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SYSTEMATIC PROCEDURE FOR THE COMPONENTS OF WATER BALANCE COMPUTATION OF THE LAKES - PART-I EVAPORATION

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ABSTRACT

Water balance of a lake depends upon the relative importance of various imputs and outputs to and from the lake. Evaporation loss constitutes one of the major output from a lake under Indian condition and as such, the water regime of a lake and yield from it are influenced by the evaporation under Indian conditions. There are a plethora of formulae and methods to estimate evaporation from a free water surface as in a lake and it is very difficult to adopt a particular method universally and more so under Indian condition. Study on hydrological aspect of lake in India has started some time back and the development of most suitable methodology for lakes in a region will need lot of experimentation and will take time to identify. In order to conserve and use our valuable water in our lakes for the community, there is an urgent need to have a first hand reasonably accurate estimate of water balance of lakes. As evaporation constitute the major output component of Indian lakes, an attempt has been made herein to review critically the major and important formulae available and their suitability for appropriate adoption and adaptation under Indian condition are examined with respect to objective of the study and desired level of accuracy.

Meyer Equation and Class A pan data multiplied by 0.7 are likely to give acceptable estimates of evaporation for design purpose. The equation which requires simple three hydrometeorological inputs (air temperature, water temperature and Wind Velocity at 10 meter height) even could be used for operational purpose if reliable measurements of air and water temperatures are made.

The energy-budget method is suitable for research purpose where very accurate estimate of evaporation is warranted. It can be resorted to for small areas only as the requirements for detailed meteorological data are great. The practical utility of the method for large lakes is limited.

1.0 Introduction:

The water balance equation for lakes for any time interval is a continuity equation. The water volume of a lake changes with time though the water exchange in lakes is much slower than in rivers. According to the law of conservation of matter, there is equilibrium between inflow components, outflow components and the change of water volume for each interval of time. Water balance is a description of this equilibrium and this forms the basis for the rational, deterministic hydrological forecasting models and are necessary for (UNESCO, 1981):

i) forecasts of lake levels for shoreline, property utilization and navigation,

ii) in the design, selection and operation of forecasting models,

iii) predicting environmental impacts i.e. preservation of living resources of a lake through the maintenance of water quality standards,

iv) to obtain valuable information base for effective management,

v) help in global studies of climate variability.

As such, water balance studies of lakes are of vital importance for the proper use, conservation and harvesting of lake water Following aspects should be taken care of while computing water balance of a lake:

i) water inflow from rivers and drainage areas along the whole length of the water body perimeter

ii) peculiarities in the formation and fall out of precipitation on the water surface

iii)water outflow to a river

iv) Peculiarities of water losses through evaporation from the water surface

v) Peculiarities in hydrology of watersheds.

2.0 Water balance equation for lakes:

A lake represents a simple open system with respect to the mass balance of the water itself. A lake has an upper water surface exposed to the atmosphere and a lower surface boundary in contact with a solid mineral surface. Water may enter and leave through both of these boundary surfaces. Incoming streams and overland flow from ground surfaces draining into the lake represent point and line sources of water input, while an overflow channel represents an output point. Consequently, a fairly simple water balance equation can be set up (Fig. 1) as follows:

$$\Delta S = \underbrace{(I_r + I_p + I_g)}_{\text{Input}} - \underbrace{(O_r + O_e + O_g)}_{\text{Output}} \qquad ....(1)$$

Where,

ΔS = net change in storage of water in the lake,

I_r = surface inflow

I_p = direct precipitation upon the lake surface

I_g = groundwater inflow to lake

O_r = surface outflow

O_e = evaporation from lake water surface

O_g = groundwater outflow

The above equation (1) based on inflow equals outflow plus or minus changes in storage is commonly used. But it can only serve the purpose adequately if all the components of water balance including groundwater are measured accurately. The relative magnitude of the water balance components vary from place to place and season to season. In fact, of all the components included in the balance equation, surface run off is the only one that have ready direct measurement or estimation, since it is confined within well defined geometric boundaries that permit determination

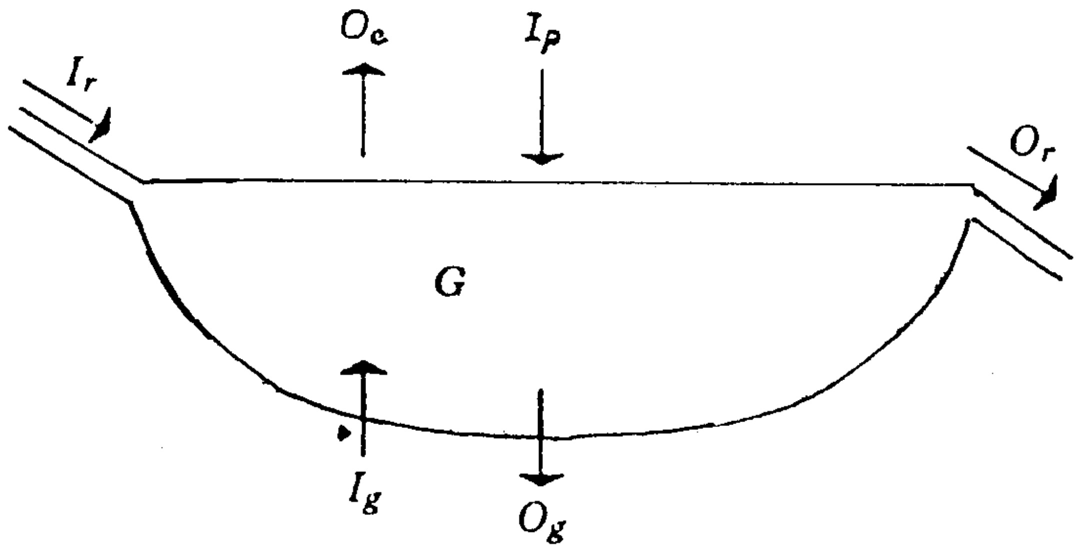

Fig. 1 - Schematic diagram of the water balance of a lake

of both rate and cross-sectional area of flow. Evaporation, out of all the components of the water balance and perhaps out of all the components of the hydrologic cycle, is the most difficult parameter to estimate owing to complex interactions between the components of land -plant atmosphere system. The choice of method used to determine evaporation which is basically an energy exchange process, depends on the required accuracy of results and the type of instrumentation available. The equatorial zone receives relatively constant solar radiation and there is a net radiation gain in this zone. As such, India being very near to the equatorial zone, evaporation from the Indian lakes is of prime importance to the water budgeting and conservation of lake water. Although the subject of evaporation from a free water surface has been studied for last two hundred years or so, the method of measurement and estimation used are still inadequate and more so for Indian lakes. Direct measurement of evaporation are not easily obtained for lakes because of extensive surfaces involved. This report, therefore, is devoted to highlight the systematic procedure to estimate evaporation from the lakes. Other important components of the lake water balance will be dealt in subsequent reports.

3.0 Evaporation:

3.1 Evaporation process

Evaporation is the change of state from a liquid to a gas. The process occurs when water molecules, which are in constant motion have sufficient energy to break through the water surface and escape into the atmosphere. Likewise, some of the water molecules contained in the water vapour in the atmosphere that are also in motion may penetrate the water surface and remain in the liquid. The net exchange of water molecules per unit time at the liquid surface determines the rate of evaporation from a water body. However, continuous evaporation can take place only when there is a supply of energy to provide the latent heat of evaporation (approximately 540 calories per gram of water evaporated at 100° C) and there is some mechanism to remove the vapour, so that the vapour pressure of the water vapour in the moist layer adjacent to the lake water surface is less than the saturated vapour pressure of the surface water of a lake (i.e., a vertical gradient of vapour pressure exists above the lake water surface). The evaporation process is driven by energy that is principally received from solar radiation and also from heat stored within the lake. However, the rate at which evaporation proceeds is a function of the vapor-pressure differential between the water surface and the atmosphere and of various other factors that affect the ability of the atmosphere to exchange water vapour (Houghton,1985).

3.2 Methods of estimating evaporation from lake water surface

The direct measurement of evaporation from a lake specially from a large lake is a difficult task. The methods available for estimating evaporation from a lake are based on a rational understanding of the effects of all the factors affecting the evaporation process. The major factors that affect the evaporation from a lake are (Houghton, 1985),

Heat Supply (Solar Radiation)

Although an evaporating surface can receive energy through the advection of sensible heat (for example, warm water discharging into a lake), the principal source of energy is solar radiation. Increases in radiation will not result directly in increased evaporation but will cause water temperature to increase, thus increasing the vapour-pressure differential and increasing overwater turbulence. Because the evaporative process cools the water surface, a constant supply of energy must be provided to the water surface if evaporation is to continue at a high rate.

Heat Storage

When the amount of energy entering a lake is high, a portion of the energy that might have gone into evaporation will be stored in the water mass, to be returned later when the energy flux in the lake is reversed. For large lakes, the heat storage can modify the monthly distribution of evaporation so much that less evaporation occurs in the summer months than in the winter. Unfortunately, most of the formulas concerning free-water evaporation are not completely applicable to lakes and reservoirs because they do not account for this heat storage factor.

Vapour -Pressure Differential

Assuming that all other factors are constant, the rate of evaporation will be proportional to the difference between the saturation vapour pressure of the water surface and the vapour pressure of the air. This relationship was first recognized by Dalton in 1802 and is often referred to as Dalton's law.

The saturation vapor pressure, e_v (in mb), of water at temperature T (in ^oC) can be found through the equation (Linsley et al. 1975):

$$e_v=33.8639[(0.00738T + 0.8072)-0.000019|1.8T+48|+0.001316 \quad ...(2)$$

However, Table is also available for the values of e_v for different temperatures in ^{o}C. The air-vapour pressure, e_a, can be computed as either the relative humidity times the saturation vapour pressure of the air temperature or as the saturation vapour pressure of the dewpoint temperature.

Wind Speed

The presence of atmospheric turbulence can greatly increase the rate of evaporation. Without wind to remove the evaporated water vapour from the lake surface, the vapour-pressure gradient over the lake will soon decrease, thus reducing future amounts of evaporation. Up to a certain velocity, an increase in wind speed will bring more fresh samples of air to the evaporating surface. Above that value, usually 40 km/hr., fresh air availability no longer becomes a limiting factor, and the evaporation will not continue to increase.

For the lake evaporation estimation techniques which use a wind function, the height at which the wind speed is measured is an important factor. Often, available wind information is taken at a height different from the height for which a particular technique is calibrated. An approximation of the wind speed for a desired height, z_2, may be found from a measurement at another level, z_1, using the formula (Derecki 1975 vide Houghton, 1985),

$$u_2 = u_1 [z_2 / z_1]^{1/7} \quad \ldots(3)$$

3.2 Methods of estimating evaporation

The various methods of estimating lake evaporation are:

1. Mass transfer method
2. Energy budget method

3. Water budget
4. Empirical formulae
5. Measurements from evaporation pans, etc.

Mass Transfer Method

The method is essentially based on the determination of the mass of water vapour transferred from the water surface of a lake to the atmosphere. This approach provides an insight to the physics of the evaporation process. The approach is based on the aerodynamic law first presented in 1802 by John Dalton. Advances in the understanding of the boundary layer problem have provided means to calculate evaporation. The rate at which water molecules leave the water surface is dependent on the temperature of the lake water surface and the atmospheric pressure. Higher water temperatures are synonymous with more vigorous molecular motion and result in more molecules leaving the water surface. On the other hand, an increase in atmospheric pressure inhibits the movement of molecules out of the water. This becomes a significant factor only when differences of elevation of more than a few thousand meters are involved. For most practical problems in hydrology, the variation in atmospheric pressure can be ignored (Bruce & Clark, 1969). Expressing these ideas of aerodynamic principles in algebraic form:

$$E = K\, f(u)\, (e_v - e_a)/p$$

$$E = K\, f(u)\, (e_v - e_a) \qquad \ldots(4)$$

where E= Evaporation in mm/day,

K= Parameter which includes the effects of air density and pressure.

$f(u)$= Empirical coefficient and is a function of the horizontal wind speed at some standard height in m.

$e_w - e_a$ = Vapour pressure difference in which e_w is the saturated vapour pressure at water surface temperature in mm of mercury and e_a is the actual vapour pressure of the overlying air at a specified height in mm of mercury.

With slight modifications of Eq. (4), a simple formula results.

$$E = Kf(u)\ (e_w - e_a)/f(z_o) \qquad \ldots(5)$$

where $f(z_o)$= Roughness parameter which is dependent on the type of the surface (water in case of evaporation).

When air passes over a lake surface, the lower atmosphere could be divided into three layers; viz., the laminar layer (near the surface), turbulent layer and the outer layer of frictional influence. The laminar layer is only a few millimeter in thickness. The temperature, humidity and wind velocity vary nearly linear with height in this layer. The overriding turbulent layer can be of several meters depending on the turbulence. However, in this layer, the temperature, humidity and wind velocity vary approximately linearly with the logarithm of height. Evaporation from a lake could be evaluated if eddy diffusion in the turbulent layer is evaluated. In other words, it becomes a problem related to velocity distribution, vapour distribution and temperature distribution to momentum and/or vapour transfer. The process of eddy diffusion is usually related the wind velocity profile as it is assumed that eddy diffusion of water vapour from a lake is identical to the eddy diffusion of momentum (Gray, 1973)

However, two cases that should be considered Wilson, 1974):

(i) When the temperature of the lake water surface is the same as the air temperature

Such cases will rarely occur and such cases should be empirically treated by the equation:

$$E= C\ f(u)\ (e_v - e_a) \qquad \ldots(6)$$

where E= evaporation in mm/day

C= An empirical constant

e_v and e_a= Saturation and actual vapour pressure of the air at $t^{\circ}C$ in mm of mercury

u=Wind speed at some standard height in m

The following equation has been empirically obtained for this case and is of general validity.

$$E= 0.35\ (e_v - e_a)\ (0.5+ 0.54\ u_2) \qquad \ldots(7)$$

where E is in mm/day, e_v and e_a are in mm of mercury, and u_2 denoting wind speed at 2m height in m/s.

(ii) When the air and lake water surface temperatures are different

This is what normally happens and a formula of the following type should be applicable for such cases.

$$E= C\ (e'_v - e_a)\ f(u) \qquad \ldots(8)$$

However, e'_v is the saturation vapour pressure of the boundary layer of air between air and water, whose temperature t' is not same as either with air or water and is virtually impossible to measure. As such, empirical formulas have been developed in the form of Eq. (6) which works fairly well for specific locations for which the constants are derived. But, such equations do not have any general validity. Such a formula is derived for Ijsselmeer in Holland and is applicable to it and similar condition. The formula is:

$$E = 0.345\ (e_v - e_a)\ (1 + 0.25\ u_6) \qquad \ldots(9)$$

where E is in mm/day, e_v is saturation vapour pressure at temperature t_v of the surface water of the lake in mm of mercury, e_a is actual vapour pressure in mm of mercury, and u_6 is wind velocity in m/s at a height of 6m above the lake surface.

Many variations of evaporation equations of this form have been developed over the years and the equations are known as mass transfer equations. Two of the most widely used mass transfer equations are those given by Sverdrup (1946), and Thornthwaite and Holzman (1939 vide Gray, 1973):

Sverdup -mixing length theory

$$E = [0.623\ \rho\ k_o^2\ u_8\ (e_v - e_8)]\ /\ p\ (\ln 800/z_o)^2 \qquad \ldots(10)$$

$$E = [0.623\ \rho\ k_o^2\ (u_8 - u_2)\ (e_2 - e_8)]\ /\ p\ (\ln 800/200)^2 \qquad \ldots(11)$$

where E= Evaporation ($gm/cm^2/sec$ or cm/sec),

ρ= Density of air (gm/cc),

k_o = Von Karman's constant

u_2 and u_8 = Wind speeds at heights of 2m and 8m respectively (cm/sec),

e_v = Saturated vapour pressure at temperature of water surface (mb),

e_2 and e_8 = Vapour pressures at heights of 2m and 8m respectively (mb),

p= atmospheric pressure, (mb), and

z_0 = roughness parameter (cm) (1 millibar=0.76 mm of Hg)

Complexity of the mass transfer equation may vary from a simple expressions such as given in Eq. (4) to complex relation like Sutton's equation for a circular lake of radius r.

Sutton's equation is:

$$E = \frac{0.623}{p} G\rho u^{(2-n)(2+n)} r^{(4+n)(2+n)} (e_v - e_a) \quad \ldots(12)$$

where E is evaporation in cm/day, ρ is mass density of the air in g/cm^3, u is the average wind velocity in cm/sec, r is the radius of the circular lake in cm, p is the atmospheric pressure in mb, e_v and e_a as previously defined in Eq.(4) and are in mb, n is an empirical constant, and G is a complex function.

Empirical mass transfer equations often require exacting and costly instrumentation, as such, their general utility is limited. However, some of them are reasonably simple equations which have been in use and yielded good results.

A commonly used empirical equation for both shallow and deep lakes was developed by Meyer in 1915 in USA and the equation takes the form (Subramanya, 1994),

$$E = K\,(e_v - e_a)(1 + W/16) \quad \ldots(13)$$

where E is daily evaporation in mm of depth, e_v is the saturated vapour pressure at the water surface temperature in mm of mercury, e_a is actual vapour pressure of overlying air at a specified height in mm of mercury, W is wind velocity in kmph measured about

9m above the water surface, and K is an empirical coefficient accounting for various other factors with a value of 0.36 for large and deep lakes and 0.50 for small shallow lakes and ponds.

Example: Find the daily evaporation E from a lake for a day during which the following mean values are obtained: air temperature 30.5°C , water temperature 17.2°C, wind speed at about 9 m height is 16 kmph, and relative humidity 20%.

Estimation of Lake evaporation (E)

Interpolating from the Table of saturated vapour pressure and temperature, e_v = 14.73 mm of mercury and e_a = 32.77x0.2=6.54 mm of mercury.

Using Eq.(13), 0.36(14.73-6.54)(1+16/16)=5.9 mm/day

This example demonstrates that a relatively simple meteorological data base which we possibly have or can be built-up could be used to have a preliminary estimate of daily evaporation from our lakes to start with.

Investigations about the utility of mass transfer equations conducted at Lake Hefner indicated that a simple equation using wind speed and vapour differences provided good and comparable results. The equation is:

$$E = Nu(e_v - e_a) \qquad \ldots(14)$$

where E is evaporation in cm/day, N is a mass transfer coefficient, u the velocity at 2 m above the water surface in m/sec, e_v and e_a as previously defined in mb.

The value of N can be computed by using an approximation:

$$N = 0.0291/A^{0.05} \qquad \ldots(15)$$

where A is the surface area of the water surface in Sq. meter. However, this equation should be used with caution for values of A less than about 4×10^{6} sq. meter as the variations in wind exposure may become important. An error of about 30% could be there for using this approximation for computing N. When evaluation of N is based on comparative studies of mass transfer and energy budget methods, average error in evaporation estimate of about 15% can be expected. Monthly mean values of evaporation from Lake Ontario for the period 1872-1965 were generated by a mass transfer method by using available onshore recorded meteorological data (Yu and Brutsaert, 1969).

Energy Budget Method

The energy budget method is an application of the law of conservation of energy. The energy available for evaporation is determined by considering the incoming energy, outgoing energy and the energy stored in the water body of a lake over a known time interval. The method has received wide application for the estimation of evaporation from a lake. The energy budget method is considered most accurate and complex for the estimation of the evaporation. The method can estimate annual evaporation within 13% error (Winter, 1981). If the period considered is less than a week, the order of the error could reduce to 5% (Subramanya, 1994). Under good conditions, the average errors of perhaps 10% for summer and 20% for winter months can be expected.

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

$$Q_o = Q_s - Q_r + Q_a - Q_{ar} + Q_v - Q_{bs} - Q_e - Q_h - Q_v \qquad \ldots(16)$$

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

Q_w = energy advected by evaporated water

All the terms are in calories per square centimeter per day (cal/cm^2-day). Heating brought about by chemical changes and biological processes is neglected as is the energy transfer that occurs at the water-ground interface. 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.

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, it has been demonstrated that 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 3-15%

in estimates of monthly evaporation, while errors on the order of 10% in measurements of reflected solar energy may cause errors of only 1-5% in calculated monthly evaporation.

To permit the determination of evaporation by Eq. (16), it is common to use the following relation:

$$B = Q_h / Q_e \qquad \ldots(17)$$

where B is known as Bowen's ratio, and

$$Q = c_p Q_e (T_e - T_b)/L \qquad \ldots(18)$$

where C_p = the specific heat of water (cal/g-^{o}C)

T_e = the temperature of evaporated water (^{o}C)

T_b = the temperature of an arbitrary datum usually taken as 0^{o}C

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

Introducing these expressions in Eq. (16) and solving for Q_e, we obtain

$$Q_e = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o - Q_v}{1 + B + c_p (T_e - T_b)/L} \qquad \ldots(19)$$

To determine the depth of water evaporated per unit time, the following expression may be used:

$$E = Q_e / \rho L \qquad \ldots(20)$$

where E = evaporation (cm /day)

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

The energy budget equation thus becomes

$$E = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o - Q_v}{\rho [c_p (T_e - T_b) + L(1+B)]} \qquad \ldots(21)$$

The Bowen ratio can be computed using

$$B = 0.61\ (p/1000) \times (T_w - T_a)/(e_w - e_a) \qquad \ldots(22)$$

where p = the atmospheric pressure (mb)

T_w = the water surface temperature (^{o}C)

T_a = the air temperature (^{o}C)

e_w = the saturation vapor pressure at the water surface temperature (mb)

e_a = the vapor pressure of the air (mb)

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

Lake Toba in north Sumatra, Indonesia, is the largest fresh water lake in Indonesia. A detailed study (Sene et al., 1991) has been done on this tropical lake perhaps for the first time to estimate evaporation on the basis of energy budget method. The investigators have indicated the applicability of energy budget method with the assumption that the average energy for evaporation is equal to the net radiation. It is estimated that the annual average evaporation from the lake is about 1.5m. Lake Kinneret (average area = 160 sq.km.) in Israel and in which river Jordan falls is a major contributor to the Israeli water supply scheme. Evaporation from this lake is quite high and has been estimated to be 30% of its water budget (Simon and Mero, 1985). A comparison of lake evaporation measured/estimated from shore pan, lake pan and energy balance method for Perch lake, a shallow (mean depth 2m) lake of 0.45 sq.km in the Canadian shield, was made. It has a dense forest cover around it. The shore sited pan with proper exposure tallied well (3% lower) with the estimate of evaporation from the energy balance method. However, the lake pan estimate was about 40% higher than the energy balance values. This over-estimation by the lake pan could be attributed to the effects of wetting of and the evaporation from the walls caused by the waves rocking the pan. Pan wall temperatures could be considerably higher than water temperatures, leading to rapid evaporation from the wetted areas (UNESCO, 1984). Bolsenga (1975) estimated evaporation of Lake Huron by the energy budget method and compared the same with the available mass transfer estimates. Data were from representative shoreline station measurements and vessel cruise measurements. Agreement between evaporation by the energy budget and by mass transfer was reasonable from February to July. For remainder of the year, the disparity is marked. There was a lack of adequate meteorological measurements on the lake or adequate techniques to extrapolate the quantities from shoreline data.

Penman (Bruce & Clark, 1969) taking advantage of both the mass transfer and energy budget concepts devised an equation which would permit evaporation to be estimated from climatological observations of frequently measured elements. The equation is:

$$E = \frac{\Delta Q_n + \gamma E_a}{\Delta + \gamma} \qquad \ldots(23)$$

where E is evaporation from a lake, Δ is the vapour process gradient at air temperature T_a, i.e., de_w/dT_a, γ is the psychometric constant and Q_n is the net radiation.

$$\begin{aligned} Q_n &= (Q_s - Q_r) - (Q_a - Q_{ar}) \qquad \ldots(24) \\ &= Q_s(1-A) - Q_b \end{aligned}$$

expressed in evaporation units, A is the albedo of water and varies between 0.05 to 0.07 for sun altitude above 55^{o}, Q_b = net energy loss by exchange of long wave radiation, E_a is mass transfer evaporation with the assumption that the lake water surface temperature for the determination of e_w is equal to air temperature, and γ =0.61.

Eq.(23) suggests that total evaporation can be decomposed into two terms, an energy term(Q_n) and a mass transfer or aerodynamic term(E_A). Δ and γ represent the saturation deficits and determine the relative importance of the two terms over a period.

Water Budget Method

It is a very simple method, but it seldom produces reliable results. The use of water balance method requires the measurement/estimation of all other components of the water balance equation (refer to Eq.1) for the estimation of lake evaporation, i.e, O_e.

If the area of the lake varies considerably as a function of water level, a volumetric unit of the components of the balance equation is preferable. For lakes with practically constant surface area, linear units are convenient.

Errors are not critically analysed in water balance studies inspite of the fact that it is very important. It is so because the residual term, whether it applies to groundwater, evaporation or any other component, includes all errors of measured parameters.

It may be pointed out that there is a tendency at present to distribute the error term by equating percentages to equal quantities of water. But, a large percentage error in a small quantity may represent an insignificant amount of water in contrast to a small percent error in a large quantity of water. It is also important that errors associated with different time period be analysed. More data over longer time periods tend to decrease the errors (Winter, 1981).

All the components of the water balance in Eq.(1) except O_e should be estimated independently on the basis of required data. The shorter the time interval, the more stringent is the requirement of accuracy for the computation or measurements of the water balance components. The accuracy of the computation of the water balances of lakes and the minimum allowable balance period are dependent upon the accuracy of the estimation of the basic water balance components, that is, surface inflow and outflow and water storage in the lake. The relative error, B_e (%), of water storage changes compared to the infow is expressed as, (UNESCO, 1981).

$$B_e = \frac{10^4 \, A_v \, \delta \bar{h}}{86400 \, Qt} \qquad \ldots(25)$$

where

A_v = Water surface area of lake or reservoir (km^2)

$\delta\bar{h}$ = error of mean level estimation (m)

Q = discharge into the water body ($m^3 \, s^{-1}$)

t = time interval or duration of balance period (in days).

Equation (25) is used to determine the length of the balance period that will ensure that the relative error, B_e, is not more than 5%, that is, within the limits of accuracy of hydrometric estimates of runoff. If B_e is less than 5% due to increase of inflow (e.g. during rainfall or snowmelt), then it is possible to reduce the length of the balance period.

Errors can be broadly classified into those of measurement and regionalization (interpretation). The latter group of errors is much more difficult to analyse. Measurement errors result from trying to measure a quantity at a point using imperfect instruments and inadequate sampling design and data collection procedures. Regionalization errors result from estimating quantities in a time space continuum from point data.

The actual measurement of each component term of the balance equation is most difficult problem. Some brief description of the errors involved in the major components are discussed.

Precipitation

Factors affecting precipitation over a lake are different from those affecting it over the surrounding land areas and is usually lower than the adjoining land area. Excessive heating of the land surface in warm weather at times lead to the formation of convective precipitation (Kuusisto, 1985). Besides, areal variability of precipitation over a lake can also be considerable. One portion of a lake may get systematically low or high precipitation rates due to the topography of the surrounding area. Also, the lake level measurement must be made at the end of a lake and should be free from any wind effect.

Precipitation into a lake is usually measured by the rain gages located near and around the lake. This conventional method gives rise to unavoidable serious wind errors. If there are islands or islets in the central part of the lake, opportunities increase for more accurate lake precipitation estimates. Rain gages on rafts and floating buoys have also been used. Care should be taken when regionalizing point precipitation values from the surroundings of a lake.

Besides human errors in reading the scale in a precipitation gage, errors could be made for gages that use dipsticks due to (i) water may creep up the stick if the stick is immersed for an extended time (ii) the water displaced by the stick, which makes the reading more by about 1 % .For heavy rains, the tipping bucket gages measures the rainfall lower by 5 % due to loss of catch during tipping operation. Weighing bucket type gages suffer from decreased sensitivity with increased volume of water in the gage. Rain gages with wind shields catch about 20 % more water than gages without wind shield (Linsley et al.,1975).

There are problems of regionalization of point rainfall data. Linsley, et al. (1975), demonstrated that the areal average determined by the arithmetic mean to be 18 % more and that determined from the Thiesson method to be 9 % more than the one determined by isohytel method, respectively. Usually, the sampling error tend to decrease with increasing network density, duration of precipitation and size of area. In general the instrument error could be to the tune of 1-5 % and placement of gages could contribute error in the range of 5-15 % for long term data and as high as 75 % for individual storms. Areal averaging of point precipitation data can be as high as 60 % for individual storm depending on storm type, duration and gage density. These observations are generally based on studies in relatively flat terrain. Errors in measuring precipitation in mountainous areas can be expected to be considerably larger.

In lake studies, hydrologists are interested to determine the amount of precipitation falling directly on the lake rather than throughout the watershed. Use of gages at the lake will have least error. But in case of a large lake, more than one gage is necessary to keep areal variability to a minimum. Gages near the lake provide the next best option if these are read after each precipitation event.

Streamflow and surface outflow

The relative magnitude of the surface inflow and precipitation to the lake depends upon the ratio of the basin area to the water body area. An increase of this ratio leads to an increase of inflow to lakes relative to precipitation (UNESCO, 1981).

The surface inflow into a lake can be subdivided into inflows from rivers and creeks and inflows from numerous small basins surrounding the lake. Part of the latter component consists of non-channelized overland flow which is often overlooked or ignored. Little studies have been made to ascertain the importance of its role in lake behaviour. The non-channelized flow travels in several routes ranging from overland runoff (sheet wash flow) to water flowing at various depths just below the land surface called as interflow (Winter, 1981). This water is important and constitute the only source of surface water for lakes which have no inflowing streams. It is often convenient to subdivide the total surface water inflows into two components:- i) main inflow and ii) lateral inflow. Main inflows refer to the total contribution of the larger gaged rivers and lateral inflow is the total direct runoff into the lake from the surrounding land area including the contribution of the small ungaged stream (UNESCO, 1981). It is debated that there is always some drainage area that contribute directly to the lake as overland flow. But, the matter is somewhat controversial (Winter, 1981).

Continuous observations of discharge should be carried out in as many inflowing rivers as possible. However, the necessary regionalization of these point discharges over the ungaged catchments may lead to serious errors. The limited accuracy of the gaged data at times leads to large errors in evaporation estimates. Kriging method has been successfully used for the point by point estimation of surface level for Lake Winnipeg in Manitoba, Canada. The method is useful for obtaining an overview of the water level for the entire surface of the lake and through an examination of sequential daily water surface profile. Unusual and unrealistic behaviour of the profile can be identified to exclude the suspect gage readings(Zrinji and Burn, 1992).

Errors in measuring stream discharge are related to instrumentation and the methods used to distribute discharge data both in time and in space. Measurement of stream discharge using devices in which flow is routed through them, such as weirs and flumes, can usually be done within 5 percent error, if recording instruments are used to continuously monitor water stage.

Accuracy of current meter discharge measurements are dependent on: (1) the velocity meter, (2) number and distribution of velocity measurements, (3) time of exposure of the meter, and (4) measurement of the cross sectional area of the channel. Tests of velocity meters indicate that errors are generally less than 3 percent at low velocities and less than 1 percent at higher velocities. Comparative calibration of meters between pre and post field use have shown that their ratings change by less than 10 percent. Errors related to estimating velocity distribution, which is related to the number of point measurements in both the vertical and horizontal directions, can be about 5 percent. Errors related to exposure time of the meter can also be about 5 percent. Considering all these factors, Fig. 2 can be used to summarize the overall error to be expected in a current meter discharge

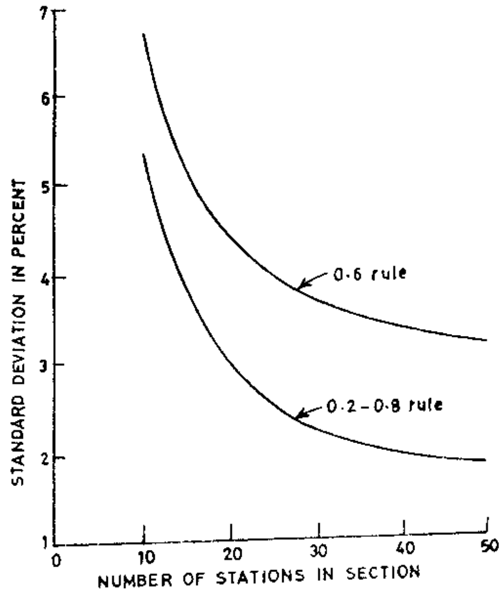

Fig. 2: Standard Deviation of Total Error of Discharge Measurements (Modified from Carter 1973)

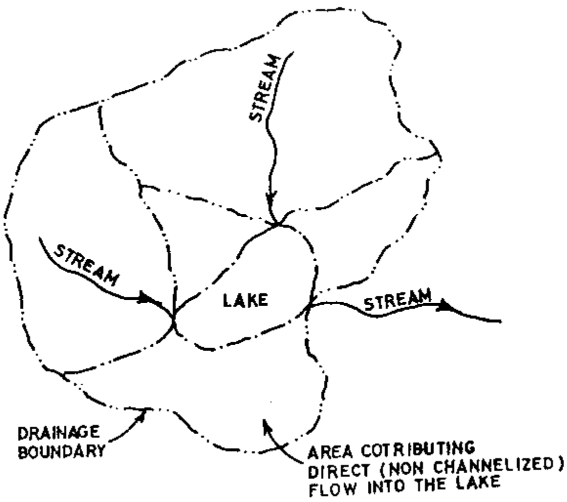

Fig. 3 Example Lake and Drainage Basin Showing Area Contributing to Nonchannelized Surface Inflow

measurement. In general, if many verticals are used, the error can be kept to less than 5 percent, but if few are used and exposure times are short, the error can be between 5 and 10 percent, or even higher.

Errors related to determination of temporal distribution of stream discharge can be great, well over 100 percent, if occasional miscellaneous measurements are averaged. If correlation techniques are used to extend records, errors in the regression have been shown to vary widely, and caution should be used. Stage discharge relationship curves can vary widely in quality; if a good relationship is developed, the error in estimating discharge is probably less than 5 to 10 percent.

Regionalization of stream discharge information from gaged to ungaged watersheds has been shown to be in error by as much as 70 percent. Use of multiple regression to estimate low flow stream discharge based on basin characteristics has been shown to be in error by greater than 100 percent in many studies, but as little as 10-15 percent in others.

Estimation of surface water input to a lake by regionalization from gaged to ungaged watersheds should be done with caution, and should not be considered an accurate technique for balance type studies.

Lake hydrologists must be aware of overland type flow into a lake, because there will always be parts of the lake's watershed that cannot be gaged by stream gaging techniques(Fig.3). This non channelised flow is important for lakes which have no inflowing streams. Lack of understanding of overland flow remains one of the serious drawbacks to water balance studies of lakes; it should receive increased attention.

Groundwater

Lakes are three- dimensional depressions in the landscape that generally intersect the water table, and the groundwater flow patterns around and below a lake may be complex (Almendinger, 1990).

Until recently, the ground water component of the lake system has been ignored or considered unimportant relative to other components of the hydrologic system. The interaction of lakes and groundwater is the least understood component of lake hydrology. Groundwater flow is next to impossible to measure directly and must be estimated from a knowledge of local gradients of the water table and aquifer properties (Strahler and Strahler, 1973).

Understanding the interaction of lakes with groundwater requires an understanding of the dynamic character of the distribution of hydraulic head in the groundwater system. This distribution of head is controlled to a large extent by the distribution of recharge, which, in turn, is directly related to infiltration and water movement through the unsaturated zone. Therefore, the entire continuum from infiltration to redistribution in the unsaturated zone to movement into and through the groundwater zone to movement through the bed of a surface water body needs to be studied (Winter, 1983).

Groundwater flow is of utmost importance in lakes which do not have inlets and outlets like seepage lakes in which water passes out as ground water discharge lowering the watertable because of which these lakes may come across extreme drop of water level and occasional complete emptying also (Reid and Wood, 1976). Seepage lakes not only lose water by ground water but are recharged by ground water only (Thurman, 1985).

The interaction between a lake and an aquifer is different to that of a spring, river or canal because the area of contact between the lake and the aquifer is much larger. Consequently the flow patterns in the vicinity of the lake tend to be more complex, so that water may flow from the aquifer into the lake in one region whereas in another region the same lake water may be transmitted from the lake back into the aquifer. This can be decided upon from the groundwater contour drawn from the regular observation of ground water levels (atleast during premonsoon and postmonsoon season) in the vicinity of a lake.

Anisotropy of the geologic materials within the groundwater system is one of the most difficult parameters to obtain. It is seldom the subject of field investigations. Lack of knowledge of the relative distribution of zones of high and low hydraulic conductivity can lead to serious misinterpretation of the interaction of lakes and groundwater.

Empirical Formulae

The energy-budget and mass-transfer methods used to estimate evaporation amounts are theoretically correct, but require data which, for many studies, are not readily available. Also, in many cases it is highly questionable whether it is economically feasible to instrument a lake sufficiently to acquire these data. As such, under such circumstances, the empirical formulae are used to estimate of evaporation from a lake.

Most of the empirical equations are based on the simple aerodynamic equation discussed earlier in Eq.4 with modifications to account for some factors affecting evaporation such as wind velocity, temperature etc. Some of the equations commonly used are (Reddi, 1992):

Fitzgerald's equation

$$E = (0.4 + 0.124u_o)(e_v - e_a) \qquad \ldots(26)$$

where E = evaporation in mm/day

e_v = saturated vapour pressure at the temperature of the water surface in mm of mercury

e_a = the actual vapour pressure of air in mm of mercury

u_o = mean wind speed at the surface in km/h.

Horton's Equation

$$E = 0.4(\psi e_v - e_a) \qquad \ldots(27)$$

where ψ is a function of wind velocity given by $\psi = 2 - e^{-0.124u}$ and E, e_v, e_a and u have the same meaning as in Eq.(26). For large areas, Horton suggested that the result of Eq.(27) be multiplied by the area factor $\{(1-P) + P(\psi-1)/(\psi+1)\}$, in which P is the fraction of the time during which the wind is turbulent and h is the relative humidity expressed as a fraction.

Meyer's equation

$$E = C(e_v - e_a)(1 + 0.06215\, u_{10}) \qquad \ldots(28)$$

where E = evaporation in mm/month

e_v = the saturation vapour pressure in mm of mercury corresponding to the mean monthly temperature of air

e_a = the actual vapour pressure in air based on mean monthly temperature and relative humidity

u = monthly mean wind velocity in km/h at 10 m above ground.

C = 15 for small shallow ponds.

For large or deep water bodies the value of C is taken as 11 and also e_s is the saturation vapour pressure corresponding to mean monthly water temperature instead of air temperature and e_a is the actual vapour pressure in air about 10 m above the water surface.

Lake Mead equation

$$E = 0.0331u_o\ (e_v - e_a)\ [1-0.03\ (T_a - T_v)] \qquad \ldots(29)$$

Where T_a = average air temperature in ^oC

T_v = average water surface temperature in ^oC

and e_v, e_a and u_o retain the same meaning as in Eq. (26)

The greatest appeal of empirical formulae is their simplicity and the fact that they permit estimates of evaporation to be made from standard meteorological data.

Though the empirical equations provide fast estimates of evaporation they must be used with caution. Their application becomes difficult if the data required in the equation is not available at a place of interest. Moreover, most of the meteorological quantities used in these equations are average daily values, whereas evaporation depends upon the total quantity of incoming solar radiation and the average may not represent the total. The equations require a measure of the surface temperature of the body of water. This measurement is, of course, very difficult to obtain. If the mean air temperature is used in place of the surface temperature, then the effects of advected energy to the lake on evaporation are not considered. This may cause considerable error in the calculated amounts of evaporation, because small errors in temperature induce large errors in the calculations. Also, measurements of the wind speed and vapour pressure should be taken at the heights specified by the equation which is used.

Ex. The following meteorological data pertain to a large lake with a waterspread of 15 km^2. The data represents the average values for a day.

Water temperature=$24^{o}C$, Air temperature=$26^{o}C$,

Wind speed at 0.5 m above the ground level (u_o)= 25.3 km/h

Relative humidity= 46%

Solution:

Saturation vapour pressure for $24^{o}C$ (e_v)= 22.43mm

Saturation vapour pressure for $26^{o}C$= 25.27mm

Therefore e_a= 0.46 x 25.27= 11.62 mm

The velocity at 0.5 m above the ground level has been taken to be mean wind speed at the surface.

(i) Fitzgerald's equation

$E = (0.4 + 0.124u)\ (e_v - e_a)$

$= \underline{38.24\ mm/day}$

(ii) Horton's equation

$E = 0.4\ (\psi e_v - e_a),\ \psi = 2-e^{-0.124u}$

$\psi=1.956$, $E=\underline{12.91 mm/day}$

(iii) Meyer's equation

$E = C(e_v - e_a)\ (1 + 0.06215\ u_{10})$

Since the evaporation is estimated for one day and it is a large lake, the value of C =11/30. Further the the wind speed to be used here is at a height of 10 m above ground level. The same can be obtained by using the EQ. (3) as:

$$\frac{u_{10}}{u_{0.5}} = \left[\frac{10}{0.5}\right]^{0.15} \quad \rightarrow \quad u_{10} = 39.65 \text{ km/h}$$

$$\underline{E = 13.73 \text{ mm/day}}$$

(iv) Lake Mead equation

$$E = 0.0331 u_0 \, (e_v - e_a) \, [1 - 0.03 \, (T_a - T_s)]$$
$$= \underline{8.51 \text{ mm/day}}$$

It may be seen that Fitzgerald's Eq. gives very high value of the estimate of evaporation whereas the other three equations give comparable value. Choice of a particular equation will depend on the local meteorological factors and physical characteristics of the lake.

Kohler et al. (1959 vide Roberts and Stall, 1966) in their U.S.Weather Bureau Technical Paper No. 37 suggested a graphical technique for evaluating the daily evaporation from shallow lakes on the basis of the energy budget and aerodynamic approach (Fig. 4). The nomograph is built around four meteorological parameters, viz., solar radiation, air temperature, dew point, and wind movement. A field worker has to enter single observations of these four parameters into the nomograph to determine daily evaporation. A typical example is illustrated in the nomograph. The four input data used were: air temperature, 27^{o}C; dewpoint, 13^{o}C; total daily wind movement, 193 km/day; and solar radiation, 25.1 megajoules per square meter. In the upper left quadrant, locate the intersection of the 25 megajoules per square meter curve and 27^{o} C temperature line. Follow the horizontal line to its intersection with the 13^{o} C dewpoint curve and project a vertical line downward to intersect the line of 193 km/day of wind. From this point proceed horizontally to the left to intersect a vertical line down

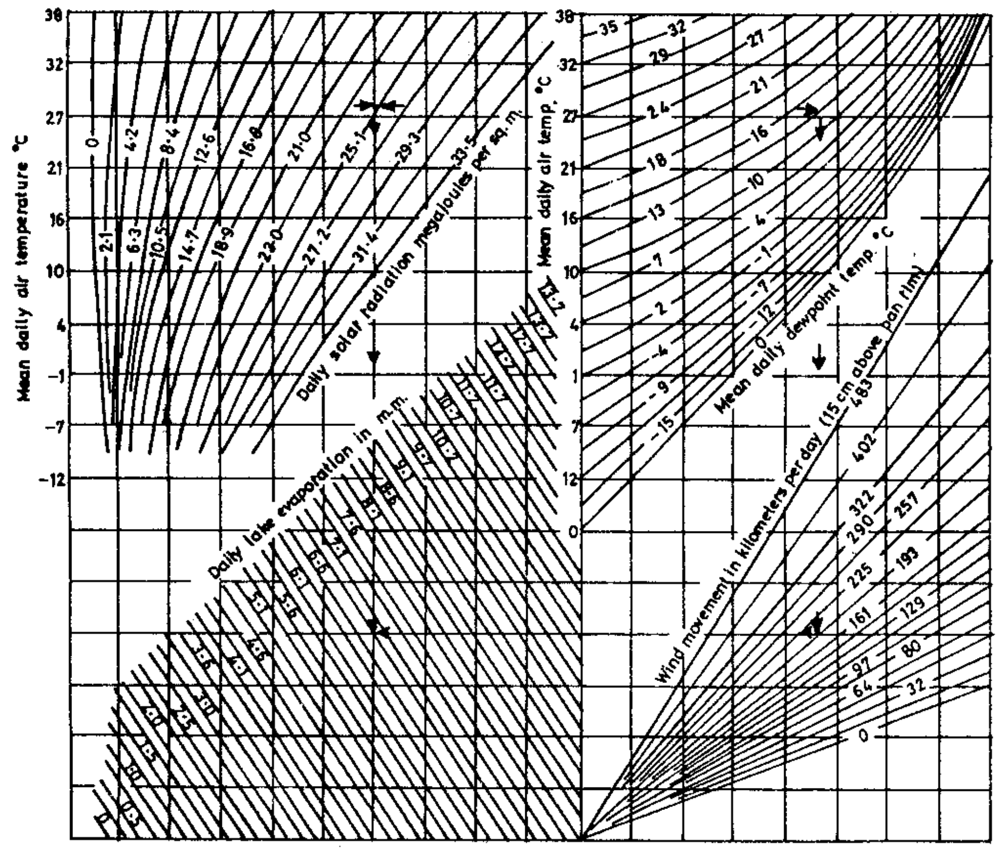

Fig. 4 Shallow lake evaporation as a function of solar radiation, air temperature, dewpoint and wind movement (Kohler, 1955)

from the first intersection point in the upper left quadrant. The daily lake evaporation, 7.1 mm, is observed at this point.

The operation of the nomograph with the four meteorological parameters is relatively simple for one set of data. However, it would be time consuming to process historical records to estimate evaporation in this way. Lamoreux (1962 vide Roberts and Stall, 1966) developed a formula to solve the 4 quadrant nomograph. The formula could be conveniently used to process large quantities of historical data to estimate daily evaporation by a digital computer. The equation is:

$$E = [e^{(T_a - 212)(0.1024 - 0.01066 \, \mathrm{Ln}\, R)} - 0.0001 + 0.0105(e_s - e_a)^{0.88}(0.37 + 0.0041 u_p)]$$
$$\times [0.015 + (T_a + 398.36)^{-2}(6.8554 \times 10^{10})\, e^{-7482.6/(T_a + 398.36)}]^{-1} \qquad \ldots(30)$$

where T_a is the mean daily air temperature in ^{o}F, R is the solar radiation in langley per day, and u_p is the total daily wind movement in miles/day, and $(e_s - e_a)$ is the saturation deficit in inch of mercury. This method is applicable for lakes and where the net advection is not appreciable. However, it is important only for monthly and seasonal distribution of annual evaporation. The method thus will be prone to over estimate evaporation under calm and humid conditions and under-estimate for dry and windy conditions (Linsley et al., 1975).

Measurement from Evaporation Pan

Measurements taken on evaporation pans have been used for many years to provide estimates of the amount of evaporation from lakes and reservoirs. The popularity of the pans stems from the fact that they are inexpensive, simple to install and easy to use and generally the annual ratio of lake evaporation to pan evaporation remains reasonably constant from year to year. In evaporation pans, the depth of evaporation during any time interval is measured as the drop in water surface level in the pan during that time interval corrected for precipitation, if any, obtained from an adjacent standard raingage. The observations are usually taken on daily basis and after taking the measurement on each day, the water level in the pan is restored to a stipulated value by adding or removing required amount of water.

The lakes have considerably different wind and thermal regimes than pans located on the land. Floating pan can only partly overcome this disadvantage. In the studies being done at the Institute of Hydraulics and Hydrology, Poondi, to find a material for floating evaporimeter whose thermal conductivity is equivalent to that of water and at the same time non leaky and light in weight, perspex sheet which is akin to glass but at the same time non brittle, non leaky and workable was chosen for the fabrication of floating evaporimeter installed at Poondi reservoir. The unit has the sliding arrangement which follows the water surface and could be fixed at the desired location. A graduated gauge of requisite least count when fixed to the frame-work shall enable the observation of water level fluctuations at the site of evaporation through the transparent perspex sheets (Makwana, 1992).

In addition to the problem of designing the ultimate evaporation pan, an even more perplexing problem is the relationship between pan evaporation and lake evaporation.

Although floating pans are supposed to get rid of some of the difficulties, it has been the aim of evaporation researchers to find lake evaporation from land based evaporation pans because these are easiest to install, service and maintain and a considerable pan-evaporation data base already exists in many places.

The rate of evaporation from a pan is greater than that from large water bodies. So a suitable pan coefficient is used to convert the pan observation to get an estimated value of evaporation for a lake. The most commonly used coefficient to estimate lake evaporation from class A pan data is 0.7. Kohler and others(1959 vide Winter,1981) had calculated evaporation from lakes by converting measured evaporation from pans by applying a coefficient. Bleney had studied the effects of high altitude on evaporation from pans and determined suitable coefficients. Studies by Bigelow has shown that the location of pans relative to the water of a reservoir has significant effect on the calculated evaporation. He concluded that evaporation from natural lakes is about five eighth of that measured from an isolated pan placed outside the vapour blanket. Further studies by Rohwer, Kohler, Mansfield showed that the evaporation coefficient ranges anywhere between 0.2 to 1.5 and this factor is dependent upon the size, depth and location of pan. In the Experimental Lake Area (ELA) in the northwestern of Ontario, Canada, the annual average lake evaporation is about 0.7 times the lake evaporation.

It is essential that the coefficient of evaporation be measured under all different conditions, which is not practically feasible in large water storage systems. The ratios of annual lake evaporation to pan evaporation are found to be consistent from year to year and region to region but exhibit considerable variation from month to month. Pan should not be used to estimate evaporation for shorter time period. Linsley et al.,(1975), pointed out that the ratio should only be applied to annual data and only if effects of wind (advected) energy into the lake and the heat transfer through the pan are considered. It may be thus possible to estimate lake evaporation within 10-15 % by applying a coefficient to pan data with due consideration to lake depth and climate regime.

It is widely recognized that the coefficient should be lower for lakes in arid regions than for lakes in humid climates. A value of 0.52 was obtained for the Salton Sea, California and 0.81 for Lake Okeechobee, Florida (Hounam, 1973 vide Kuusisto, 1985). The annual average Class A pan coefficient to be 0.69 for lake Hefner, Oklahoma. This is in fair agreement with the results of other investigations indicating that the evaporation pan method of determining annual lake evaporation may be accurate to within perhaps 10 or 15 percent, provided care is taken in measuring pan evaporation and selecting the coefficient to be used. In cold climates where lakes are ice covered in winter, the Class A pan coefficient for the open water season also tends to be large (Jarvinen, 1978). In addition to the regional variation of pan coefficients, there is a remarkable seasonal variation for many climates. The monthly evaporation pan coefficients vary more widely and with a greater range of probable error than the annual coefficients. The coefficients tend to be smaller than the annual average in the winter and spring and larger in summer because of the lag between lake water temperature and the pan water temperature. Since the temperature lag is greater for deep lakes, it is expected that the monthly variation in the coefficients is greater for deep lakes for a climate which has large seasonal variations in temperature. From studies made on seven Australian lakes, it was inferred that the effect of heat storage on lake evaporation is less in tropical climates than in subtropical climates and monthly pan coefficients show much smaller variation in the former than in the latter and depth seems to be the most important parameter in determining the variability of monthly lake to pan coefficients for subtropical lakes (Garrett and Hoy, 1978).

Obviously the use of constant value for each month can lead to serious errors. Butler (1957) suggested the method of computation of a reasonable value for class A evaporation pan coefficient for a given month.

$$C = \left[\frac{(e_o - e_z)_{Lake}}{(e_o - e_z)_{pan}} \right] * 0.69 \qquad \ldots.(31)$$

where, C = pan coefficient for a given month,

e_o = vapour pressure of Saturated air at the temperature of the water surface for the given month,

e_z = corresponding average vapour pressure of air at some specific height (usually 2m) above the water surface.

US Weather Bureau measured evaporation at about 450 locations in the continental United States. The data collected from these pans were used to estimate the evaporation rate of nearby lakes after multiplying the pan evaporation by a constant called the pan coefficient. This had been done on the assumption that the lake evaporation is a fixed percentage of the pan evaporation.

Ferguson et al. (1970), analysed 5-year class A evaporation pan data supplemented by 10 year climatological data from 100 stations across Canada. The mean monthly and annual lake evaporation maps over Canada were prepared. In this investigation, lake evaporation, determined from class A pan evaporation is usually obtained from pan data by applying a correction to allow for energy transfer through the bottom and the sides of the pan.

Lake evaporation from the data available from 104 evaporation stations (1959 to 1977) in India was estimated by using the lake to pan coefficients (Ramasastri, 1987). As the evaporation stations pan in India are mesh covered, the observed class A evaporation pan data are adjusted by the factor 1.144 to obtain the evaporation from open pan. He used the monthly coefficients as 0.6 in cold dry winter months, 0.8 in hot humid summer months and 0.7 in the transition months between the winter and the summer months. As winter in the southern portion of India is less severe and brief, different months of winter season were considered for the southern and northern portion of India. For this purpose, 22^{o} latitude has been taken as the demarcating line. The pan coefficients considered were:

North of 22^{o} Lat.	Nov.-Feb.	Mar.-Apr. Sept.-Oct.	May-Aug.
South of 22^{o} Lat.	Dec-Jan	Feb.- Mar. Sept.-Nov.	May-Aug.
	0.6	0.7	0.8

With above-mentioned coefficients, the lake evaporation was estimated at all the 104 stations using the period averages for the length of the data available. The isopleths were drawn with the estimated values which provided the values of the evaporation taking place from large lakes (Fig. 5) for monsoon (Jun-Oct.), non-monsoon (Nov.-May) and the calendar year. It was observed that the evaporation was highest during the months of April and May and decreases during rainy season and winter months.

Mean monthly and annual evaporation data collected at more than 200 stations by IMD proved to be very valuable in field estimations. The volume of water lost due to evaporation from a lake in a month is, thus, calculated as (Subramanya, 1994):

$$V_E = A\ E_{pm}\ C \qquad \ldots(32)$$

where V_E = Volume of water lost in evaporation in a month (cu m)

A = Average lake area during the month (sq m)

E_{pm} = Pan evaporation in a month (m)

C = Relevant pan coefficient

The monthly pan coefficient could be estimated from the Eq. (31).

Evaporation from a water surface is a continuous process. Under typical Indian condition, evaporation loss from a water body is about 160 cm in a year with enhanced values in arid region. The quantity of stored water lost by evaporation in a year is indeed considerable as the surface area of many man-made and natural lakes in India are very large.

4.0 Conclusions and Discussions:

The Meyer Equation (No. 28), Kohler et al. Equation (No. 30), the Energy Budget method and 0.7 times the class "A" pan evaporation gave estimates of seasonal evaporation which were within 2 % of that determined by the Water Budget method. However, these methods except the Energy Budget method did not give same

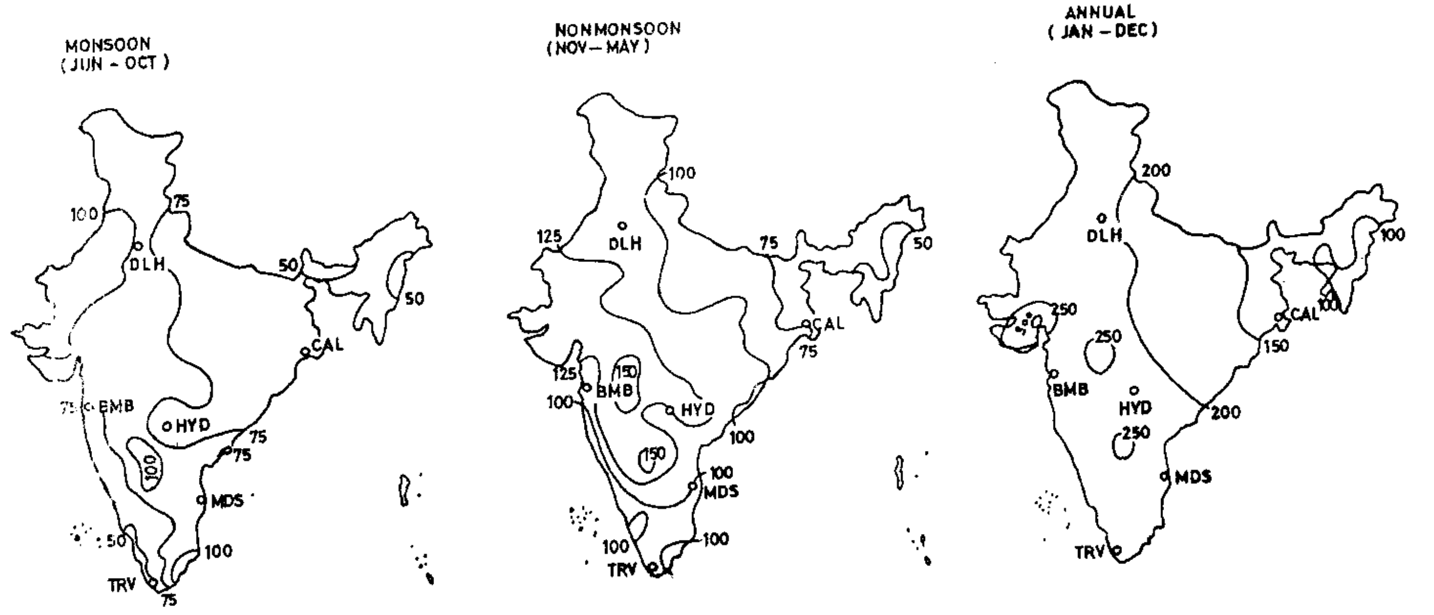

FIG.5 : FREE WATER SURFACE EVAPORATION (Cm)
(Vide Ramasastri, 1987)

agreement on a monthly basis. Energy Budget method alone gave consistent results seasonally and monthly. These findings were available from the comparative study conducted by Mckay and Stichling (1962 vide Gray 1973) for the Weyburn Reservoir located in southern Saskatchewan. The energy budget method which requires special instrumentation and measurement may be used for operational problem where high accuracy is demanded. It was observed that Class "A" pan data and Meyer Equation also give acceptable estimates of evaporation and thus could be resorted to if the evaporation estimates are required for design purposes. They opined that even Meyer Equation could be used for both design and operational purposes if reliable measurements or estimates of air and water surface temperatures are available.

As such, a water surface temperature network is required for having the water temperature - air temperature relationships more precisely. Small errors in temperature estimates introduce large errors in the estimate of evaporation.

The floating pan may yield good quality estimates of seasonal evaporation if its coefficient is established. However, in view of the servicing difficulty, the pan is not likely to gain general favour.

For a shallow lake, the evaporation will normally approximate closely to the seasonal air temperature regime. This means that maximum rates of evaporation from a shallow lake will be during the summer season and minimum rates in the winter season. However, in case of large and deep lakes, water temperatures commonly lag behind the temperature of the overlying layer of atmosphere. During the summer and early spring months, considerable depths of water are slowly and gradually warmed up by a part of incoming solar radiation which would otherwise be available for

evaporation. Subsequent release by stored heat in winter and autumn seasons provides a heat energy in excess of that received directly from the sun for evaporation. Thus, the net result of this heat storage
on relative air and water temperature are:
(i) in summer, the water temperatures are lower than the air temperatures, and
(ii) in winter, the water temperatures are higher than the air temperatures (Ward, 1967).

The mass transfer methods give satisfactory results in many cases and normally use easily measurable variables and have simple model form. However, wind speed and air temperature are generally measured at inconsistent heights resulting in a large number of equations with similar and identical structure. The wide ranging inconsistency in meteorological data collection procedures and standards has given rise to over one hundred evaporation formulae and has made it impossible for a fruitful comparative study.

Evaporation estimations from the mass transfer based evaporation equations were found to be particularly sensitive to vapour pressure gradient, less sensitive to wind speed and least sensitive to temperature data. Systematic errors in input data influence the evaporation estimates to more or less the same magnitude for both monthly and daily cases, while random errors were found to have a much more significant effect on evaporation estimates in the monthly cases than in the daily cases. (Singh and Xu, 1997).

An inspection of the mass transfer equation reveals that only meteorological factors;- i.e., vapour pressure gradient, wind speed, temperature, have major influence on evaporation. Factors like air pressure, density of water and water surface elevation for a given location may not greatly affect the rate of change of evaporation.

The energy budget method is reliable in theory and suitable for research purposes only in small areas because of their requirements for detailed meteorological data, such as net radiation, sensible heat flux, etc. Their practical utility for large lake is limited. Though a simplified general version of energy budget is available, still it requires an evaluation of the net radiation which is not so easily obtainable for many applied engineering problems.

The water budget method is simple in theory. But, rarely the method produces reliable results in practice. The main difficulty is that some of the variables, such as seepage rate are difficult to measure. There are other problems also associated with the water budget method.

It is generally accepted that empirical formulae are reliable in the areas and over the periods for which they are developed, but large errors are not uncommon when they are extrapolated to other climatic areas without recalibrating the constants involved in the formulae. This is because of the fact that the formulae relate evaporation to meteorological factors using regression analyses. The empirical formulae have a limited range of applicability because (i) their variables may not be easily measurable in other places, or some existing data may not be utilised, (ii) they are accurate only in a limited range as their model structure may be only partially correct, and (iii) it is difficult to compare one method with another due to method-specific model variables, e,g., the requirements for measurement of temperature and wind speed may be at different heights above the water or ground surface.

For the estimation of evaporation from the pan measurement, the thermal regimes of pans and lakes are usually markedly different. Seasonal changes in subsurface heat storage are not reflected in pan observations. Floating pan, which has its inherent operational and logistic difficulty, can be used to overcome some difficulty. As discussed earlier, the rate of evaporation from a pan is greater than that from a lake. Therefore, by using a suitable pan coefficient as discussed in detail in the earlier section, an estimate of lake evaporation could be made.

In fine, it may be summarised that there is no universal and sacrosanct method to estimate lake evaporation. There are a plethora of methods and formulae and choosing a method or formula amongst them will depend on the purpose and the desired accuracy associated with it, and data available. An attempt has been made herein to highlight the usage and capabilities of some of the important methods/formulae in vogue to assist the field engineers/investigators to choose and work with the most suitable method/formula either to make a first hand estimate of lake evaporation specially for a scenario where virtually no evaporation data/estimate is available or to make an elaborate and detailed study on lake evaporation for a lake where some preliminary studies/estimations are available.

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