

WATER BALANCE OF LAKES

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INTRODUCTION

Lakes either man made or naturally occurring are constantly used from time immemorial in many parts of the world. Particularly in mountainous regions, the importance of lakes increases because in most cases they are the only source of water for a variety of purposes such as irrigation, drinking, hydropower generation, pisciculture and recreation. Further, a lake plays a significant role in shaping the hydrological, ecological, and environmental balance of that region. Lake is also a place for sanctuary for migrating birds, developing flora and fauna and an excellent spot for habitation of aquatic biota which are important for maintaining the ecological and environmental balance and also the hydrological cycle. All the above factors are either directly or indirectly linked up with the overall economic development of the region and the country as a whole. Any change in the lake water both in quality and quantity which may arise for various reasons, will certainly hamper the development of the area. The impact of degradation may be less at the initial stage, however, the cumulative impact in a longer period would be significant and in many cases, it may be dangerous. Problems may be serious particularly in cases when the lake catchment is urbanised. It is, therefore, necessary to study the hydrological features of a lake to understand the growing problems and to take the suitable measures well in time to save the lake and adjoining area from ecological, hydrological and environmental disasters.

The water balance of lakes provides very useful information about the availability of water in lakes at any time. In order to utilize the lake water in a planned and systematic manner and to manage the required availability of water in a lake, the knowledge of different components of water balance of lake is essential. The general principles of water balance and computational techniques using conventional methods as well as isotopic techniques have been discussed in the following sections.

WATER BALANCE OF LAKE

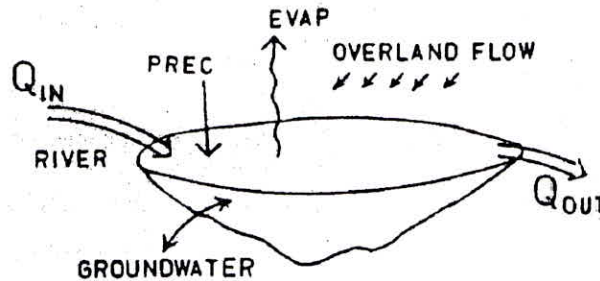
The water balance of a lake is same as the water balance of any drainage area. The lake water balance approach physically accounts for the components of outflow from the system, inflow to the system and changes of storage within the system. In general, the water balance equation

for a lake can be expressed in its simplified to form as given below

$$Q_{in} - Q_{out} = \Delta V / t \quad \text{or} \quad \Delta V = (I - Q - E) \Delta t \quad \dots\dots(1)$$

Where,

- ΔV = change in water storage of the lake
- I = inflow to lake
- Q = outflow from lake
- E = net evaporation,
- Δt = time interval for which the water balance is to be determined.



After considering the different input and output parameters, equation (1) can be written specifically as,

$$\Delta V = (S_I + SS_I + P_I - S_O - SS_O - E_O) \Delta t \quad \dots\dots\dots(2)$$

where,

- ΔV = change in water storage;
- S_I = Surface inflow,
- SS_I = Subsurface inflow;
- P_I = Lake precipitation,
- S_O = Surface outflow;
- SS_O = Subsurface outflow,
- E_O = Lake evaporation;

The surface inflow includes inflow from the surface of lake catchment such as overland flow, stream/nalah runoff, spring inflow, waste water drain also flows in the form of channels etc depending upon the lake catchment. Surface outflow component includes the water drained out under controlled conditions, water flow out in the form of streams or rivers and the water withdrawn from the lake to meet out various requirements of lake catchment area or outside. Subsurface inflow represents the groundwater contribution to the lake and subsurface outflow represents seepage/leakage from the lake to surrounding groundwater aquifer. Change in lake storage represents the lake level fluctuation.

TECHNIQUES FOR WATER BALANCE ESTIMATION

The water balance of lakes can be computed by using conventional techniques. However, there are few components like sub-surface inflow and outflow which are difficult to estimate using the conventional techniques with the desired accuracy. Therefore, isotopic techniques are employed to compute these parameters. The details of isotopic techniques are discussed in

the subsequent sections while conventional methods for computing water balance components of lakes are described below.

Conventional Method

The estimate of following parameters is required to estimate the water balance using conventional techniques.

a) Lake Precipitation (P_1)

Precipitation into a lake is usually estimated from the raingauges located near and around the lake. This method gives rise to unavoidable wind errors. A more accurate estimate of lake precipitation can be achieved by installing raingauges on the islands, if any, in the central part of the lake. Raingauges on the rafts have also been used. Care should be taken in regionalizing point values from the surroundings of the lake because lake precipitation is generally lower than precipitation over surrounding land areas. The ideal requirement is to average the values from raingauges installed all around the lake, upstream as well as downstream portions. The total number of raingauges to be installed should depend upon the size of the lake and the variability of rainfall over different parts of the lake.

b) Surface Inflow (S_1)

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. Continuous observation of discharge is carried out in as many inflowing rivers as possible. In case of the perennial rivers, rating curves can be computed for manipulating the discharge values for any time period. The main source of error is due to regionalization of point discharge data from the gauged basin over the ungauged catchments. Therefore, the portion of gauged catchments should be as large as possible. For the seasonal inflows into the lake through streams, nalas etc., the direct measurement of discharge is difficult, particularly if the number of such inflows is large. Instead, a relationship should be developed between precipitation and discharge through such inflows from which discharge can be computed. In this case, however, the precipitation should be measured very accurately. To measure the surface inflow from nalas and small streams V-notches or weirs can be permanently installed at suitable places. Alternatively, pigmy type current meters can be used to measure the flow velocity of the streams or nalas. In case of continuously flowing nalas, automatic water level recorders can also be installed at 'V' notches or weirs in order to have a continuous record of discharge.

c) Lake Evaporation (E_0)

The estimation of evaporation is rather complicated. Several equations for estimating evaporation from free water surface exist in literature. In a recent and most comprehensive article, Winter *et al* (1995) have evaluated 11 equations for the estimation of lake evaporation. Of the three types of approaches to the estimation of evaporation, viz. energy balance, mass transfer and combination (Penman type), the first one yields more reliable results, if data on all associated energy terms are available. Alternatively, if sufficient data is

available, the combination methods also yield better results. The equation for the estimation of evaporation is (Jensen, 1974):

$$E = \frac{\Delta}{\gamma + \Delta} * (Q_N + G) + \frac{\gamma}{\gamma + \Delta} * 15.36 * (0.5 + 0.01 * U_2) * (e_0 + e_a) \quad \dots\dots\dots (3)$$

Where, γ and Δ are the weighting factors. Δ is the slope of the vapour pressure versus temperature curve and is expressed in mb/EC. The following equation can be used for the computation of Δ , which is the slope of the vapour pressure vs. temperature curve.

$$\Delta = \frac{25083}{(T + 237.3)^2} \exp\left(\frac{17.3 * T}{T + 237.3}\right) \quad \dots\dots\dots (4)$$

where T is the mean air temperature in EC. γ , the psychrometric constant is computed using the following equation:

$$\gamma = \frac{0.61 * P}{1000} \quad \dots\dots\dots (5)$$

where P is the mean atmospheric pressure in mb.

Q_N is the net solar radiation. If the solar radiation data for a site is not available, the net radiation can be computed using the following relationship:

$$Q_N = Q_S * (1 - \alpha) - Q_{LN} \quad \dots\dots\dots (6)$$

where, α is the albedo (reflection coefficient), Q_S is the global solar radiation and Q_{LN} is the net long wave radiation. In general the value for α in case of lake evaporation studies is taken as 0.06. Alternatively, the α can be computed by using the values presented for different amount of cloud cover for different latitudes (Ter-Makaryantz, 1960). Rao *et al.* (1971), after analysing meteorological data collected from India Meteorological Department (IMD) Stations from all over India, proposed the following equation for the computation of Q_S .

$$Q_S = Q_A * [0.325 * \cos \Phi + 0.385 * \left(\frac{n}{N}\right)] \quad \dots\dots\dots (7)$$

where Φ is the latitude and n and N are actual and maximum possible sunshine hours, respectively. Q_A , the extraterrestrial radiation and N for the lake site can be computed using the Duffie-Beckman equations (Allen, 1996). Q_{LN} shall be computed by the following equation (Shuttleworth, 1992):

$$Q_{LN} = - f \epsilon' \sigma T^4 \quad \dots\dots\dots (8)$$

where, σ is the Stefan Boltzmann constant, T is the absolute temperature in Kelvin, f is the adjustment factor for cloud cover, which is roughly equal to $(0.1 + 0.9 * n/N)$; Singh, 1992) and ϵ' is the emissivity factor, which can be computed using the Idso-Jackson equation. U_2 is the wind velocity in m/s at 2m above surface. e_0 and e_a are the saturated vapour pressure (mb) at

the water surface temperature and actual vapour pressure (mb) at air temperature, respectively. e_0 can be calculated using the following equation (Linsely et al., 1975):

$$e_0 = 33.8639 * [(0.00738 T_0 + 0.8072) - 0.000019 * T_0 + 48 + 0.001316] \dots (9)$$

The actual vapour pressure e_a can be computed by multiplying the relative humidity with the saturated vapour pressure at air temperature.

In general the errors associated with such empirical methods are of the order of 15 to 20%. This error can be substantially reduced if one uses site specific data, instead of using data collected from far-away meteorological centres. Pan evaporimeters installed at the lake site also give reliable results, if the pan-coefficients are properly evaluated. The detailed about the estimation of evaporation is covered in later chapter.

d) Surface Outflow (S_0)

In general, surface outflow is the most accurate component in the lake water budget. The volume of water used for different purposes are generally known precisely which can be used in the water balance of lakes directly. For both regulated and unregulated lakes, the error in this component can usually be of the order of 15%.

e) Storage Changes (ΔS)

The change in storage of lakes is directly estimated from the data on the dimension of lake, depth of water column, and water level fluctuations. The sources of error in the estimation of storage change involve inaccuracies in the measurement of water level fluctuations. Some other factors have minor influences; these include changes of water density, sedimentation, copious aquatic vegetation and ice phenomenon. Computation of the lake volume using depth contours can be made by using the following methods (Zumberge, and Ayers, 1964).

- i) Volume of lake may be determined by measuring the area enclosed by the depth contours (the location coordinates plotted against depth) and the area under the curve so obtained may be planimetered or otherwise measured.
- ii) In another method, the area enclosed by successive pairs of depth contours is averaged out and multiplied by the contour interval to yield a series of volume elements which are then summed using the following relation,

$$V_{A_1A_2} = h/3 (A_1 + A_2 + \sqrt{A_1A_2}) \dots (10)$$

where, h is the contour interval, A_1 is the area enclosed by the upper contour and A_2 is that enclosed by the lower contour. Summation of the results of repeated successive applications of above equation will yield the lake volume.

f) Subsurface Components: Inflow (SS_i) and Outflow (SS_o)

The subsurface components of a lake may include inflow from groundwater and seepage through the dam, if the lake is regulated. Variations in bank storage can be considered as a

special case of underground flow. The proportion of underground flow in the hydrologic budget of lakes varies widely. There is no direct method for computing the subsurface components of a lake. Normally the net groundwater flow (subsurface inflow minus subsurface outflow) is calculated as a residual of the lake water balance, with knowledge of all other components. This, however, does not provide the information on the magnitude of the subsurface inflow and outflow.

The groundwater inflow/outflow rates may be estimated by groundwater modelling techniques, through use of flownets or seepage meters. In the absence of an appropriate network of piezometers both inside and around the lake, it may not be feasible to use the groundwater modelling studies or flownets. Too often the logistic problems hinder the use of seepage meters. Alternatively natural solute and stable isotope tracers may be used to obtain the information on the magnitude of subsurface components.

Isotopes have found their most useful application in lake water balance studies in the determination of subsurface inflow and outflow which are difficult to evaluate otherwise. Under favourable circumstances, the evaporation term can also be evaluated, however, accuracy is generally poor. The other parameters are normally directly measured using the conventional methods. Both environmental as well as artificial isotopes are used for water balance studies of lakes.

Environmental Isotope Technique

The estimation of sub-surface water balance components of lakes using environmental isotopes is easily possible in isotopically well-mixed lakes in which evaporation has significantly raised the heavy isotope content with respect to inflow. However, in particular cases, isotopes might provide an independent tool to check water balance computations made by other methods. In this section, the use of environmental stable isotopes and environmental tritium is discussed for evaluating difficult water balance components.

Application of Environmental Stable Isotope

The avenues for application of stable isotope techniques to lake hydrology opened up with the contributions of Craig (1961), Dansgaard (1964) and Craig & Gordon (1965). Dinçer (1968) through his studies on the water balance of three Turkish lakes showed the possibilities of successful application of such techniques to the scientific community. Following his study, several investigators have tried to apply the technique in different environmental setups, from desiccating lakes of semi-arid regions to seasonally stratified lakes of temperate climates. The detailed reviews on the application of environmental isotopes in Lake Hydrology have been provided by Pearson & Coplen (1978) and Gonfiantini (1986).

Two stable isotopes of oxygen (^{18}O and ^{16}O) and of hydrogen (H and D) and the natural variation of their ratio ($^{18}\text{O}/^{16}\text{O}$ and D/H) in water cycle are of great significance. These natural variations, which are mainly due to the influence of temperature, are used to study several hydrological problems. The stable isotopic compositions of water bodies are reported with respect to the international standard, V-SMOW (Vienna - Standard Mean Oceanic Water), and are denoted by δ , expressed in permil (λ).

$$\delta_x (\text{‰}) = \frac{R_{x(\text{Sample})} - R_{x(\text{Standard})}}{R_{x(\text{Standard})}} * 1000 \quad \dots\dots\dots (11)$$

where R stands for the ratio of rarer to the common isotope, subscript x denotes $^{18}\text{O}/^{16}\text{O}$ or D/H.

The variation of δ_x in global precipitation has been studied by several investigators (Dansgaard, 1964; Rozanski *et al.*, 1993). An interesting relationship between δ_{18} and δ_D exists in the precipitation, and based on this a Global Meteoric Water Line (GMWL) has been constructed by Craig (1961). The equation for this line is $\delta_D = 8 \delta_{18} + 10$. While Local Meteoric Water Line (LMWL) constructed for different locations often does not coincide with GMWL, the slope is usually close to 8. Similarly δ_{18} and δ_D of surface water bodies subjected to evaporation will exhibit a linear relationship with a slope of (usually between 3.5 and 6) that will be much less than LMWL. Such lines are called Local Evaporation Lines (LEL).

Established conventional techniques can be used to measure or estimate all the water balance components given in equation (2) within certain error limits (Winter, 1981) except the sub-surface ones. Although seepage meters have been used (Lee, 1977), the variation of subsurface flow in time and space and the practical problems of installation and periodical sampling in deep lakes, restricts their use in this application.

The two unknown variables SS_i and SS_o can be determined, if equation (2) is solved simultaneously with the isotope mass balance equation. The isotope mass-balance equation for the lake is given by

$$\frac{d(\delta_L V)}{dt} = \delta_s S_I + \delta_p P_I + \delta_G SS_I - \delta_{LE} S_O - \delta_{LH} SS_O - \delta_E E_O \quad \dots\dots\dots (12)$$

where t is the time period for which the balance is being computed, other notations with different subscripts are as given in equation (2), and δ -notations are the corresponding isotope ratios. The equation has been simplified by assigning δ_L of the lake as δ_{LE} and δ_{LH} to the surface and sub-surface outflow components, due to stratification in the lake. A separate component should be included in the water balance equation if the water is pumped out from a lake by installing deep tubewells as it may contain the component of lake as well as ground water. In this equation except δ_E , all other components are directly measurable. The factors which may influence the isotope values of individual components are outlined below:

$\delta_{precipitation}$:

As noted in the preceding section, the $\delta^{18}\text{O}$ and δ_D of precipitation exhibits a linear relationship. The standard deviation in δ_p values for independent rain events may be significant depending upon the season and amount of rainfall. The following effects have been recognised as influencing the isotopic composition of precipitation: latitude effect, continental effect, altitude effect, seasonal effect and amount effect. For most hydrological studies, the impact of

individual rainfall events on isotopic composition of the system being investigated is usually negligible, and it is sufficient to have the long-term averaged isotopic composition. But in specific cases, involving small lake basins, short-term averaged data are essential.

The adiabatic lapse rate, of temperature with altitude is about 0.6°C/100m. For δ_{18} , the temperature dependence during adiabatic cooling is about 0.5‰/°C. This means that, there may be about 0.3‰ variation in δ_{18} per 100 meter variation in altitude. Eriksson (1965) has shown that under average conditions, as the vertical flux of water vapour is largely due to turbulence, the altitude effect will be about half of this. Fontes & Olivry (1976) reported a variation of 0.16‰/100m on the slope of Mt. Cameroun. Ramesh & Sarin (1992) have proposed 0.38‰/100m for the Himalayan precipitation; although this is not supported by isotope data on precipitation but inferred through the river waters, it is comparable to the values reported for other regions by Yurtsever and Gat (1981).

In order to consider the seasonal effect, rain water is collected with a device, like a conventional rain gauge or even a rain gauge itself can be used, on each day for a week. The collected rain water is transferred to a 1 litre glass/polythene airtight bottle. After a week, the bottle is stirred well and a 20 ml glass/polythene airtight vial is filled up to the brim and sealed. This sample represents the weekly averaged isotopic composition. The same procedure is repeated for the next three weeks of a month. The remaining sample in the 1 litre bottle can be transferred to another large container and 20 ml of sample is obtained for monthly isotopic composition after stirring the water at the end of the month. Similar procedure is followed for other months. The annual weighted mean isotopic composition (δ_{wp}) can be determined by using the following expression (Yurtsever & Gat, 1981):

$$\delta_{wp} = \frac{\sum P_i \delta_i}{P} \quad \dots \quad (13)$$

where P is the total annual precipitation, P_i and δ_i are the amount of precipitation and corresponding isotopic composition for the time period i, n is the number of weeks or months, as the case may be.

Dinçer (1968) has suggested that the intersection of the LEL and LMWL could be used to estimate the mean isotopic composition of the inflow waters to the lake. One has to consider both time and space variations, to establish the isotopic composition of the groundwater in a given environment. In terrain which is faulted, the aquifers may exhibit different isotopic composition with time. Therefore, the spatial variation within an aquifer is more significant. The groundwater in such terrains could be compared to a set of discrete water parcels which move along a flow path determined by the hydraulic gradient.

δ_{lake}

The isotopic composition of a lake (δ_L) is an important aspect, as there may be both spatial and temporal variations. While it is easy to consider the temporal variations, it is rather difficult to arrive at the true value of δ_L , if mixing is not complete (Gat, 1970). In such cases,

two or three different equations (one each for hypolimnion, mesolimnion and epilimnion) should be used, taking into account eddy diffusion coefficient and hypolimnion entrainment.

$\delta_{Evaporation}$:

Evaporation is the most important process by which the lake undergoes on enrichment in heavier isotope content and gives the lake an isotopic signature different from the local groundwater system. This difference makes it possible to study the water balance, particularly the groundwater component. This natural labelling is a result of fractionation. The mechanism of evaporation is best explained by the Craig and Gordon (1965) model. The simplified evaporation model by Craig & Gordon (1965) can be written as,

$$\delta_E = \frac{(\alpha^+ \delta_l - h\delta_a - \epsilon)}{(1-h + 10^{-3} \Delta\epsilon)} \quad \dots (14)$$

where δ_a is the isotopic content of the atmospheric water vapour, h is the relative humidity normalised to temperature at the interface. The other parameters are related to fractionation. When the water vapour and water body are in thermal equilibrium, the equilibrium fractionation factor (α^+) is given by

$$\alpha^+ = \frac{(R_x)_{water}}{(R_x)_{vapour}} \quad \dots (15)$$

where, R is the ratio of concentrations of heavier to lighter isotopes

The α^+ values, in the case of ^{18}O , decreases monotonously with increasing temperature until it becomes one at the critical temperature of water. However, for D , it decreases to 0.996 at 275EC and then raises to become one at the critical temperature (Gonfiantini, 1986). The α^+ values for water - vapour system at different interface temperature have been determined experimentally (Bottinga & Craig, 1969; Kakiuchi & Matsuo, 1979). But, the most commonly used ones are presented by Majoube, 1971.

$$\ln \alpha^+ [^{18}O] = 1137 * T^{-2} - 0.4156 * T^{-1} - 0.00207 \quad \dots (16a)$$

$$\ln \alpha^+ [D] = 24844 * T^{-2} - 76.248 * T^{-1} + 0.05261 \quad \dots (16b)$$

where T is the temperature in degrees Kelvin. The air temperature is usually considered for the purpose, rather than the lake water surface temperature.

The equilibrium enrichment factor $\epsilon^+ = (\alpha^+ - 1) * 1000 \lambda$. (the α^+ values are always greater than one in temperatures of interest). In lake studies α^* and ϵ^* are used in place of α^+ and ϵ^+ , where α^* is the inverse of α^+ and $\epsilon^* = (1 - \alpha^*) * 1000$, (α^* is less than one).

The kinetic enrichment component $\Delta\epsilon$ of the total (or effective) enrichment factor ϵ , represents the additional enrichment due to the relative difference of resistance coefficients of

isotope molecules, caused during transport of vapour in air. Therefore, the total enrichment (ϵ) is:

$$\epsilon = \epsilon^* + \Delta\epsilon = 1000(1 - \alpha^*) + \Delta\epsilon \quad \dots\dots (17)$$

Several experiments have been conducted to find a practical way to determine the value of $\Delta\epsilon$. Craig and Gordon (1965) suggested a value of 5λ for ^{18}O and 20λ for D. Gat (1970) conducted a multiple pan experiment to measure $\Delta\epsilon$ experimentally, and observed a diurnal variation. They concluded that for the experimental period, $\Delta\epsilon$ values for ^{18}O is between 13 and 16 times (1-h), where h is the relative humidity. However, on the basis of results obtained from wind tunnel experiments, Vogt (1976) proposed the following values, which are used widely:

$$\Delta\epsilon_D \text{‰} = 12.4(1 - h) \quad \dots\dots(18a)$$

$$\Delta\epsilon_{18} \text{‰} = 14.3(1 - h) \quad \dots\dots(18b)$$

Gat (1970) has outlined the greater sensitivity of $\Delta\epsilon_D$ to the evaporation conditions. Zuber (1983) has cautioned that the laboratory experimental data of $\Delta\epsilon_D$ may differ greatly from observed values for a given set of conditions.

Suitable way to determine or by-pass δ_a have been attempted by several investigators. Different methods have been used by various investigators [Dinçer (1968), Gat (1970), Gat(1981), Allison *et al.* (1979), Stolf *et al.* (1979), and Krabbenhoft *et al.* (1990)]. However, keeping in view the merits and demerits of different methods and their suitability, δ_p - δ_a equilibrium assumption method can be used.

δ_p - δ_a equilibrium assumption method:

The relationship between isotopic composition of atmospheric water vapour and precipitation was reported by Zimmermann *et al.* (1967). Craig & Horibe (1967) reported that the isotopic composition of continental water vapour, on rainy days, is in equilibrium with that of precipitation. This implies that there could also be a period when they are not in equilibrium. Zuber (1983) assumed that δ_a is in equilibrium with the mean isotopic content of precipitation, so that

$$\delta_a = \alpha^* \delta_p - \epsilon \quad \dots\dots (19)$$

The study by Krabbenhoft *et al.* (1990) lend support to this assumption. Overall, given the constraints, it seems to be reasonable that if short-term δ_p is available, this method could be used to estimate δ_a .

Application in different types of lakes

The basic isotope mass balance equation can be simplified by taking the type of lake and lake environment into consideration. A lake that is gaining substantial quantity of water from the adjacent aquifers is called as "discharge lakes" or "seepage lakes", and a lake that loses water to adjacent aquifer is called as "recharge lake" or "groundwater lake". Lakes that lose water

only through evaporation is called as "terminal lakes". There are variations to this definitions if a lake is having no inflow or if the inflow compensates the evaporation. A lake that has substantial subsurface inflow and subsurface outflow is called as a "flow-through lake". Gonfiantini (1986) has presented a range of lake conditions and the equations which describe the evolution of the isotopic composition of the lakes, which can be used to compute the isotopic mass balance. Some probable lake types have been presented below:

Case I: Flow-Through Lakes: Water Balance sing Isotopes

In this case , the isotopic water balance equation can be written as,

$$\frac{d(\delta_L V)}{dt} = \delta_{S_i} S_{Ii} + \delta_P P_I + \delta_G S_{S_i} - \delta_L S_{O_i} - \delta_E E_{O_i} - \delta_L S_{S_{O_i}} \quad \dots\dots\dots (20)$$

The above equation has been simplified by assigning the lake δ_l to the outflow components (δ_{is} - δ_{ib}) in equation (12). In this equation except δ_E , all other components are directly measurable.

Zuber (1983) presented the following equation to estimate the ratio of total outflow (O) to evaporation loss, which can then be used with conventional data to estimate the subsurface outflow components:

$$x = \frac{O}{E_o} = \frac{\left(\frac{S_i}{E_o}\right)(\delta_G - \delta_{S_i}) + \left(\frac{P_i}{E_o}\right)(\delta_G - \delta_P) + \delta_E - \delta_G}{(\delta_G - \delta_L)} \quad \dots\dots\dots (21)$$

He suggested that the above equation (21) may further be simplified by assuming δ_G is equal to δ_P , in the absence of groundwater data. Zuber (1983) calculated the subsurface outflow from Lake Chala, Tanzania using oxygen isotope ratios. He reported a value of $3.9 * 10^6$ m³/yr, which was in very good agreement with the value of $4.14 * 10^6$ m³/yr reported by Payne (1983), who used injected artificial tritium for whole body tracing.

In the study of Lake Bagry, Poland also, Zuber (1983) used equation (21) to estimate the subsurface inflow and outflow of the lake. Lake Bagry did not have any surface inflow or surface outflow. The investigator used the long term averages of precipitation, temperature, humidity and evaporation data of the study area. Though oxygen isotopic composition of the lake showed seasonal variation, Zuber used the mean value by integrating the area below δ_l - time curve. The residence time (V/O) calculated by using the results of isotope method compared well with that obtained through electrical analogy approach, which confirmed the validity of the isotope approach, as adopted by Zuber.

Krabbenhof et al. (1990) estimated the groundwater exchange with Lake Sparkling, Wisconsin, USA using the stable isotope mass balance approach using oxygen isotopes. Since the surface inflow to the lake was insignificant and the lake was in a isotopically steady state condition they derived an equation that satisfies the above observations, such that:

$$SS_I = \frac{P_I(\delta_L - \delta_P) + E_o(\delta_E - \delta_L)}{\delta_G - \delta_L} \dots (22)$$

The above equation facilitated them to determine the groundwater inflow independent of groundwater outflow rate. They estimated that the subsurface flow accounted for about 27% of the total inflow and 50% of the total outflow. The results were found comparable to those obtained by a three dimensional flow and solute transport model for the lake - aquifer system.

The Sparkling lake study was later extended by Krabbenhoft et al. (1994), considering the lake to be a Index lake, to study the groundwater components of other nearby lakes with negligible surface components. They studied the groundwater exchange of Lake Crystal, Lake Palette, Lake Big Musky. Assuming isotopic steady state of the lake, they used the groundwater inflow information of Sparkling lake to evaluate the subsurface components of the three lakes. The success of this approach depends very much on the accuracy of the estimation of the water balance components of the Index lake.

LaBaugh et al. (1997) estimated the groundwater component in the water balance of Williams lake, Minnesota, USA using a multi-tracer approach and flow net approach. The investigators found a large difference between the results obtained through isotope approach and flow net approach. They reasoned that the discrepancy arose mainly due to the errors attached to the estimation of evaporation losses and the isotopic composition of the lake evaporates through indirect methods. The investigators reported that both the values obtained for groundwater inflow (seepage) and the range of uncertainty varied with the selection procedure of the groundwater isotopic index. There was a difference of -2λ between the isotopic composition groundwater samples from the seepage zones and the weighted precipitation index. They also considered two different values of lake evaporates as input to the water balance equations. The investigators did not, however, consider the selective recharge process in their study area. They concluded that the oxygen isotope was a better tracer in accounting the exchange of waters of lake Williams with groundwater than the four chemical tracers viz. chloride, sodium, magnesium or dissolved organic carbon that were tried. Further, they suggested that the combined use of hydrogeological and chemical approaches will improve the estimation of the lake - aquifer exchanges.

Case II: Desiccating terminal lakes with no inflow:

For a terminal lake with no inflow, the variation of the isotopic composition could be written as (Gonfiantini, 1986):

$$\frac{d\delta_i}{d \ln f} = \frac{h(\delta - \delta_a) - (\delta_i + 1)(\Delta\epsilon + \frac{\epsilon^+}{\alpha^+})}{1 - h + \Delta\epsilon} \dots (23)$$

where f is the fraction of remaining water. Assuming environmental conditions of evaporation as constant i.e., h, δ_a , $\Delta\epsilon$ and ϵ^+ as constant, equation (25) can be integrated to get

$$\delta_i = \left(\delta_{i_0} - \frac{A}{B} \right) f^B + \frac{A}{B} \dots (24)$$

where, δ_{10} , is the initial lake isotopic composition, f is the fraction of water remaining, and A and B are as follows:

$$A = \frac{h\delta_o + \Delta\epsilon - \frac{\epsilon}{\alpha}}{1 - h + \Delta\epsilon} \quad \dots\dots (24a)$$

$$B = \frac{h - \Delta\epsilon - \frac{\epsilon}{\alpha}}{1 - h + \Delta\epsilon} \quad \dots\dots (24b)$$

From equation (26) it is seen that as f tends to zero (during final stages of evaporation) the term (A/B) is then the final isotopic composition of the lake and it is independent of initial lake isotopic composition.

Case III: Lake with slow and unidirectional change in volume:

For lakes / reservoirs with long residence time and slow and unidirectional change in volume the following equation holds, if we assume the isotopic composition of inflow waters and atmospheric air, the environmental conditions and also the inflow, outflow and evaporation rates to be constant (Gonfiantini, 1986):

$$\delta_L = \left(\delta_{Lo} \frac{\delta_L - AX}{1 + BX} \right) + f^{\frac{-(1+BX)}{(1-X-Y)}} + \frac{\delta_L + AX}{1 + BX} \quad \dots\dots (25)$$

where, $X = \text{Evaporation/Inflow}$ and $Y = \text{Outflow/Inflow}$.

Case IV: Lakes of leaky evaporation pan type:

This type of the lake is similar to the one given under Case III, but with no inflow. i.e., $I = \text{zero}$.

$$\delta_L = \left(\delta_{Lo} - \frac{A}{B} \right) f^Z + \frac{A}{B} \quad \dots\dots (26)$$

where, $Z = E / (E + Q)$, that is the ratio of evaporation to total losses from the lake.

For the water balance study of Quebra Unhas reservoir in Brazil, Stolf et al. (1979) used this approach. Since there was negligible surface inflow and no surface outflow to the reservoir, the investigators used the leaking evaporation an analogy to estimate the evaporation and subsurface outflow from the reservoir. Equation (28) was used with the oxygen isotope data of the pan form which the values of parameters A and B were evaluated, which were then reused in the equation (28) to evaluate the parameter Z of the reservoir. The investigators estimated that -67% of the total losses from reservoir was accounted by evaporation and -33% by subsurface outflow. An independent tracer, viz. chloride ion was also used for the study which indicated that subsurface outflow accounted for -25% of total losses. However,

the investigators reported that the results of two independent tracers are in reasonable agreement, if the error associated to the computations are taken into consideration.

Case V: Lake with constant volume:

For lakes with constant volume i.e. $dv/dt = 0$, we have $I = Q + E$. Assuming $E/I = X$ and $Q/I = 1 - X$, where X is the fraction of water lost to evaporation and $V/I = T$, where T is the mean residence time of the lake we get:

$$\frac{\delta_L}{\delta_i} = -\frac{I}{T} [(1 + BX) \delta_L - \delta_i - AX] \quad \dots (27)$$

Integrating equation (27) and considering evaporation conditions and the isotopic composition of inflow waters to be constant, we get:

$$\delta_L = \left(\frac{\delta_i + AX}{1 + BX} \right) + \left(\delta_{L_0} - \frac{\delta_i + AX}{1 + BX} \right) e^{-(1+BX)\frac{t}{T}} \quad \dots (28)$$

In the above equation when t tends to infinity, δ_L tends to δ_s , a steady state value is then achieved, which is defined by:

$$\delta_s = \frac{\delta_i + AX}{1 + BX} \quad \dots (29)$$

The groundwater components in the water balance of two young artificial lakes in Germany viz. Lake Wiesensee and Lake Waidsee were investigated by Zimmermann (1979). The lake had no surface inflow or outflow. Zimmermann considered the lakes to be of constant volume, although there was a 6% decrease in the volume. During the investigation period from 1970 - 1974, the lakes showed a continuous enrichment in the heavy isotope (δD) due to evaporation, on which the seasonal variations in the precipitation were superimposed. The investigator correlated $\ln(\delta_i^s - \delta_i)$ with time (t), and selected the value of δ_i^s by trial basis using equation (30). The values of parameters A and B were computed using the field humidity and δ_a values. In the study, the results attached to the estimated evaporation and subsurface inflow and outflow were of the order of 30%. The investigator, however, concluded that the stable isotope method is suitable for computing subsurface components as there is no other method to evaluate these components with comparable accuracy. Zuber (1983) using equation (23) reinterpreted the data on Lake Waidsee presented by Zimmermann. Based on the discrepancy in the value obtained for X , Zuber questioned the results of Zimmermann (1979). Further, Zuber showed that the assumptions involved in the calculations influences the results of the parameter X .

Case VI: Terminal lakes where evaporation compensates inflow:

When the evaporation loss from a lake exactly compensates the inflow to the lake, then the steady state isotopic composition of the lake could be determined from (Gonfiantini, 1986):

$$\delta_s^T = \alpha^+ \delta_i (1-h + \Delta\varepsilon) + \alpha^+ h \delta_a + \alpha^+ \Delta\varepsilon + \varepsilon^+ \quad \dots\dots (30)$$

δ_s^T is then the maximum enrichment in heavy isotope which can be reached by a constant volume lake. Correction terms have to be included in equation (30) as in certain terminal lakes the salt effects may dominate.

Dincer (1968) used the presence of Lake Burdur, in Turkey which is of terminal type, as an index lake for molecular exchange in the region where two fresh water lakes viz. Lake Egridir and Lake Beysehir were situated. He used the following equation to estimate the water balance of the fresh water lakes:

$$\frac{d(\delta_L V)}{dt} = I(\delta_i - \delta_L) - E \left(\frac{\delta_L - \delta_L^T}{\alpha(1-h)} + \delta_i^T - \delta_L \right) \quad \dots\dots (31)$$

where, I is the inflow rate, E is the evaporation rate, δ is the isotopic composition and subscripts I and L denote inflow and mean lake values respectively. The superscript T denotes the data pertaining to terminal lake (index lake). The investigator stressed the sensitive nature of the isotopic composition of lake water in using the index lake method. An error of 1% in estimated mean isotopic composition resulted in 5% change in the inflow. However, the method was found less sensitive to the fractionation factors, relative humidity and kinetic effects.

Choice of isotope: ^{18}O or D

While both isotopes can be used for calculating the lake water balance, there may be a slight difference in results arrived by using the two isotopes. Zimmermann and Ehhalt (1970), Friedman *et al.* (1976) and Allison *et al.* (1979) have advocated the use of D rather than ^{18}O , because of the smaller influence of the poorly known kinetic separation factor ($\Delta\varepsilon$) in the case of D. The kinetic separation in diffusion controlled transport region affects ^{18}O relatively more than D. Yurtsever & Gat (1981) have indicated that because of the low sample volume required for D/H analysis by most laboratories, investigators prefer to test for δ_D as acquiring atmospheric vapour samples at high altitudes could be difficult. On the other hand, Gat (1970) showed that the kinetic term is better understood in the case of ^{18}O than in the case of D, and preferred to use ^{18}O . The laboratory determined $\Delta\varepsilon_D$ values differ greatly from the observed field values (Zuber, 1983).

Application of Environmental Tritium

If a lake is considered isotopically well mixed then, the tritium content of the lake outflow (both surface and subsurface) is equal to that of the lake; usually this is a case for a water body wherein mixing is fast relative to the residence time of water. Assuming that the lake is at hydrological steady state, the tritium balance corresponding to the basic water balance of a lake (eqn. 2) can be written as:

$$VdT_L/dt = T_S S_I + T_{SS} SS_I + T_P P_I - T_L S_O - T_L SS_O - T_E E_O - T_L \lambda.V \quad \dots\dots(32)$$

where T represents the respective tritium content of the water balance components in eqn. 4

(in tritium units, TU) and the last term being the decay term (where λ is the radioactive decay constant of tritium = 0.0557 a^{-1}). T_E given in the above equation can be defined as,

$$T_E = (\alpha \cdot T_S - h \cdot T_A) / \{ \alpha_K (1-h) \} \quad \dots \dots (33)$$

where, h is the relative humidity, T_A is the tritium content of water vapour over the lake, α_K is the equilibrium fractionation factor for tritium between liquid and vapour phases (=0.9), T_L is the tritium content of the lake water and α_K is the kinetic fractionation factor (=1.1). The change produced by evaporation and molecular exchange can be positive or negative, depending on the relative tritium content of lake water and precipitation. If the Equation 35 is divided by volume V, we get

$$dT_L / dt = T_S K_{SI} + T_{SS} K_{SSI} + T_P K_{PI} - T_L K_{SO} - T_L K_{SSO} - T_E K_{EO} - T_L \cdot \lambda. \quad \dots \dots (34)$$

where K represents the ratio of the volume of water that enters or leaves the water in different forms to that of the lake volume. For a steady state system (i.e., no increase or decrease of volume with time), the water retention time (τ) is defined as the inverse of some of either K_{SI} , K_{SSI} and K_{PI} or of K_{SO} , K_{SSO} and K_{EO} . Since, the lake volume is assumed to be a constant, from the Equation 4, it can be obtained that;

$$K_{SI} + K_{SSI} + K_{PI} = K_{SO} + K_{SSO} + K_{EO} \quad \dots \dots (35)$$

In addition, the tritium content in water vapour over the lake may be assumed to be in equilibrium with local meteoric water ($T_A = \alpha T_P$) which may be considered a reasonable assumption in case of a mid continental temperate zone lake. If we assume piston flow of groundwater,

$$T_{G,t} = T_{P,t-\tau} e^{-\lambda\tau} \quad \dots \dots (36)$$

where $T_{G,t}$ is the tritium content of the groundwater at any time t and $T_{P,t-\tau}$ is the tritium content of the precipitation at a time t- τ (where τ is the residence time of ground water). Now discretising Equation 35 in a time step of one year, we get,

$$T_{L,t} - T_{L,t-1} = T_{p,t} \cdot K_{PI} + T_{s,t} \cdot K_{SO} + T_{p,t-\tau} \cdot e^{-\lambda\tau} \cdot K_{SSI} - [\{ -\alpha T_{L,t} - h \cdot \alpha \cdot T_{p,t} \} / \{ \alpha_K (1-h) \}] \cdot K_{EO} - T_{L,t} \cdot K_{SSO} - T_{L,t} \cdot K_{SO} - T_{L,t} \cdot \lambda \quad \dots \dots (37)$$

In case, the lake catchment is not comparatively large in size, $T_{S,t} = T_{P,t}$, i.e., the stream or channel water undergo negligible change in their tritium content during their travel from the catchment to the lake in the form of atmospheric exchange. Therefore, the above equation can be arranged to get,

$$T_{L,t} \cdot [1 + \alpha \cdot K_{EO} / \{ \alpha_K (1-h) \} + K_{SSO} + K_{SO} + \lambda] = T_{L,t-1} + T_{p,t-\tau} \cdot e^{-\lambda\tau} \cdot K_{SSI} + T_{p,t} [K_{PI} + K_{SI} + \{ h \cdot \alpha \cdot K_{EO} / \alpha_K (1-h) \}] \quad \dots \dots (38)$$

Using Equation 38, the most difficult water balance components, namely K_{SSO} and K_{SSI} , in the above equation are expressed in terms of T as, $K_{SSO} = (1/T) - K_{SO} - K_{EO}$ and $K_{SSI} = (1/T) - K_{SI} - K_{PI}$ respectively. Therefore,

$$T_{L,t} [1 - K_{EO} \{1 - \alpha / \{\alpha_k (1-h)\}\} + (1/T) + \lambda] = T_{L,t-1} + T_{p,t} [K_{PI} + K_{SI} + h \cdot \alpha \cdot K_{EO} / \{\alpha_k (1-h)\}] + T_{p,t-\tau} \cdot e^{-\lambda\tau} \{(1/T) - K_{SI} - K_{PI}\} \quad \dots\dots\dots (39)$$

The tritium contents of precipitation, spring water, surface runoff, lake water (at different depths and locations) and the water which is either drained or flowout in the form of streams are measured at different times. The tritium contents in the vapours can be obtained by using Equation 34. By using the above equations and the different measured values of tritium contents, not only the values of SS_t and SS_o can be measured but the residence time of groundwater as well as residence time of water in the lake can be computed which may be of great importance to understand the hydrological behaviour of the lake and for its management.

Poorly mixed lakes

The study of water balance of a poorly mixed lake may be difficult, however with a dense network of sampling and using a properly weighted mean value to solve the mass balance equations with a reasonable degree of accuracy may be attempted. Fontes et al. (1970) utilised the systematic variation in the isotopic composition in a shallow lake, disregarding mixing and considered each sample as representative of different stages of evaporation of the lake. While studying a large swamp area, Dincer et al. (1979) combined isotopic and water balance equations with the salt balance equations. They suggested that the method allowed the calculation of the ratio of evaporation to evapotranspiration.

Gonfiantini et al. (1962) investigated the water balance of lake Bracciano, Italy using isotope approach. They suggested that 90% of the error in the calculated parameter X, was due to the errors associated with humidity and isotopic composition of the atmospheric air. and it is difficult to measure or estimate these parameters with reasonable accuracy. Zimmermann and Ehhalt (1970) used hydrogen isotope ratio to study the subsurface components of Lake Neuseidl, Austria, which is a shallow lake with poor mixing. The investigators neglected the kinetic enrichment effect in their computations. They concluded that large errors associated with δ_E resulted the determination of evaporation rate not better than 50% accuracy.

The case of the seasonal stratification is not uncommon in temperate lakes, in tropical or subtropical mountainous lakes and permanent stratification in deep lakes. Under stratified conditions the upper portion of the lake, called epilimnion, is enriched in heavier isotopes (by evaporation) compared to the bottom waters, called hypolimnion. In the water balance study of Lake Tiberias, which is a stratified lake, Gat (1970) used an approach to include the effect of isotope exchange between the epilimnion and hypolimnion (eddy diffusion) and also the increase in the volume of the epilimnion (hypolimnion entrainment) due to the subsidence of thermocline. Gat (1995) suggested that the effect due to exchange with hypolimnion is usually not significant, if exchange is the only process by which the isotopic composition of the hypolimnion changes. However, the effect of hypolimnion entrainment will generally be significant.

Artificial Radioisotope Techniques

Tritium can be used for water balance studies of lakes provided the size of the lake is not much and the lake water is not used for drinking purpose. If tritiated water is injected to raise

the tritium concentration of the whole lake about 100 times above the natural level, then theoretically, lake's tritium content should decrease with time (Gilath and Gonfiantini, 1983). If the tritium concentration of inflow water and atmospheric moisture are much smaller than that of the spiked lake, the decrease in the tritium concentration can be written in terms of water budget components assuming that the lake is homogenous and the water with some tritium content flows out from the lake

$$C_t = C_0 \exp - (t/V) [I - E + E/\alpha(1-h)] - \lambda t \quad \dots\dots(40)$$

where,

- V = volume of the lake,
- I = inflow rate to lake (surface runoff + precipitation + ground water),
- E = net evaporation rate,
- h = mean relative humidity,
- α = tritium fractionation factor at evaporation,
- λ = tritium decay constant (0.0558 a^{-1}),
- C_0, C_t = tritium concentrations of lake at time 0 and at time t.

Therefore, by knowing the concentration of tritium tracer injected at different times, the unknown parameters of the water balance equation can be evaluated. Tritiated water is injected in the lake and is allowed to mix so that the tritium concentration is homogenised in the entire lake. Lake water samples are collected at regular intervals and tritium concentration is measured in each sample using a liquid scintillation spectrometer. Using these values, the required parameters like subsurface inflow and outflow can be computed at any particular time.

As is obvious, this method can be used only for small lakes as in large lakes, a huge quantity of tritium will be required to raise the tritium concentration of the lake to a desired level. Further, this method requires an accurate evaluation of the evaporation and relative humidity if being used for estimating groundwater inflow/outflow components. The tracer in fact interacts with atmospheric moisture and necessary corrections should be made to compensate this effect.

Another way to use the radioisotopes for the measurement of inflow into the lake through visible gullies, nalas, and springs is to mix either a gamma ray radioisotope like Br-82 or tritium at a suitable point in the inflowing channel. In order to compute the subsurface inflow/outflow a special tracer probe is used which comprises of a tracer injection and mixing device, inflatable rubber packers to isolate a portion of the bore-hole and a built-in scintillation detector assembly.

UNCERTAINTIES IN THE ESTIMATION OF WATER BALANCE COMPONENTS

The overall accuracy of the water budget method depends on the accuracy of each flow component used in the budget. Winter (1981) observed that water budgets of lakes that are determined without estimates of errors could be misleading. In case of certain components of the water budget, the amount of error could not be correctly estimated, due to the nature of the estimation methods. This could be achieved based on available information and therefore,

the errors in the estimation of those components are - at best - judged. For detail see the Winter (1981) paper.

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