily applicable to the determination of local rates of evaporation. We shall also include in this the approach by Thornthwaite and Holzman which is the best known and widely used formula. The second approach relies on estimating the amount of energy used in the change from the liquid to vapour phase and hence the rate at which water is being removed.

(iii) Correlation with climatic records

(iv) Direct measurement from pans

5. Formula based on Meteorological Elements

5.1 Observations of evaporation from lakes, reservoirs and pans have been used in the development of many empirical formulae in which evaporation is expressed as a function of various meteorological elements such as temperature, relative humidity, barometric pressure, wind velocity and solar radiation. Dalton (1802) was the first to point out that evaporation is proportional to the difference between vapour pressure of the air at the water surface and that of the overlying air, although apparently he never expressed this relationship in mathematical terms.

5.2 In Rohwer's "Evaporation from free water surface", published in 1931, a number of evaporation formulae are presented and discussed. Most of these formulae contain the expression $(e_s - e_d)$, e_s being the vapour pressure at the surface of the liquid and e_d the vapour pressure of the air, and a factor w expressing the influence of wind velocity. Rohwer's empirical formula is:

$$E = (0.44 + 0.118w) (e_s - e_d)$$

5.3 A few other empirical formulae have already been discussed in the previous section. Thornthwaite and Holzman (1939) have carried out experiments on theoretical and observational procedures which have given promising results. In their method they measure moisture at two or more levels using theoretical developments of Prandtl (1932, 1935), Rossby (1932, 1935) and Sverdrup (1936). From their formula evaporation can be determined by knowing the vapour pressures and wind velocities at two levels. We shall be referring to this again in theoretical methods.

6. Theoretical Methods - (i) Evaporation as a Diffusion Process

6.1 Evaporation from smooth and rough surfaces

6.1.1 The moisture concentration of air can be represented by ρq , where ρ is the density of air and q the specific humidity. This will be considered as a conservative property, that is, a property which is altered locally (except at boundaries) by processes of diffusion and advection only, and we may write

$$\begin{aligned} \frac{\partial(\rho q)}{\partial t} &= \frac{\partial}{\partial x} \left(\frac{A_x}{\rho} \frac{\partial(\rho q)}{\partial x} \right) + \frac{\partial}{\partial y} \left(\frac{A_y}{\rho} \frac{\partial(\rho q)}{\partial y} \right) \\ &+ \frac{\partial}{\partial z} \left(\frac{A_z}{\rho} \frac{\partial(\rho q)}{\partial z} \right) \\ &- \frac{\partial}{\partial x} (\rho q w_x) - \frac{\partial}{\partial y} (\rho q w_y) - \frac{\partial}{\partial z} (\rho q w_z). \end{aligned} \quad \dots(6.1)$$

Here $\frac{A_x}{\rho}$, $\frac{A_y}{\rho}$ and $\frac{A_z}{\rho}$ represent the diffusion coefficients (of dimensions $L^2 T^{-1}$) which are supposed to be different in different directions and w_x , w_y and w_z are velocity components. This definition implies that the following considerations are not valid if droplets are present in such numbers that condensation on or evaporation from droplets, cannot be neglected. Assuming stationary conditions ($\frac{\partial(\rho q)}{\partial t} = 0$), motion along x - axis only ($w_y = w_z = 0$) and neglecting horizontal diffusion ($A_x \frac{\partial(\rho q)}{\partial x} = A_y \frac{\partial(\rho q)}{\partial y} = 0$), we can rewrite equation (6.1) as

$$\frac{\partial}{\partial z} \left(\frac{A_z}{\rho} \frac{\partial(\rho q)}{\partial z} \right) = \frac{\partial(\rho q w_x)}{\partial x} \quad \dots(6.2)$$

Equation (6.1) can also be applied to evaporation from ocean surface which is a water surface of infinite extension, directly above which horizontal gradients of moisture content are negligible. Very near the sea surface the ver-

tical velocity can be neglected and density can be considered constant. Then eqn. (6.1) reduces to

$$\frac{d}{dz} \left(A \frac{dq}{dz} \right) = 0$$

or

$$A \frac{dq}{dz} = \text{const.} \quad \dots(6.3)$$

This expresses that near the boundary surface the vertical flux of water vapour is independent of height. If the vertical flux is directed upwards, $\frac{dq}{dz}$ must be negative. In this case the flux must equal the evaporation from water surface and, therefore evaporation E is given by

$$E = -A \frac{dq}{dz} (g \text{ cm}^{-2} \text{ Sec}^{-1}). \quad \dots(6.4)$$

This is the general form of an equation for evaporation. This shows that the problem of evaporation is the problem of finding A , the eddy diffusivity. At some distance from surface we may write

$$A \approx \rho k_0 w_* z \quad \dots(6.5)$$

Defining for convenience a term Γ , (Montgomery, 1940),

$$\Gamma \equiv - \frac{1}{q_s - q} \frac{dq}{d \ln z} \quad \dots(6.6)$$

(where q_s and q are the specific humidities at the surface and at height z respectively) and putting (6.5) and (6.6) in (6.4) we get

$$E = k_0 \rho w_* \Gamma (q_s - q). \quad \dots(6.7)$$

Γ is called the evaporation coefficient, k_0 von Kármán's universal turbulence constant ($k_0 = 0.4$) and w_* the "friction velocity" which is defined by

$$w_* = \sqrt{\tau / \rho} \quad \dots(6.8)$$

where τ is shearing stress.

In equation (6.7) we shall determine the evaporation coefficient Γ , which is the unknown factor, for smooth and rough surfaces.

6.2 Evaporation from smooth surface

Over a smooth surface, there is a thin boundary layer in which the flow is laminar. Above this laminar layer the turbulent layer begins. Placing the origin of the vertical axis at the top of the laminar layer we have in the turbulent layer.

$$A = \rho (\nu + k_0 w_* z) \quad \dots(6.9)$$

and in the laminar layer

$$A = \mu = \rho \nu \quad \dots(6.10)$$

here μ is the viscosity and ν kinematic viscosity of air.

Shearing stress τ is defined as

$$\tau = A \frac{dw}{dz} \quad \dots(6.11)$$

It is supposed to be independent of height, near the sea surface and equal to

τ_0 , the wind stress at the sea surface. The friction velocity can be

expressed by the wind at any level, w , as

$$w_* = Yw \quad \dots(6.12)$$

where Y is the resistance coefficient which according to von Kármán, can be

obtained from the equation

$$\frac{1}{Y} + \frac{1}{k_0} \ln \frac{1}{Y} = 5.5 + \frac{1}{k_0} \ln \frac{w_* z}{\nu} \quad \dots(6.13)$$

Further, from dimensional consideration, Montgomery (1940) expresses the

thickness of the laminar layer by

$$\delta = \lambda \frac{\nu}{w_*}$$

where λ is a constant for which von Kármán found the value 11.5 and Montgomery obtained 7.8.

The above picture of the variation with height of the eddy viscosity is fully applicable in the case of eddy diffusivity also. We may expect close to the sea surface a layer through which the flux of water vapour takes place by molecular diffusion, and the thickness of which is equal to that of the laminar layer. The eddy diffusivity in the turbulent layer may, therefore, be written as

$$A' = \rho(k + k_0 w_* z), \quad \dots(6.14)$$

where k is the coefficient of diffusion of water vapour through air. Where $z \gg \delta$, eddy diffusivity differs only imperceptibly from the eddy viscosity.

The flux of water vapour through the two layers is according to equation (6.4)

$$E = -\rho k \left(\frac{dq}{dz} \right)_l = -\rho(k + k_0 w_* z) \left(\frac{dq}{dz} \right)_t \quad \dots(6.15)$$

where the subscripts l and t denote that the differentials apply to the laminar and the turbulent layer, respectively. Integrating we get ($z = 0$ at $q = q_0$ at the top of the laminar layer)

$$q_s - q_0 = \frac{E}{\rho k} \delta = \frac{E}{\rho w_*} \frac{\lambda \nu}{k} \quad \dots(6.16)$$

$$q_0 - q = \frac{E}{k_0 \rho w_*} \ln \frac{k + k_0 w_* z}{k} \quad \dots(6.17)$$

For $z \gg \delta$, k is small compared to $k_0 w_* z$ and we may rewrite equation (6.17) as

$$q_0 - q = \frac{E}{k_0 \rho w_*} \ln \frac{k_0 w_* z}{k} \quad \dots(6.18)$$

Adding (6.16) and (6.18) we get

$$q_s - q = \frac{E}{k_0 \rho w_*} \left[\lambda k_0 \frac{\nu}{k} + \ln \frac{k_0 w_* z}{k} \right] \quad \dots(6.19)$$

Rearranging we get

$$E = k_0 \rho w_* \frac{q_s - q}{\lambda k_0 \frac{\nu}{k} + \ln \frac{k_0 w_* z}{k}} \quad \dots(6.20)$$

Substituting w from equation (6.12)

$$E = k_0 \rho Y w \frac{q_s - q}{\lambda k_0 \frac{\nu}{k} + \ln \frac{k_0 w_* z}{k}} \quad \dots(6.21)$$

Hence from equation (6.7) we get

$$\Gamma_a = \left[\lambda k_0 \frac{\nu}{k} + \ln \frac{k_0 w_* z}{k} \right]^{-1} \quad \dots(6.22)$$

Here $\frac{\nu}{k}$ is independent of temperature ($\nu/k = 0.602$), but k itself increases a little with temperature (at 0°C $k = 0.22$, at 20°C $k = 0.25 \text{ cm}^2 \text{ sec}^{-1}$).

With $k = 0.24 \text{ cm}^2 \text{ sec}^{-1}$ and $a = 800$, Γ_a is shown in Fig. 6 as a function of w .

According to Rossby (1936), the sea surface has the character of hydrodynamically smooth surface at wind velocities upto $6-7 \text{ m sec}^{-1}$ as measured at a height of about 8 m. At wind velocities exceeding $7-8 \text{ m sec}^{-1}$ the sea surface appears to be hydrodynamically rough.

6.3 Evaporation from Rough Surface

6.3.1 Various attempts have been made to describe evaporation from rough surfaces. Each worker in the field has assumed different character as to the nature of diffusivity close to the surface.

6.3.2 Millar (1937) and Montgomery (1940) have assumed that next to the surface there is a true diffusion layer followed by an intermediate layer of thickness Z in which the eddy diffusivity corresponds to that over a smooth surface. At Z the eddy diffusivity suddenly increases to the

value of the eddy viscosity over a rough surface, and for $z > z_0$,

$$A = k_0 \rho w_{*z} (z + z_0) \text{ where } z_0 \text{ is the roughness length of the surface.}$$

6.3.3 The second assumption is that next to the surface there is a true diffusion layer of thickness $\delta = \frac{\lambda y}{w_{*z}}$, where w_{*z} applies to a rough surface. From the top of this layer the eddy diffusivity increases at the same rate as the eddy viscosity above a rough surface. These assumptions are similar to those of Bunker and others (1949).

6.3.4 While studying evaporation from oceans Sverdrup (1937) assumed that next to the surface there is a true diffusion layer of thickness $\delta = \frac{\lambda y}{w_{*z}}$ where w_{*z} applies to a rough surface. At the top of this layer the diffusivity increases abruptly from the molecular value to the value over a rough surface $A = k_0 \rho w_{*z} (\delta + z_0)$ and for $z > \delta$, $A = k_0 \rho w_{*z} (z + z_0)$. In another paper Sverdrup (1946) came to the conclusion that there does not exist a true diffusion layer. The eddy diffusivity is at all levels identical with the eddy viscosity i.e. $A = k_0 \rho w_{*z} (z + z_0)$.

6.3.5 Norris (1948) while agreeing with Millar and Montgomery on the existence of laminar layer, the intermediate layer and the outer turbulent layer, assumed that at the top of the intermediate layer, at $z = Z$, the vertical flux of water vapour increases abruptly from E_0 to E where $E_0/Z = (w_{*0} / w_{*z})^2$.

6.3.6 On the basis of the above assumptions we arrive at five different equations to express Γ_a . These are

$$1. \Gamma_a = \left[\ln \frac{z}{z_0} + \frac{w_{*z}}{w_{*0}} \left(\lambda k_0 \frac{y}{k} + \ln \frac{k_0 w_{*z} z}{k} \right) \right]^{-1}, \dots (6.23)$$

$$2. \Gamma_a = \left(\lambda k_0 \frac{y}{k} + \ln \frac{k_0 w_{*z} z}{k} \right)^{-1}, \dots (6.24)$$

$$3. \Gamma_a = \left(\lambda k_0 \frac{y}{k} + \ln \frac{z}{\delta + z_0} \right)^{-1}, \dots (6.25)$$

$$4. \Gamma_a = \left(\ln \frac{z}{z_0} \right)^{-1}, \dots (6.26)$$

$$5. \Gamma_a = \left(\ln \frac{z}{z_0} - 1.76 \frac{w_{*z}}{w_{*0}} \right)^{-1}, \dots (6.27)$$

6.3.7 In Fig. 6 five Γ_a curves for $a = 800 \text{ cm}$ are shown as functions of the wind velocity, corresponding to five sets of assumptions. From Montgomery's derivation we find that the evaporation coefficient for a fully rough surface

is one half of that for smooth surface, but since the friction velocity increases in passing from a smooth to a rough surface, it is concluded that evaporation from a smooth sea is much the same as that from rough sea, except for the contribution from spray. This, as we find, does not agree with the work of Norris (1948), who assumed that the ratio of the flux of vapour in the intermediate and turbulent layers is proportional to the ratio of the shearing stress in these layers, and therefore to the square of the ratio of the friction velocities appropriate to the two layers. On this basis, Norris concluded that the ratio of evaporation from a fully rough surface is four times that for a smooth surface.

6.3.8 On the whole, the weight of observational evidence is in favour of the theories of Norris and Sverdrup and against Montgomery. For winds (at 6 m) less than 800 cm sec^{-1} , all three theories lead to much the same result and are in moderate agreement with observation. For winds exceeding 800 m sec^{-1} Montgomery's formula gives rates of evaporation far below those observed, whereas there is a measure of agreement with the estimates of Sverdrup and Norris.

7. Theoretical Methods — (ii) Energy Balance Method

7.1 This method is based on the principle of conservation of energy. This consists in equating the energy received from the sun to the energy used by evaporation, warming, back radiation etc. Energy balance in the case of ocean can be written as

$$Q_0 = Q_z + Q_b + Q_h \dots (7.1)$$

where

Q_0 = Shortwave radiation from sun and sky.

Q_z = Upward flux of long-wave radiation.

Q_b = Heat used in evaporation.

Q_h = Loss of heat by conduction to the atmosphere.

All the above quantities are expressed in $\text{g cal cm}^{-2} \text{min}^{-1}$. It is assumed that other sources of heat gains or losses, such as conduction of heat from interior of the earth, energy changes related to the chemical and bio-chemical processes in sea and friction losses, are negligible and that average temperature of the oceans remains nearly constant from one year to the other.

Introducing Bowen Ratio, $R = \frac{Q_h}{Q_b}$ equation (7.1) becomes

$$Q_0 = \frac{Q_b - Q_z}{1 + R} \dots (7.2)$$

Now $Q_0 = LE$, where L is the latent heat of vapourisation.

Hence the rate of evaporation per unit area is

$$E = \frac{Q_0 - Q_r}{L(1+R)} \quad \dots(7.3)$$

The problem here is to determine R . The other quantity $(Q_0 - Q_r)$, radiation surplus, is known or can be measured easily. Schmidt determined the value of $(Q_0 - Q_r)$ by using measurements of radiation at sea combined with data pertaining to cloudiness and sea surface temperature (Fig. 7). Mosby (1936) computed it by using certain empirical relations between the incoming radiation and the altitude of the sun. His values agree well with those of Schmidt (Fig.7) except between latitudes $20^\circ N$ and $20^\circ S$ where Schmidt's results show a minimum radiation surplus near the equator whereas Mosby obtains no such minimum. This difference is due to the fact that Schmidt has introduced a considerably greater cloudiness at the equator than at $30^\circ N$ or $30^\circ S$ (5.9, 4.2 and 4.0 respectively) whereas Mosby has used nearly the same values of cloudiness in all latitudes between $30^\circ N$ and $30^\circ S$ (values between 5.6 and 5.2). McEwan (1938) has also computed the radiation surplus between latitudes 20° and $50^\circ N$ in the eastern North Pacific. His method is very elaborate one. The agreement between his result and those of Schmidt and Mosby is quite good (Fig. 7).

7.2 According to Bowen R can be computed if temperature T_a and the vapour pressure e_a at a height 'a' above the sea surface is known and if the temperature at the sea surface T_s has been recorded. The vapour pressure at the sea surface is obtained by using the equation.

$$e_s = e_a(1-0.00053S) \quad \dots(7.4)$$

where S is the salinity in parts per thousand,

Bowen's formula can be derived in a simple manner. If the eddy coefficients for diffusion of water vapour and conduction of heat are assumed to be same, the upward fluxes of latent energy of water vapour and of heat can be written as

$$Q_0 = - \frac{L \times 0.621}{p} \times A \frac{dq}{dz} \quad \dots(7.5)$$

and

$$Q_h = - C_p A \frac{dT}{dz} \quad \dots(7.6)$$

where p is the atmospheric pressure, A the eddy conductivity (or diffusivity), C_p the specific heat of air and T the air temperature. To be more appropriate, potential temperature should be used, but near the sea surface, due to large vertical temperature gradient, only a small error is introduced by using the ordinary temperature. It follows that

$$R = \frac{Q_0}{Q_h} = \frac{C_p p}{0.621 L} \frac{dT/dz}{dq/dz} \quad \dots(7.7)$$

Replacing $\frac{dT/dz}{dq/dz}$ by $(T_s - T_a) / (e_s - e_a)$ and putting $C_p = 0.245$, $p = 1000$ mb and $L = 585$, one obtains R , the Bowen Ratio as

$$R = 0.64 \frac{T_s - T_a}{e_s - e_a} \quad \dots(7.8)$$

7.3 The Bowen ratio has been used extensively by Cawings and Richardson (1927) in the study of evaporation from lakes and by Jacobs (1942) in his examination of energy exchange between the ocean and the atmosphere.

7.4 Jacobs determined the Bowen ratio as a function of latitude for North Atlantic and North Pacific Oceans. Though in both cases R increases with latitude, the values found for the North Atlantic were consistently smaller than those of the North Pacific. This was due to the fact that careful observations from specially equipped ships have shown that in the tropics of the Atlantic Ocean, the air temperature is about $0.8^\circ C$ lower than the sea surface temperature, but the values found in climatological charts show a smaller difference or even higher air temperature and Jacobs had used these values and therefore his R values are probably too small in low latitudes which require a correction.

7.5 We give below the average values of R for the two northern oceans.

Table II

N. Lat. (deg.)	R (Jacobs)	R (Corrected)	R (in intervals of latitude)
70	0.53	0.53	0.45
60	0.37	0.37	0.31
50	0.25	0.25	0.21
40	0.18	0.18	0.15
30	0.08	0.13	0.11
20	0.02	0.10	0.10
10	0.00	0.10	0.10
0	0.00	0.10	0.10

8. Theoretical Methods - (iii) Method of Thornthwaite and Holzman

8.1 Jeffreys (1918) and Giblett (1921) were among the earliest investigators who studied the problem of evaporation in terms of atmospheric turbulence. However, because of various non-rigorous assumptions regarding the nature of the Austausch coefficient, their results are limited and can be applied only to the evaporation from bodies of water. Sutton (1934), making use of Taylor's (1922) researches, has extended Jeffreys's analysis by assuming that the Austausch coefficient varies with height. His theoretical work was found to be in very good agreement with experimental evaporation measurements for variously shaped areas. Schmidt (1935) has made use of a formula devised by Ertel (1930) for calculating the Austausch coefficient and claims to have measured the actual evaporation from a meadow.

8.2 With the help of researches on turbulence by von Kármán (1930, 1934, 1935, 1937) and Rossby (1932, 1935), Thornthwaite and Holzman (1939) formulated a method for determining evaporation from observations of moisture concentration and wind velocity in the turbulent layer. The derivation of the formula follows directly from an expression for the Austausch coefficient obtained from the concepts of the mixing length and the shearing stress as developed by Prandtl (1932, 1935) and von Kármán. The formula is:

$$E = \frac{k_0^2 \rho (q_1 - q_2) w_2}{\ln \frac{z_2}{z_1} \ln \frac{z_2}{z_0}} \quad \dots(8.1)$$

in which

- k_0 = von Kármán's coefficient
- ρ = density of air
- q_1 and q_2 = moisture concentration at lower and upper levels respectively
- w_2 = wind velocity at upper levels
- z_2 and z_1 = height of upper and lower instruments respectively
- z_0 = roughness coefficient

The roughness coefficient is determined from observations of wind velocity at two levels by means of the following formula

$$\ln z_0 = \frac{w_2 \ln z_1 - w_1 \ln z_2}{w_2 - w_1} \quad \dots(8.2)$$

Since the intensity of turbulent mixing is dependent on the wind velocity and the roughness coefficient and since the latter can be determined from wind velocities at two levels, it is possible to simplify the evaporation formula to the following:

$$E = \frac{k_0^2 \rho (q_1 - q_2) (w_2 - w_1)}{(\ln \frac{z_2}{z_1})^2} \quad \dots(8.3)$$

8.4 The formula giving evaporation in inches per hour for an installation where the upper observations are taken at 28.6 ft. and the lower ones at 2 ft. above the ground is

$$E = \frac{0.0274 P (q_1 - q_2) (w_2 - w_1)}{T + 459.4} \quad \dots(8.4)$$

where P is pressure in inches of mercury, q_1 and q_2 are expressed in gm per kg, the wind velocity 'w' in miles per hour and temperature T in °F. If expressed in terms of vapour pressure the formula is simplified to

$$E = \frac{17.1 (e_1 - e_2) (w_2 - w_1)}{T + 459.4} \quad \dots(8.5)$$

where e_1 and e_2 are vapour pressure in inches of mercury at lower and upper levels respectively.

The formula is rigorously correct for an adiabatic atmosphere. Some corrections are made for other conditions but these are minor except when inversions are strong or when adiabatic lapse rate is considerably exceeded.

9. Correlation with Climatic Records

9.1 This is another method by which evaporation from large areas can be determined, provided complete and accurate records are available. Thus, if total precipitation and runoff by surface and ground water of a place and also the change in the storage in the area are known, the difference will give the total amount of evaporation taking place from soil, free water surface, snow surface and vegetation. This method has been applied with some success by the United States Geological Survey.

9.2 In his method Thornthwaite (1943) obtained all available observations of water losses from land areas in different parts of U.S.A. These include water requirements in various irrigation districts in the western region of

the country and the data on evapotranspiration from different types of cover on the weighing lysimeters of the Soil Conservation Service in Coshocton, Ohio. They also include computations of water loss from small watersheds from values of precipitation and runoff for months when it was apparent that there was no reduction of evapotranspiration because of a moisture deficiency in the soil and where it was possible to estimate the lag in the runoff and thus to assign the water available for runoff to the months when the rainfall occurred.

9.3 From these observations Thornthwaite derived a general equation by means of which daily, monthly and annual evapotranspiration can be determined from records of temperature, length of day and precipitation since runoff can then be computed as a difference between precipitation and evapotranspiration.

10. Direct Measurement from Pans

10.1 This method is most widely used for evaporation measurements. Although direct measurements from pans have limitations on theoretical grounds, it has, nevertheless, advantages in some applications (Kohler, 1955). A variety of pans and special surfaces have been designed for this purpose.

10.2 There are three types of exposure employed for pans installations—sunken, above ground and floating. Divergent views persist as to the best exposure. Some important advantages and disadvantages of these are given below:

- (i) **Floating Pans:** The evaporation from a pan floating in a lake more nearly approximates evaporation from the lake itself. But it is influenced by the lake in which it is immersed and therefore it is not necessarily a good 'climatic' indicator. Observational difficulties are the chief disadvantages of floating pans and splashing frequently renders the data unreliable.
- (ii) **Sunken Pans:** Burying the pans tends to eliminate objectionable boundary effects, such as radiation on the side-wall and heat exchange between the atmosphere and the pan proper. But there are some operational problems. These pans collect more trash, are difficult to clean and leaks cannot be easily detected and rectified. Height of vegetation adjacent to the pan is also quite critical. Moreover, appreciable heat exchange does take

place between the pan and soil depending on soil type, moisture content and vegetation cover. It is advisable to use a large pan to reduce these effects relatively.

- (iii) **Surface Pans:** These types of pan experience greater evaporation than sunken pans because of added radiant energy intercepted by the side walls. Kohler (1955) in his studies has indicated that this factor has no adverse effect although heat transfer at the pan/air interface does introduce geographical (climaticological) variations. The advantage in use of this type of pan, is the ease of operation and maintenance.

10.3 Evaporimeters in Use

- (a) **Open tank evaporimeter:** This is a cylindrical tank 4 ft. in diameter and 10 inches deep. It is set on timbers with the bottom 4 inches from the ground. This permits air to circulate beneath it. It is questionable just how much such an arrangement reduces the lag in the diurnal changes of the water temperature as compared with air temperature. Since the mass of water in the tank is relatively large, this lag may be considerable. There may also be a rim effect because of variations in the roughness of the surrounding ground.
- (b) **Piche Evaporimeter:** This consists of graduated glass tube open at one end and closed at the other. The tube is filled with water and a porous disc placed over the open end. The tube is then inverted. The porous disc is wet as long as water is in the tube. Evaporation is measured by noting the drop of the water level in the graduated tube. The whole apparatus is placed inside the Stevenson Screen. For this reason, it responds to vapour pressure deficit and variation in wind speed but not to variation in solar or sky energy. In dry regions it has been found that the porous disc dries around the edges and so evaporating surface is not always of the same area. Another difficulty is that the tube barely holds enough water for one day's evaporation. Because of the shape of the disc and the manner of mounting it in contact with the open end of the glass tube, it is difficult to standardise the size and effectiveness of the evaporating surface.

(c) **Bellani Plate Anemometer:** This consists of essentially a porcelain flask, the walls of which are water proofed and the flat part porous. This porous plate is about 7.5 cm in diameter. The flask is operated in an inverted position so that the flat part is exposed to the sky horizontally. Water is supplied to the flask from a reservoir. By means of a special valve, water can flow to the flask but not away from it as would be the case when rain or dew falls on the porous plate. The mass of the instrument is small so that its temperature follows fluctuations of air temperature fairly closely. Since there is no protruding rim, there can be no obstruction to wind flow over the plate. The instrument can be placed at any desired height. Thus it is possible to place it at the optimum level for ease of operation and away from the influence of wind speed due to variations in surface roughness.

11. Conclusion

11.1 We have seen that the study of evaporation has presented both theoretical and practical difficulties. The technique for evaluating evaporation either by direct measurement of outgoing moisture or by computation from energy balance, turbulent transfer or any other theory is still inexact.

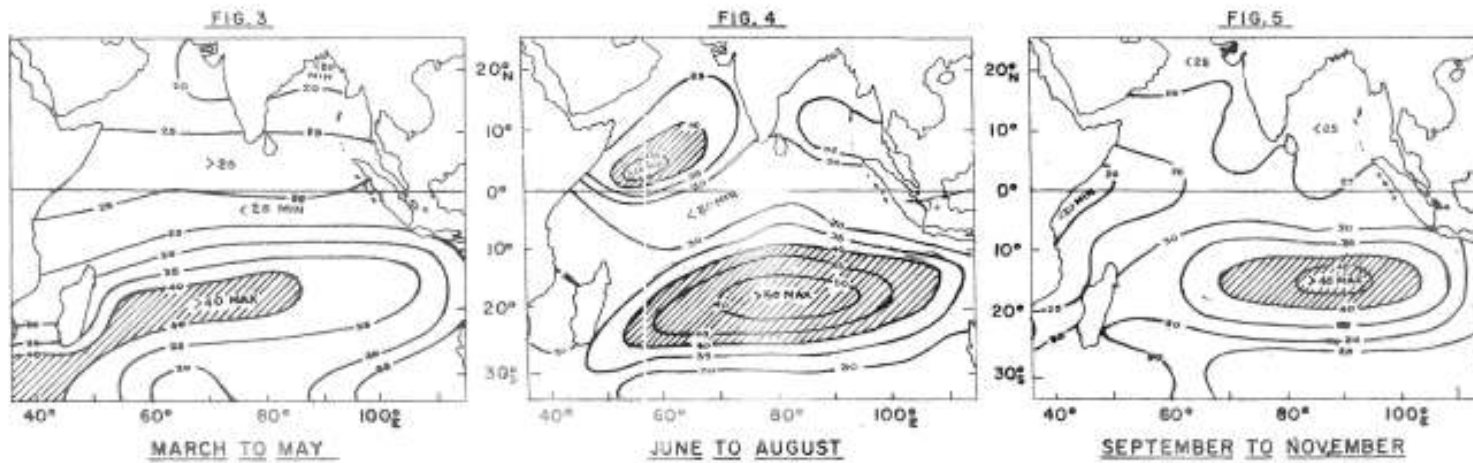
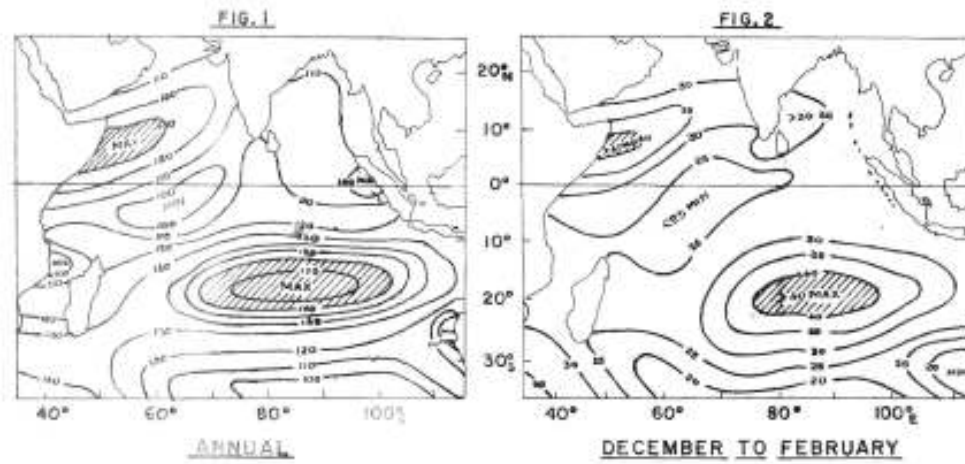
11.2 Even though the evaporation pan is a very old tool, it will, probably for many years to come, continue to be the only cheap and generally accepted instrument. Many experimental studies have been conducted in the past either in nature (Rohwer 1933, 1934, Hickox 1946, Kohler 1952, Fitzpatrick 1963, Fritschen and van Bavel 1963) or in wind tunnels (Hiras 1929, Powell and Griffiths 1935, Miller 1937, Shepherd, et. al. 1938, Pasquill 1943, Yamamoto, 1950), but in spite of this large amount of data there is no evidence in the literature that the phenomenon of evaporation from pan is completely understood.

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ANNUAL AND SEASONAL EVAPORATION (CM.) OVER THE INDIAN OCEAN



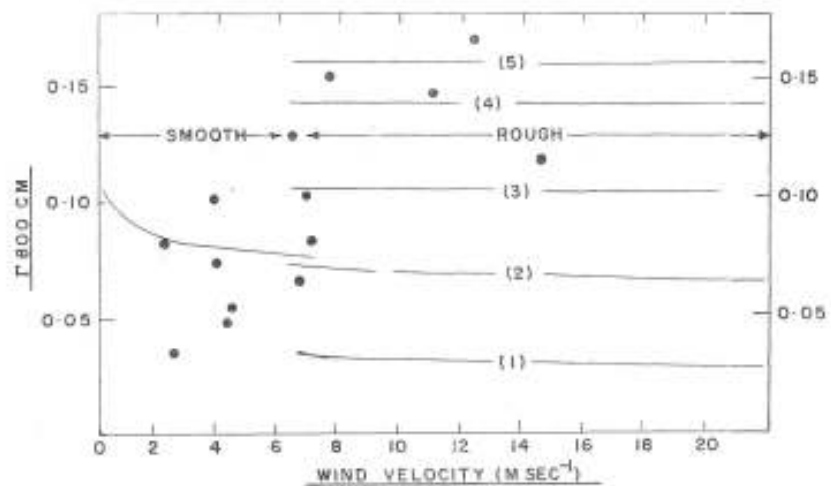


FIG. 6: Theoretical values of the evaporation coefficient as function of the wind velocity at 800 cm, computed on the basis of assumptions; observed values are shown by dots.
(from Compendium of Meteorology Pg.1077)

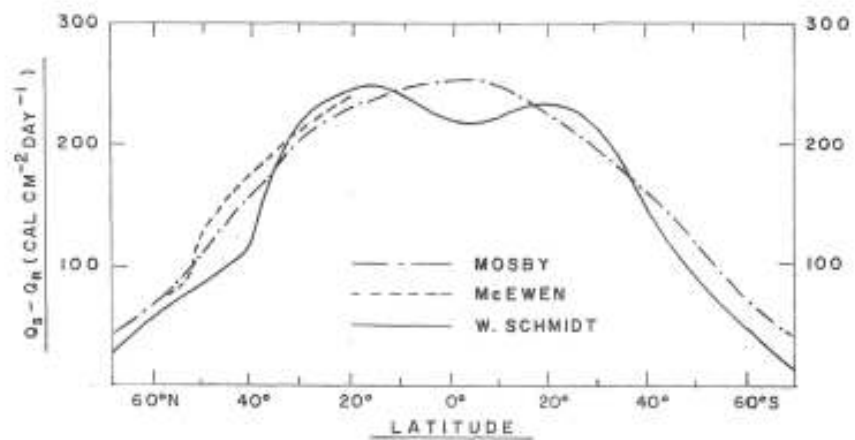


FIG. 7: Average annual surplus of energy which the oceans receive in different latitudes by radiation processes (from Compendium of Meteorology page 1073)