Methods for the quantification of evaporation from lakes

prepared for the World Meteorological Organization's
Commission for Hydrology

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CONTENTS

LIST OF SYMBOLS

A	available energy
a	regression parameter (gradient)
A_s	area of the water surface
b	regression parameter (intercept)
C	mass transfer coefficient
c	specific heat of water
c_a	specific heat of air
c_j	intercept of the linear regression between departures of the daily evaporation from the mean daily evaporation and the daily meteorological variable and the mean daily meteorological variable for the j^{th} month of the year
c_p	specific heat of air at constant pressure
E	evaporation rate from a water body
E_i	estimated daily evaporation on day i
$\overline{E}_m$	monthly mean daily evaporation of month m
E_p	evaporation rate of an evaporation pan
e	vapour pressure of the air at the reference height.
e_s^*	saturated vapour pressure of the air at the water surface temperature
e_p	saturated vapour pressure of the air at pan surface temperature
e_a^*	saturated vapour pressure of the air at air temperature
$f(u)$	wind function of wind speed u
F_{in}	heat fluxes associated with inflows
F_{out}	heat fluxes associated with outflows
F_P	heat inflow associated with precipitation
G	heat conduction occurring between the water and its substrate
G_s	soil heat flux
g_j	slope of the linear regression between departures of the daily evaporation from the mean daily evaporation and the daily meteorological variable and the mean daily meteorological variable for the j^{th} month of the year
H	flux of sensible heat
K	empirical constant
$K\downarrow$	incoming short-wave (solar) radiation
k	von Karman's constant
$L\downarrow$	incoming long-wave (thermal) radiation
$L\uparrow$	outgoing long-wave (thermal) radiation
L_e	effective length of the water body (km)
m	ratio of observed sunshine hours to total possible hours of sunshine in a day

N	change in the energy storage in the water
P	atmospheric pressure
P	mean rate of precipitation over a sampling period
p	cloudiness factor
Q_{ri}	surface inflow rate
Q_{ro}	surface outflow rate
Q_{gi}	groundwater and seepage inflow rate
Q_{go}	groundwater and seepage outflow rate
R_n	net input of radiation at the surface of the water body
R_n'	net radiation in units of equivalent depth of water
R_{n*}	net radiation when the water temperature is equal to the wet bulb temperature
r_a	aerodynamic resistance
r_s	bulk surface resistance
S	incident short-wave radiation
S_d	incoming diffuse solar radiation
S_n	net short-wave radiation at the surface of the water body
S_t	solar radiation incident at the top of the atmosphere
S_0	incoming direct solar radiation
T_a	air temperature at a reference height
T_b	arbitrary base temperature
T_e	equilibrium temperature
T_n	wet-bulb temperature
T_s	temperature of the water at the surface
$T_{w,i}$	water temperature at the end of the current day
$T_{w,i-1}$	water temperature at the end of the previous day
t	length of the model time step
u_z	wind speed at z m above the surface
V	water stored in water body
V_i	value of the meteorological variable on day *i*
$\bar{V}_m$	mean daily value of the meteorological variable of month *m*
z	water depth
z_{mix}	summer mixing depth of the water body (*m*)
z_o	roughness length
z_r	height of the meteorological observations above the surface
α_S	albedo for short wave radiation
α_L	albedo for long-wave radiation
α	Priestley-Taylor coefficient
β	Bowen ratio

ε clear-sky atmospheric emissivity

ε_m ratio of the molecular weight for water to that for dry air

Δ slope of the saturated vapour pressure-temperature curve at air temperature

Δt time step

ΔT_w change in spatially averaged temperature of the water body

Δ_w slope of the temperature-saturation water vapour curve at the wet bulb temperature (kPa $^oC^{-1}$)

σ Stefan-Boltzmann constant

ϕ atmospheric pressure

ρ density of water

ρ_a density of air

τ time constant

γ psychometric constant

λ latent heat of vaporisation

λE flux of latent heat (evaporation rate in energy flux units)

1 INTRODUCTION

At the twelfth meeting of the UN World Meteorological Organization Commission for Hydrology in October 2004 one of the work items proposed for the following inter-sessional period was the identification of methods of assessing evaporative water losses from reservoirs and lakes. This report has been produced as a result of that request.

A wide variety of methods for estimating open water evaporation has been reported in the literature and used in practice. They can be categorised into major types of approach - pan evaporation, mass balance, energy budget models, bulk transfer models, combination models, equilibrium temperature methods and empirical approaches.

The form of this report is a description of the major methods for determining lake evaporation, using both aspects of measurement and of calculation. Within the description of each method comment is made about the general applicability of the approach with respect to data needs and relevance of results, together with key experience from applications.

An important subsequent section of the report is a tabulation of values derived for lake and reservoir evaporation by a range of methods and from a range of sources. These are grouped by WMO Region and offer numerical values assessed under particular conditions.

2 FACTORS AFFECTING EVAPORATION RATES FROM LAKES

The estimation of evaporation from lakes and reservoirs is not a simple matter as there are a number of factors that can affect the evaporation rates, notably the climate and physiography of the water body and its surroundings. In addition, the water has the potential to transport stored heat within the water body itself and into and out of it. The rate of evaporation is, however, fundamentally controlled by the available energy and the ease with which water vapour diffuses into the atmosphere.

The available energy is a combination of the net radiation at the lake's surface and the amount of heat stored in the water. The net radiation, R_n, that is, the amount of energy captured by the lake, is normally the dominant factor controlling the annual evaporation rate. It is the difference between the downward, $K\downarrow$, and reflected, $(1-\alpha_s)K\uparrow$, global solar radiations, where α_s is the albedo or reflection coefficient, plus the difference between the upward, $L\uparrow$, and downward, $L\downarrow$, longwave radiation.

Thus, the albedo is an important characteristic of a lake. There are a number of factors which affect the albedo, for example the proportion of direct to diffuse downward solar radiation, the turbidity of the water and, in the case of shallow lakes, the reflection coefficient of the lake bottom. The proportion of direct to diffuse solar radiation matters because the albedo is a function of the elevation angle of the incoming solar radiation: the values of albedo are relatively constant, at a low value, at elevations greater than 37˚. However as the elevation angle decreases below this value, the albedo of water increases exponentially (see for example Davies, 1972). Thus, on a cloudy day when the downward solar radiation is entirely diffuse, the albedo of the lake, averaged over a day, will have a higher albedo than on a sunny day when the direct downward solar radiation will be dominant. Although a value of 0.08 is commonly used for the albedo of water, there are a number of factors that can significantly alter this, for example due to the possibility of reflectance from the bottom, differences in the waves on the surface, and differences in the amount and type of suspended particles, all of which will tend to increase the albedo.

The exchange of radiant energy between the lake surface and the atmosphere in the form of long-wave (thermal) radiation is significant. The downward longwave radiation is related to the temperature and humidity structure of the atmosphere and the cloud cover because its dominant source is the water vapour molecules in the atmosphere. The reflectivity of the water surface is normally around 0.02; however, it is often taken to have a value of zero, which does not introduce any significant error. The upward longwave radiation emitted from the surface of the lake and can be calculated from the Stefan-Boltzman law:

$$L^{\uparrow} = \varepsilon\sigma(T_s + 273.13)^4 \quad (1)$$

where ε is the effective emissivity, σ is the Stefan-Boltzman constant (4.9×10^{-9} MJ m^{-2} K^{-4} d^{-1}) and T_s is the temperature of the water at the surface, in degrees Celsius. Richter (1988), on the basis of measurements, concluded that the average value of the effective emissivity is 0.98. This is supported by more recent work, e.g. Ogawa et al. (2002).

Seasonal variations in the evaporation rate can be significantly affected by the heat storage capacity of the water body which is, to a large extent, determined by its depth. In higher latitudes, where there is a strong seasonal variation in the sun's elevation at noon, the increasing incoming solar radiation serves to warm the water body during the spring and early summer. During the autumn and early winter, as the incoming solar radiation decreases, the water body cools as the stored energy is released. The result is that the evaporation rate can be de-coupled in time from the net radiation. It is generally considered that the effect can be ignored for water bodies with a depth less than 0.5 m. and that the effect reaches a maximum (i.e. the seasonal evaporation ceases to change) once the depth increases beyond 4.5 m (because little of the incoming solar radiation penetrates below this depth).

The situation becomes more complex if a water body becomes thermally stratified. Stratification occurs in large, deep water bodies (at mid and high latitudes) and may accentuate the time lag between the net radiation and the evaporation rate. The temperature dependence of the density of water is a key factor (the maximum density occurs at a temperature of 4^{o} C).

During the early spring, most large, temperate lakes exhibit a nearly uniform temperature distribution with depth (homothermal conditions). As the year progresses and the weather warms up, the water body receives heat at an increasingly rapid rate. Initially, the water body remains homothermal because the heat that is received at the surface layers is transported to deeper layers by wind-induced currents and turbulence. As the rate of heating continues to increase, it begins to exceed the rate of transfer to deeper layers with the result that the temperature of the surface layers increases faster than those of the deeper layers. As the heating continues, a point of inflection develops in the temperature depth profile and a well-mixed upper layer (the epilimnion), with relatively intense gradients at its bottom boundary is formed. The plane of maximum temperature gradient is known as the thermocline. During the remainder of the heating period, the thermocline slowly descends into the lake. Once

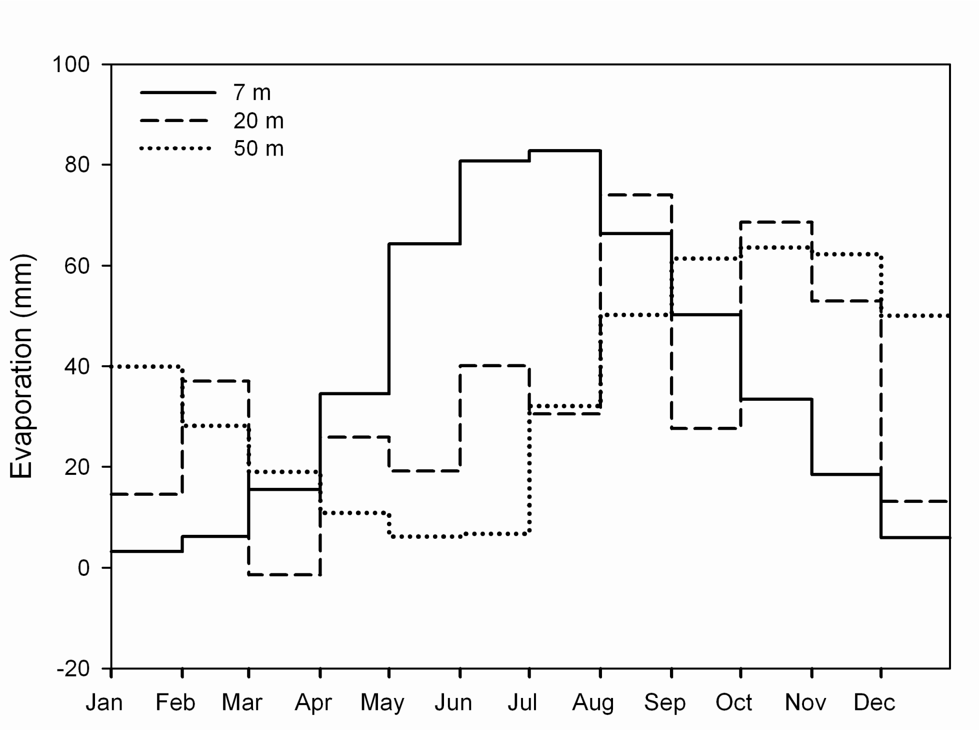

Figure 1 Simulated monthly average evaporation for three water depths (temperate climate).

a thermocline has formed, the deeper regions of the lake are relatively uninfluenced by changes in surface conditions. The maximum thickness of the epilimnion is dominantly a function of the surface area of the water body and the climate.

In the autumn, after the water body has attained its maximum heat content, the thermocline moves down rapidly into the deeper layers of the lake, often referred to as turnover. This is because the wind mixing is augmented by convective mixing due to surface cooling (resulting in an increase in density so that the water sinks). The thermocline continues to move down rapidly as the well-mixed upper layers cool further, until the whole water body again attains homothermal conditions.

A 'reverse' stratification can be created in winter, especially in continental climates, but the cool layer is much thinner than the epilimnion of summer. Sufficient cooling may permit the water body to freeze over, whilst retaining the temperature of the deeper water in the range of 2-4ºC. If the minimum (winter) temperature of the water body is greater than 4ºC then there is only one turnover (in the autumn). Large water bodies that are rarely stratified are generally tropical with high temperatures.

The net result of the heat storage is that water temperatures are lower than air temperatures during the summer and vice versa during the winter. Thus, the evaporation rates from large deep water bodies may be higher in winter than in summer. This is illustrated by Figure1, which shows the simulated mean monthly evaporation, calculated using the model of Mironov *et al*. (2003) for a location near London. This clearly shows that, as the water depth increases, the maximum evaporation shifts from about a month after the summer solstice to four months

The heat transferred into a lake by inflows and outflows of water may be a significant factor in the energy budget of the lake and thus the evaporation rate. Possible Inflow includes seepage from groundwater bodies, changes in bank storage, rivers flowing into the lake and land surface run off; whilst outflow includes rivers, controlled withdrawals (reservoirs) and leakage to groundwater.

In the following chapters specific approaches to quantifying lake evaporation are presented and discussed in the light of these factors.

3 PAN EVAPORATION

Pans have been used to estimate evaporation for over two hundred years. They have an intuitive appeal of apparent simplicity but it is difficult to reliably use data from pans except in specific circumstances. Hounam (1973) carried out a review of methods for estimating lake evaporation from measurements of pan evaporation and much of the following is drawn from this source.

A pan that has found wide use around the world is the US Class A pan. This is a circular galvanized iron tank with a diameter of 1.21 m and is 0.255 m deep and the interior is usually painted black. It is mounted on an open wood frame so that air circulates round and under the pan. The water level is kept about 50 mm below the rim. The level is measured daily using a hook gauge and allowance must be made for any rainfall recorded in the previous 24 hours. In the standard setup, a thermometer measures the water temperature and a three-cup anemometer measures the wind speed 150 mm above the pan rim.

Another pan that has found world-wide use is the USSR GGI-3000 pan. This is a cylindrical tank with a diameter of 0.618 m and is 0.6 m deep at the walls and 0.685 m deep at the centre and painted white. The pan is sunk in the ground with the rim approximately 75 mm above the surface.

Measurements of pan evaporation can rarely be used directly as estimates of evaporation from a water body because of the differences in size between the pan and the water body and, possibly, differences in the overlying air. Winter (1981) suggests that the direct use of data from pans located some distance away from the water body can result in significant errors. The two main approaches to estimating the evaporation of a water body from pan measurements are the use of pan coefficients and pan conversions.

Pan coefficients are simply the ratio of the water body evaporation to pan evaporation. Numerous coefficients have been reported in the literature, although most apply to the US Class A pan. The coefficients are generally specific to the pan type, its location and the nature of the water body and they may, in addition, vary with time. This variation with time takes account of the lag, due to heat storage, in large water bodies whereas the pans are too small for any lag effect. Lapworth (1965), for example, compared the evaporation from a 16 hectare reservoir near London, UK, calculated using the water balance, with that of a US Class A pan over a seven year period: for annual totals, the pan coefficient was 0.7, with a strong monthly variation in the pan coefficients which varied between 0.47 and 1.18. Winter (1981) suggested errors of 10% for measurement errors, 50% for application of pan coefficients and 15% for areal averaging.

Pan conversions are achieved by taking the ratio of the bulk mass transfer equations of the lake and the pan:

$$\overline{E} = K \frac{(\overline{e}_s^* - \overline{e}\ )}{(\overline{e}_p - \overline{e})} \overline{E}_p \tag{2}$$

where $\overline{E}$ is the mean evaporation rate from the water body, $\overline{E}_p$ is the mean evaporation rate of the pan, K is an empirical constant, $\overline{e}_s^*$ is the mean saturated vapour pressure of the air at the water surface temperature, $\overline{e}_p$ mean saturation vapour pressure at pan surface temperature and $\overline{e}$ is the mean vapour pressure of the air at reference height. This method is, however, dependent on knowing the surface temperatures of the lake and the pan, which is rarely practical. In addition, an empirical coefficient is still required which has to be determined for the specific situation.

The evaporation from a pan will be enhanced if it is surrounded by a dry surface: this is called the oasis effect. Energy from the surrounding surface will be transferred horizontally from the dry surface and provides extra energy for evaporation of water in the pan. In addition, specific pans differ due to their different constructions. The US Class A pan suffers from the disadvantage that the sides are exposed to the sun with the result that it reaches a higher temperature (and consequently increased evaporation) than pans sunk in the ground. Conversely, sunken pans can sometimes overestimate evaporation due to heat transfer from the surrounding soil. Leakage is also much more difficult to detect in sunken pans and they are vulnerable to splash in and splash out. In times of hot weather, wildlife may use the pans as sources of drinking water. Attempts to overcome this by covering the pans with mesh have resulted in significant reductions in evaporation.

Despite their apparent simplicity, all pans need to be carefully maintained. The water level must be kept close to the prescribed level and regular cleaning and periodic re-painting are necessary. The siting of the pan can have a major impact on the measurements: a pan sited on bare soil may, for example, record higher evaporation rates than one sited on grass because the air moving over the pan will tend to be drier (Allen *et al.*, 1998).

4 MASS BALANCE

The mass balance method of measuring open water evaporation is simple in principle, evaporation being calculated as the change in volume of water stored and the difference between inflow and outflow, i.e.

$$E = P + \frac{(Q_{ri} + Q_{gi}) - (Q_{ro} + Q_{go}) - dV/dt}{A_s} \tag{3}$$

where E is the evaporation rate from the water body, P is the mean rate of precipitation over the sampling period, Q_{ri} is the surface inflow rate, Q_{ro} is the surface outflow rate, Q_{gi} is the groundwater and seepage inflow rate, Q_{go} is the groundwater and seepage outflow rate, V is the water stored and A_s is the surface area.

The relative importance of the terms depends on the hydrological and physiographical setting. The feasibility of determining evaporation depends primarily on the relative magnitudes of the terms: it is difficult to obtain a reliable estimate whenever the evaporation is of the same order of magnitude as the errors inherent in the measurements. The method is therefore unsuited to water bodies with large flow rates.

Depending on the size of the lake, one or more raingauges are required to estimate precipitation. In most cases, precipitation is estimated from gauges on the surrounding land. Differences in the properties of the land and water surfaces, in particular through the partition of the incoming energy by the land surface between the latent heat flux and the sensible heat flux into the atmosphere, may result in a large water body having a distinct micro-climate with the result that the precipitation may be appreciably different from that on the land.

The surface outflow of larger water bodies may be measured to a reasonable accuracy but surface inflow is generally known less accurately as, commonly, only the major water courses are measured. If flow is seasonal, surface inflow during the summer may be small enough in comparison with the evaporation for the evaporation to be calculated with reasonable accuracy during this period. For example Gangopaghaya *et al.* (1966) pointed out that, in the case of the Franklin D. Roosevelt Lake on the Colorado River, the errors in measuring the outflow would result in an uncertainty that was ten times the amount of evaporation. In the case of a Lake Hefner study, the measured inflows and outflows were 10% greater than the evaporation over the 16 month period (Harbeck *et al.,* 1954). The volumes of groundwater and seepage inflow and outflow are usually unknown. In some situations it may be possible to assume that these are negligible. A further complication can arise if bank or groundwater storage occurs. Gangopaghaya *et al.* (1966) have pointed out that this can increase the total storage capacity by as much as 12% with the consequent error in the estimation of evaporation if this is not taken into account. Water level recorders and a reliable depth-storage relationship are required. The use of more than one water level recorder should be used for large lakes, in order to avoid errors due to seiches (standing waves) and wind and pressure conditions.

An example of a very detailed analysis of the mass balance of a lake is provided in the work of Harbeck *et al.* (1954) on Lake Hefner (13.8 km^2 surface area near Oklahoma, USA) over a 16 month period. They estimated that the error in the monthly estimates of evaporation was less than five percent, which must be taken as the highest accuracy that is likely to be achieved using this method. In the UK, Lapworth (1965) estimated the evaporation from a 16 hectare man-made reservoir near London over a period of seven years. An assumption was made that seepages were negligible, there were no inflows and outflows (except for a single lowering) during the period of the study, and the rainfall inputs were measured with a raingauge at the site. An assessment of the errors suggested that the estimated evaporation was within 5% of the true value.

In view of the possible errors, the mass balance method is unlikely to be applicable over periods shorter than a month.

5 ENERGY BUDGET

In this approach evaporation from a water body is estimated as the energy component required to close the energy budget when all the remaining components of the budget of the water body are known, that is, it is the residual component. The energy associated with evaporation is of two categories; first, the heat required to convert liquid water into water vapour (vaporisation) and, second, the energy of the water vapour molecules carried from the water body (advection). The latent heat of vaporisation ranges between 2.5 and 2.4 MJ kg^{-1} for liquid water between 0°C and 40°C.

The energy budget of a water body is given by

$$N = S(1-\alpha_s) + L_{\downarrow}(1-\alpha_L) - L_{\uparrow} - \lambda E - c(T_s - T_b)E - H + F_{in} - F_{out} + F_p - G \qquad (4)$$

where N is the change in the energy storage in the water, S and $L\downarrow$ are the incident short and long-wave radiation respectively, and α_S and α_L are the albedos (reflectivities) for short and long-wave radiation, $L\uparrow$ is the long-wave radiative loss from the water, λE is the flux of latent heat (evaporation rate in energy flux units; λ is the latent heat of vaporisation and E is the evaporation rate in mass units), c is the specific heat of water, T_s and T_b are the temperature of the evaporated water and an arbitrary base temperature respectively, H is the flux of sensible heat (the energy used in warming the atmosphere in contact with the water which is then convected upwards), F_{in} and F_{out} are the heat fluxes associated with water flows in and out of the water body, F_P is the heat inflow associated with precipitation, and G is the heat conduction occurring between the water and its substrate. All the energy components are in units of energy per unit surface area of the water.

The three radiation terms together give the net radiation, R_n such that rewriting equation (4) gives

$$\lambda E + c(T_s - T_b)E = R_n - H + N + F_{in} - F_{out} + F_P - G \qquad (5)$$

Usually the sensible heat term (the amount of energy directly warming the air) cannot be readily determined and it is eliminated from equation (5) through use of the Bowen ratio, β which is defined as the ratio between the sensible and latent heat fluxes. It can be expressed thus

$$\beta = \frac{H}{\lambda E} = \frac{c_p \phi (T_s - T_a)}{\varepsilon_m \lambda (e_s^* - e)} \qquad (6)$$

where c_p is the specific heat of air at constant pressure, ϕ is the atmospheric pressure, T_s and T_a are the temperatures of the water surface and the air at a reference height, ε_m is the ratio of the molecular weight of water to that of dry air, and e_s^* and e are the saturated vapour pressure of the air at the water surface temperature and the vapour pressure of the air at the reference height. The ratio $c_p\phi / \varepsilon\lambda \equiv \gamma$ is also known as the psychometric coefficient. More detail on the Bowen ratio and many other aspects concerning evaporation physics are given in Brutsaert (1982).

From Equation 6, $H = \beta\lambda E$ which, when substituted into equation (5), gives the evaporation rate,

$$E = \frac{R_n + N + F_{in} - F_{out} + F_P - G}{\lambda(1+\beta) + c(T_s - T_b)} \tag{7}$$

The second term in the denominator represents a correction term for the difference between the temperature of the evaporated water and an arbitrary base temperature.

By suitable selection of averaging period it is sometimes possible to neglect the F_{in}, F_{out} and G terms. Indeed, it is usually the case that the energy content of a water body is chiefly governed by the exchange of energy through the surface, rather than the inflows, including precipitation, and outflows and the water-substrate interface (Henderson-Sellers, 1986). This would certainly be the case if the volumes of water flowing in and out of the water body are small compared to the overall volume, or the temperatures are close to the temperature of the water body. Therefore, the last four terms in the numerator of equation (7) can often be neglected if $T_b = T_s$ and the energy budget is then given by

$$E = \frac{R_n + N}{\lambda(1+\beta)} \tag{8}$$

This is sometimes referred to as the reduced energy budget equation (Simon and Mero, 1985; Assouline and Mahrer, 1993; dos Reis and Dias, 1998).

The energy budget method consists of determining, by measurement or estimation, the different terms in either equation (7) or (8). After the direct measurement of evaporation, the energy budget is widely considered to be the most accurate method of estimating evaporation (Assouline and Mahrer, 1993 quoting Hoy and Stephens, 1977). As such it is often used as a reference method against which other methods are validated or calibrated. The accuracy depends upon the timescale and size of the water body. Because of the heat storage, the larger the water body, the longer the time interval required between measurements of the temperature profile to attain acceptable accuracy in the temperature differences. In a classic study at Lake Hefner (Anderson, 1954) an accuracy of 5% in the evaporation estimate was achieved for periods of a week or more but decreased for shorter periods. For a shallow (average depth 0.6 m) lake, Stewart and Rouse (1976) assumed that daily values were sufficiently accurate to use them as a standard against which an alternative method was validated.

Disadvantages of the energy balance method are the large number of measurements needed, the frequency of the measurements, and the difficulties inherent in making some of them. Consequently it is expensive and has not often been used in the more complete form of equation (7): exceptions include the Lake Hefner study (Anderson, 1954), the work by Stauffer (1991) on Lake Mendota, Wisconsin, and Sturrock *et al.* (1992) on Williams Lake, north central Minnesota, and a comparative study of evaporation from two lakes in Florida (Sacks *et al.,* 1994).

To enhance accuracy, measurements of surface and profile water temperatures and the micrometeorology should be made at representative points over the water body. This has often been achieved using an anchored instrumented raft (e.g. Anderson, 1954; Assouline and Mahrer, 1993; Sturrock *et al.,* 1992). For ease of maintenance and cost, however, measurements have been made over land and sometimes data used from distant weather stations. Work has been done to determine the effect on the accuracy of the evaporation estimates of using land-based and distant data sources (e.g. Keijman, 1974; Rosenberry *et al.,* 1993; Winter and Rosenberry, 1995).

The importance of the net radiation in the energy budget makes its accurate measurement or estimation paramount. Modern instrumentation allows the direct measurement of the net radiation to an accuracy of about 5%. Where net radiometers are unavailable, R_n is calculated either from measurements or from estimates of the radiation components and over the years there has been much work on improving the accuracy of this approach. A review of the many equations that have been developed to allow the short and long-wave radiation to be estimated from astronomical, meteorological and climatological data is given by Henderson-Sellers (1986), and Brutsaert (1982). Major factors affecting the value of the incoming solar radiation, S, are atmospheric scattering, absorption and reflection, so that cloud amount and type are important, as well as season and latitude. The reflected component depends upon the albedo, which in turn varies depending upon the degree of cloudiness with solar elevations angle. The long-wave radiation, $L\downarrow$, emitted by the atmosphere can be calculated from vertical profiles of temperature and humidity: such data are, however, not often available and it is usual to calculate it using the Stefan-Boltzmann relationship

$$L_{\downarrow} = \varepsilon\sigma T_a^4 \tag{9}$$

where T_a is the air temperature near the surface, σ is the Stefan-Boltzmann constant and ε is the clear-sky atmospheric emissivity which can be calculated from air temperature and humidity near the surface. Like S, $L\downarrow$ is also affected by cloudiness. The Stefan-Boltzmann equation, with appropriate surface values for the temperature and emissivity, is also used to calculate long-wave radiative loss, $L\uparrow$, from the water. Stannard and Rosenberry (1991) found that measuring the incoming radiation and modelling the outgoing radiation resulted in overestimates of lake net radiation compared with directly measured values. One possible reason for this was differences in incoming radiation between the lake and the site where they measured it, 4.5 km away. Whether estimated or measured, the radiation values are integrated to produce period estimates consistent with the other measurements.

Estimation of the Bowen ratio, β, requires measurement of air temperature and specific humidity above the water and temperature and saturated specific humidity at the temperature of the water surface. This is usually achieved using wet and dry bulb thermometers at a reference height on a raft and a thermistor within the top few centimetres of the water.

The change in heat storage N per unit surface area is calculated from the following:

$$N = \rho c d \frac{\Delta T_w}{\Delta t} \tag{10}$$

where ρ, c, d and ΔT_w are the density, specific heat, depth, and change in spatially averaged temperature of the water body in time step Δt. For pans and shallow lakes that are well mixed, T_w can be approximated by the surface temperature (Keijman, 1974). This however begs the question as to a suitable average value for the surface temperature; in calm conditions and high solar radiation, spatial variation in surface temperature can be large over short time scales. For deep lakes it is necessary to conduct thermal surveys consisting of temperature profiles with depth, measured ideally at a sufficient number of stations to produce a good average. For example, in the detailed Lake Hefner study, surveys were made at weekly intervals at 16 stations and daily at one of two stations (Anderson, 1954), while at Williams Lake, surveys were made fortnightly at 16 stations (Sturrock *et al.*,1992). Selection of the appropriate time interval, which will depend upon the size of the water body, can result in the value of N being small enough to be neglected.

Estimation of the energy advected in and out of the lake requires that the inflow and outflow are gauged accurately and the water temperature measured. Inflow includes rivers and land surface run off, bank storage and seepage from groundwater. Outflow includes rivers, controlled withdrawals (reservoirs) and leakage to groundwater. Where inflow or outflow are large relative to the volume of the water body, and cannot be accurately gauged, the energy balance method may become unusable. However, in many lakes the relative inflow and outflow are small (e.g. Williams Lake, Sturrock *et al.,* 1992). Sturrock *et al.* (1992) calculated the groundwater volumes using Darcy's Law and used water temperatures to determine inflow and leakage. The energy advected by rainfall is usually determined from the recorded rainfall and the wet-bulb temperature recorded during rainfall. Sacks *et al.* (1994), Stauffer (1991) and Sturrock *et al.* (1992) concluded that for the lakes that they studied, the advected energy was trivial compared to the other terms, being around 1% of the radiation terms. However, of the advected terms, Sacks *et al.* (1994) and Stauffer (1991) found that the largest was that due to precipitation. In non-natural or semi-natural water bodies, other advective components may be large but easy to measure, e.g. reservoirs and cooling ponds.

In some circumstances, the heat conduction term G can be significant, Sturrock *et al.* (1992) found that in the summer neglecting it made a 7% difference to the estimated evaporation from Williams Lake (average depth 5.2 m) in Minnesota. Stauffer (1991) states that ignoring sediment heat exchange can be a major source of error in estimation of evaporation and that the Lake Hefner results may be in error through ignoring this component. He used annual sine-wave functions to model the sediment-water heat exchange (Likens and Johnson, 1969).

Comparisons have been made of the evaporation estimated using the energy balance with direct measurements using eddy correlation equipment mounted over lakes (Sene *et al.,* 1991; Stannard and Rosenberry, 1991; Assouline and Mahrer, 1993). These show that, for deep lakes, the hourly or daily evaporation rates are determined primarily by the wind speed and atmospheric stability, with the energy being supplied from the heat storage in the lake. In consequence, estimates of evaporation on a short timescale determined from the energy balance method for deep lakes may not be accurate. Assouline and Mahrer (1993) found that, for a period of high wind speeds and sensible heat advection, the daily average evaporation rate estimated from the energy budget method was 2.8 mm day^{-1} compared to 4.1 mm day^{-1} measured using eddy correlation. However, they also found much closer agreement for a second period when wind speeds were lower and advection was less. Good agreement between the energy budget and eddy correlation estimates of evaporation can be obtained for longer time scales. Anderson (1954) gives an accuracy of evaporation estimates of 5% for periods of a week or more for Lake Hefner.

6 BULK OR MASS TRANSFER

A simple derivation of the bulk transfer equation is given by Sene *et al.,* (1991). It has the form

$$E = Cu(e_s^* - e) \tag{11}$$

where C is the mass transfer coefficient, u is the wind speed and e_s^* and e are the saturated vapour pressure of the air at the water surface temperature and the vapour pressure of the air at the reference height. The mass transfer coefficient can be thought of as the total drag coefficient; the combination of skin friction and a force resulting from the deceleration of the wind in the direction of flow. It can be shown that the mass transfer coefficient and the roughness lengths used in the Penman-Monteith equation are linked. Over a uniform surface C can be calculated from theory which indicates that it is a function of the atmospheric stability and the roughness of the surface which itself is affected by the wind speed (Brutsaert, 1982). The value of the coefficient has often been determined for sea surfaces although there is considerable scatter in the results (Brutsaert, 1982). For most inland water bodies the conditions of surface uniformity are not met and it is necessary to make more restrictive assumptions to obtain a theoretical solution to the evaporation and heat transfer equations. The value of C reflects the transfer characteristics of the particular water body which are determined by its geometry, plant cover, and the topography, land use and climate of the surrounding land. Moreover the value of the coefficient is specific for the characteristics of the site used to record the meteorological data such that a value derived for wind speed measured at 2 m will not be correct for use with wind speeds measured at 10 m, even at the same site. Over the years meteorological data have been inconsistently measured using a variety of different standards, resulting, according to Singh and Xu (1997), in over 100 such evaporation formulae. It is therefore not possible to find a value of C that is applicable to all water bodies. Rather, it is best to determine it empirically for a particular water body from the ratio of the mean evaporation rate (measured using a standard method, for example eddy correlation or the energy budget) to the mean vapour pressure gradient. Nevertheless attempts have been made to produce a generally applicable value. On the basis of an extensive measurement programme on reservoirs in the western USA, Harbeck (1962) suggested an expression for C that incorporated the area of the water body. In appropriate units (Shuttleworth, 1993) the transfer equation is

$$E = 2.909 A_s^{-0.05} u_2 (e_s^* - e) \tag{12}$$

where A_s is the area of the water surface in m^2, and u_2 is the wind speed at 2 m above the water surface. This is suitable for lakes in the range of 50 m < $A_s^{0.5}$ < 100 km that are in a relatively arid environment. A similar expression for pans in the range 0.5 m < $A_s^{0.5}$ < 5 m, is also given by Shuttleworth based on the work of Brutsaert and Yu (1968), namely

$$E = 3.623 A_s^{-0.066} u_2 (e_s^* - e) \tag{13}$$

The weak inverse dependence of the transfer coefficient on the size of the water body reflects the effect of the reduced efficiency of turbulent transfer over the smooth water surface (Shuttleworth, 1993). However some observations indicate that transfer is enhanced over large water bodies. Venalainen *et al.* (1998), for example, observed, from direct micrometeorological and eddy correlation measurements over two lakes in Sweden, that evaporation rates were greater from the larger of the two lakes. They attributed this to the effect of increased wind speed more than compensating for opposing effect of increased humidity of the air associated with the larger distance travelled by the air over water. They also noted that evaporation from lakes with forest at the edge would be reduced through sheltering: apparently the reduction in turbulence associated with the reduced wind speeds more than compensates for the increased aerodynamic roughness of the forest.

An alternative form for the mass transfer equation dating from the 19th century has also been widely used. This takes the form

$$E = f(u)(e_s^* - e) \tag{14}$$

where $f(u)$ is a function of the wind speed, $f(u) = a + bu$ with empirical constants a and b, allowing for free convection, i.e. evaporation when there is no wind. Sweers (1976) reviewed wind speed functions and concluded that, for a temperate climate, best results were obtained using the wind function of McMillan (1973) adjusted for the area of the water body in relation to the lake studied by McMillan. This function is

$$f(u) = \left(\frac{5 \times 10^6}{A_s} \right)^{0.05} (3.6 + 2.5 u_3) \tag{15}$$

where u_3 is the wind speed measured over the water at 3 m above the surface.

Once a value for C has been determined, this method requires routine measurements of wind speed and vapour pressure at the same height as the measurements used in the determination of C. Unless the water body is less than a few metres across these measurements should be made over the water so that they are representative of conditions prevailing over most of the water surface. In addition, to determine e_s^* the average surface temperature of the water must also be measured.

When evaporation estimates are required on hourly or daily time scales then the effects of atmospheric stability must also be considered (e.g. Stauffer, 1991) but for long-term estimates these effects can usually be neglected.

Other functional forms of balance equations, some of which include air temperature, have been used. Singh and Xu (1997) tested 13 mass transfer equations, transformed into seven generalised forms using climatological data from northwest Canada. They compared monthly evaporation estimates with pan data at four sites after calibration of each equation for each site. Agreement was generally good between the estimates and measurements for a particular site but the equations did not give good results at sites for which they were not calibrated. On a monthly time scale the humidity gradient, rather than the wind speed, primarily controlled the evaporation. This is at variance with the observations on two Swedish lakes (Venalainen *et al.,* 1998) and demonstrates a limitation of pan estimates of evaporation.

Simon and Mero (1985) decided against the mass transfer method to estimate evaporation from Lake Kinneret in Israel because of inconsistent results and large scatter in estimates of the transfer coefficient. In contrast Sacks *et al.* (1994) found good agreement (generally within 8%) between the energy-budget evaporation and monthly mass-transfer evaporation for a shallow lake in Florida, but larger discrepancies (mean monthly difference of 24%) for a similar but deeper lake, also in Florida. Correcting the mass transfer coefficient for stability effects (Stauffer, 1991; Harbeck *et al.*,1958) did not improve estimates. Sacks *et al.* (1994) suggested that the differences might be a smoothing effect caused by using long-term mean vapour pressure gradients, one of the main difficulties with this method being that it is sensitive to the errors in the vapour pressure gradient. They also found that using the Harbeck (1962) form for the mass transfer coefficient produced lower values that resulted in underestimates of the evaporation from the shallow lake by 14% and from the deep lake by 27%. This was in contrast to Sturrock *et al.* (1992) who found that the Harbeck prescription gave higher values than those based upon energy budget estimates.

7 COMBINATION EQUATIONS

In the last fifty years possibly the most widely used formula to estimate evaporation from water, or vegetation, has been the Penman equation (Penman, 1948). Its success when applied in many different locations is attributable to its physical basis. Linacre (1993) presents a table comparing monthly or annual measured evaporation with Penman estimates for a wide range of water bodies from around the world. The median value of the ratio of measured to estimated evaporation is 0.99 with a standard deviation of 0.12.

Penman combined the mass transfer and energy budget approaches and eliminated the requirement for surface temperature to obtain his expression for the evaporation in mm per day from open water:

$$E = \frac{\Delta R_n^{'}}{\Delta + \gamma} + \frac{\gamma f(u)(e_a^* - e)}{\Delta + \gamma} \tag{16}$$

where $R_n^{'}$ is the net radiation in units of equivalent depth of water (mm day^{-1}), Δ is the slope of the saturated vapour pressure-temperature curve and γ is the psychometric coefficient (or c_p/λ). Penman subsequently modified this to a form commonly known as Penman E_T, the evaporation rate expected from short well-watered vegetation. The open water form (equation (16)) does not allow for heat storage and was not intended for use in estimating evaporation from deep water bodies, with or without components of advected energy. To incorporate advected energy, R_n is replaced in equation (16) with A, the available energy, which is the sum of net radiation and any energy advected into the water body less any that goes into storage.

When air travels a long distance over a wet surface it will tend to saturation so that the second term in equation (16) tends to zero. The first term represents the lower limit of evaporation and is referred to as the equilibrium rate. In practice, however, equilibrium evaporation is rarely found because the atmosphere near the surface is never truly homogeneous and, even over oceans, atmospheric humidity deficits develop. Priestley and Taylor (1972) analysed data collected over oceans and extensive saturated land surfaces and found that the evaporation values were fitted using

$$E = \alpha \frac{\Delta A}{\Delta + \gamma} \tag{17}$$

where A is the available energy and the constant α accounts for the evaporation arising from the humidity deficit in addition to the equilibrium term. The equation is now known as the Priestley-Taylor equation. Priestley and Taylor found that the average value of α was 1.26 from the data they examined and there has been subsequent corroboration of this value by other studies. de Bruin and Keijman (1979) used the Priestley-Taylor equation to estimate the evaporation from a large shallow lake (Lake Flevo, 460 km^2, mean depth 3 m) in the Netherlands and found very good agreement with daily evaporation measured by the energy budget and water budget methods during the summer and early autumn with $\alpha = 1.25$. However, they also found diurnal variation in the value of α which they attributed to the variation of the difference between air and water temperatures and suggested that the conditions producing such variation would be expected from many lakes. They also found evidence of seasonality in the value of α, of at least the same magnitude as the diurnal variation in evaporation. This variation is the result of some evaporation occurring when the available energy was zero. de Bruin and Keijman also found very good agreement between the evaporation estimated from the energy budget and that estimated using the following formula

$$E = \frac{\Delta A}{\lambda(0.85\Delta + 0.63\gamma)} \tag{18}$$

derived from the Priestley-Taylor equation, the relationship between α and β and an empirical relationship, $\beta = 0.63\,\gamma/\Delta - 0.15$, given by Hicks and Hess (1977).

Stewart and Rouse (1976) derived a variation of equation (17) by using a linear function of incoming solar radiation to replace the net radiation and heat storage. The parameters, a and b, of the function were obtained by regression and the values are necessarily specific to their lake. The resulting equation is identical to the formula of Makkink (1957) who used it to estimate the evaporation from well-watered grass and is

$$E = a\frac{\Delta}{\Delta + \gamma}S + b \tag{19}$$

A disadvantage of the Priestley-Taylor equation is the requirement for measured $R_n^{'}$ and N values, especially the latter (the change in the heat stored in the water): it is often not possible or is too expensive to make adequate measurements of N for a large water body. de Bruin (1978) overcame this difficulty by combining the Penman and Priestley-Taylor equations, thus eliminating the energy term to give the relationship

$$E = \left(\frac{\alpha}{\alpha - 1}\right)\left(\frac{\gamma}{\Delta + \gamma}\right) f(u)(e_a^* - e) \tag{20}$$

This formula requires only measurements of air temperature, humidity deficit and wind speed at 2 m. de Bruin tested the method by using a form of the wind function given by Sweers (1976) with time-averaged input data measured at the centre of Lake Flevo to calculate evaporation for varying time intervals. He found good agreement with estimates from the energy budget method for intervals of 10 days or more. He also found that the Priestley-Taylor coefficient was not constant for intervals of a day or less.

A more general form of combination equation is given by the Penman-Monteith equation (Monteith, 1965). It was developed to describe the evaporation of water vapour from the sub-stomatal cavities of plants into the atmosphere. Essentially the evaporation rate is obtained from the simultaneous solution of diffusion equations for heat and water vapour, and the energy balance equation. When applied to open water it takes the form

$$E = \frac{1}{\lambda}\left[\frac{\Delta A + \rho c_p (e_a^* - e) / r_a}{\Delta + \gamma}\right] \tag{21}$$

where the aerodynamic resistance r_a is the resistance that the water molecules encounter in moving from the water surface to a reference height in the atmosphere and is inversely proportional to the wind speed. This equation has the same physical basis as the Penman equation but does not contain the empirical calibration factors inherent in the wind function used by Penman. It thus is often considered to represent the best description of the evaporation process and in this sense is often preferred to other estimates provided the necessary input data are available, the same proviso as required by the Penman model. The heat storage and net energy advected into the water body need to be included in the available energy, A. Accurate estimates also require that the value of the aerodynamic resistance, r_a, accounts for the effects of surface roughness, size of the water body, and atmospheric stability.

The combination equations proper require values of net radiation, air temperature, vapour pressure and wind speed. Fewer input data are required by the simpler, derived equations like the Priestley-Taylor equation. Unlike the energy balance and mass balance methods, they do not require values of surface temperature but for accurate estimates of evaporation it is necessary to estimate or measure the heat storage in the water, unless the time interval over which evaporation estimates are required is such that the heat storage can be neglected. Linacre (1993) derived a simplified version of the Penman equation requiring just air temperature, wind speed and dew point data and he suggested two different methods for estimating solar irradiance, one of which used rainfall as a surrogate indicator of cloudiness and the other which accounted for temperature variation with latitude, altitude and distance from the sea. Using this method with monthly or longer average input data, he obtained good agreement (within 5%) with measured evaporation rates for a range of different sites in Australia, USA and Copenhagen.

As with the other methods, uncertainties in the evaporation estimates are larger for bodies of deep water because of the larger heat storage component. For large water bodies this component is determined primarily by the surface energy exchange which in turn is affected by the atmospheric stability and must be allowed for when daily, or shorter, estimates are needed. When water bodies exceed a certain depth stratification occurs and the heat storage has to be determined from measured temperature profiles or hydrodynamic models. For lakes in tropical climates the water temperature can be nearly constant all year round so that the change in heat storage can be neglected in cases (Sene *et al.,* 1991).

On the basis of data collected from the literature Linacre (1993) states that the probable error associated with monthly or annual evaporation estimated using the Penman equation with monthly data is about 8%.

Stewart and Rouse (1976) tested the Priestley-Taylor equation (equation (17)) using data from a small (105 m^2) shallow (mean depth 0.6 m) lake and found very good agreement with evaporation estimated by the energy budget method on a half-hourly and daily basis. They concluded that evaporation could be estimated within 5% using this method. They also tested the Makkink formula (equation (19)) and found that it gave estimates of evaporation to within 10% over fortnightly to monthly intervals.

In addition to the uncertainty connected with the heat storage and the measurement errors of the driving data there can also be systematic uncertainty associated with the aerodynamic resistance in the Penman-Monteith equation. Near the edge of a body of water the aerodynamic resistance will be determined chiefly by the aerodynamic roughness of the surroundings in the direction of the prevailing wind. For example if there is forest in that direction then it will generate large turbulent eddies but it will also reduce the wind speed. The effect of the surroundings of the water body on the aerodynamic resistance will reduce with distance. Usually, because the water is smoother than most other surfaces, the wind speed will increase with distance over water resulting in a smaller value of resistance, unless the higher wind speed causes waves with associated increased roughness. Shuttleworth (1993) suggested that using the aerodynamic resistance implicit in the Penman equation (equation (16)) in the Penman-Monteith equation (equation (21)) might result in overestimates of evaporation from very large lakes of 10-15%: this is, however, probably an upper limit because Shuttleworth does not appear to have taken into account the increase in wind speed that occurs.

The model of de Bruin removes the requirement to know the heat storage term but its effects will be reflected in variation in the value of the Priestley-Taylor ‘constant’ α. If the appropriate value is not known then errors may be quite large because of the sensitivity of the evaporation estimate to this parameter. The model is also sensitive to errors in the vapour pressure gradient.

8 EQUILIBRIUM TEMPERATURE METHOD

Useful open water evaporation methods have been derived from a detailed consideration of the heat transfer processes occurring at the surface of a water body. These require the same driving data as the combination equations except for the water heat storage which is calculated within the models. Edinger *et al.* (1968) introduced the concept of an equilibrium temperature and associated time constant, determined from meteorological data, towards which the water temperature is driven by the net heat exchange, that is, when the water is at equilibrium temperature the net rate of heat exchange is zero. From this they were able to derive an expression for the temperature of a well-mixed body of water as a function of time and water depth. Once the water temperature is estimated then it can be used to estimate the evaporative and sensible heat fluxes, the heat storage and the long wave radiative loss from the water. A similar approach was taken by Keijman (1974) who then used the calculated heat storage in the Penman equation (see equation (16)) to estimate the evaporation from the shallow Lake Flevo. de Bruin (1982) used a slightly different approach to obtain an expression for the water temperature that also used an equilibrium temperature, but one that was constant and equal to the mean value of that used by Keijman. Using this model with ten-day means of standard land-based meteorological data, de Bruin achieved good agreement between measured and predicted water temperatures over several years for two reservoirs of different depths in the Netherlands. This type of work was extended by Fraedrich *et al.* (1977) by considering the effect of energy advected to a reservoir by inflow and outflow: two characteristic temperatures and related time constants allow simulation of the energetics of the reservoir in response to surface-transfer and hydrological forcing mechanisms. The model was applied to a large shallow reservoir and good agreement was achieved between simulated and predicted monthly averages of water temperatures and upward long-wave radiation. The water temperature was used to calculate the energy storage term, which in turn was used in the Penman (16) and Priestley-Taylor (18) equations to estimate evaporation rates: the best estimate of evaporation was derived from the Penman equation.

With regard to aspects of data requirements, Keijman (1974) used daily mean values of dry and wet bulb air temperature, and wind speed together with sunshine duration, measured around the perimeter of a lake, from which he estimated net radiation, to drive his model. He also compared the effect of using the data collected at the centre of Lake Flevo with data collected at two stations on the perimeter of the lake. Equally good results were achieved when using data from a perimeter station if it was downwind of the lake. Fraedrich *et al.* (1977) used monthly mean weather data together with rates and temperatures of the inflow and outflow to drive their more sophisticated model.

There is little reported in the literature on the detail of errors associated with these equilibrium temperature methods. de Bruin (1978) found good agreement between estimated and measured lake surface temperatures. Good agreement of surface temperatures was also found by Keijman (1974) and reflected in estimates of daily lake evaporation estimated by the Penman equation that had a standard error of 0.6 mm. Fraedrich *et al.* (1977) also found that the Penman equation gave better estimates of evaporation using this method than the Priestley-Taylor equation.

9 EMPIRICAL FACTORS

In operational estimates of evaporation, empirical factors to convert evaporation rates measured or estimated for one type of land surface (the reference evaporation) to those of another have a useful practical record of application, particularly in areas where data are sparse. They are comparable to the use of pan coefficients to convert measurements of evaporation from evaporation pans to those of other water bodies or land surfaces and generally consist of multiplying the reference evaporation by an empirical factor.

Although the source of the reference evaporation could be any method, in practice, as well as pans, it has frequently been combination equations because these equations use relatively readily-available meteorological data and have proven to be robust at estimating evaporation.

Penman (1948), for example, gives the following factors to convert evaporation rates from 'turf with a plentiful water supply' to an open water surface in southern England exposed to the same weather conditions:

midwinter (November - February)	1.67
spring and autumn (March, April, September, October)	1.43
midsummer (May - August)	1.25

These values were derived from measurements of evaporation at a single site, Rothamsted Experimental Station in southern Britain, using cylinders 0.76 m in diameter and 1.83 m deep and so the use of these factors outside these conditions should be treated with caution. Measurements of evaporation from water were used to calibrate Penman's model of evaporation and so these factors should be used with estimates of reference evaporation calculated using this model.

Finch (2003) used the measurements of Lapworth (1965), made over a seven year period from a 17 hectare lake southeast of London, to derive monthly empirical factors to be applied to the grass potential evaporation calculated using the Penman equation and a form of the Penman-Monteith equation. These factors should similarly be used with caution in different climatic regions.

Doorenbos and Pruitt (1984) list empirical factors (or crop coefficients) to allow evaporation to be estimated for a wide range of land surfaces from time series of evaporation calculated using a modified version, involving changing the wind function, of the Penman (1948) model for grass. The factors given for open water evaporation are:

humid environment – light to moderate wind	1.1
humid environment – strong wind	1.15
dry environment – light to moderate wind	1.15
dry environment –strong wind	1.2

These coefficients can be used for calculating annual totals of evaporation for all water bodies and monthly totals for shallow water bodies (less than 5 m). They are considered appropriate for the estimation of monthly totals of evaporation from deep water bodies in equatorial regions but Doorenbos and Pruitt warn that, when applied to deep water bodies (greater than 25 m) with a change in climate during the year, in spring and the early summer the correct coefficients may be 20-30% lower due to heat storage in the water body: conversely, due to heat release, the correct coefficients may be 20-30% higher in later summer and autumn.

Allen *et al.* (1998) have given crop coefficients for use with Penman-Monteith estimates of evaporation, for a hypothetical crop with a bulk surface resistance of 70 s m^{-1} and a height of 0.12 m (which can be taken as corresponding to short grass freely supplied with water). The coefficient given for water bodies in subhumid climates or tropics and water bodies less than 2 m in depth is 1.05. Two coefficients are given for water bodies greater than 5 m depth, clear of turbidity, in temperate climates. A value of 1.25 is recommended for the autumn and winter when the water body is releasing thermal energy and 0.65 when the water body is gaining thermal energy (spring and summer). Allen *et al.* urge caution in using these coefficients.

Morton (1983 a, b) forwarded a pragmatic approach to lake evaporation recognising that fully descriptive process methods would not for some time become operationally routine in many areas. His approach is based on the conceptual and empirical relationship between areal and potential evaporation, with an extension to estimate lake evaporation from monthly temperature, humidity and sunshine (or radiation) observations over land, with approximate adjustments for lake depth and salinity.

The potential errors in using empirical factors arise from measurement errors inherent in the meteorological data used to calculate the reference evaporation (or direct pan estimates if these are the source of basic data to transpose) and the appropriateness of the transposition. The dominant driving variable is the net radiation which is generally derived from measurements of the sunshine hours or incoming solar radiation which in the case of modern instruments are generally accurate to around ±5%. It is essential that the meteorological data used to calculate the reference evaporation are representative of the meteorological conditions over the water body. It is difficult to quantify the error that may arise from failing to do this but it could amount to around 10%. In general meteorological data should be used from a nearby site that reflects the general topography and land cover in the vicinity of the water body.

The main source of error in empirical transposition is likely to be the use of an inappropriate coefficient for the water body in question. To achieve a high level of accuracy, the coefficient(s) should be determined for each particular water body and should vary throughout the year. In practice, it is frequently not feasible to do this and thus the coefficients should only be relied upon when used in the conditions that they were determined. In particular, differences in the depth of water, and possibly the surface area, may result in errors of up to 30% in evaporation totals for time periods less than a year. The use of a single set of empirical factors for every year will potentially result in errors due to variations in the weather from year to year. For monthly estimates, this can average between 15 and 20%. In addition, the empirical factors used should have been developed for the estimates of potential evaporation of a specific source: for example the differences between estimates of potential evaporation by the Penman (1948) model and the MORECS (Hough and Jones, 1997) implementation of the Penman-Monteith model are likely to result in differences in estimates of open water evaporation using the same set of empirical factors of around 30%.

10 EXAMPLE VALUES OF LAKE EVAPORATION BY WMO REGION

See table on next four pages.

WMO Regions:

I	Africa
II	Asia
III	South America
IV	North America, Central America and the Caribbean
V	South-West Pacific
VI	Europe

Location	Lake area	WMO Region	Determination date	Method of determination	Evaporation values	Notes	Source
Aswan High Dam Lake, Egypt, Sudan	~ 5250 km^2	I	1979 - 1983	pan and isotope experiments	7.7 - 21.6 mm day^{-1}	monthly variation given	Aly *et al.*, 1993
Lake Ziway, Ethiopia	mean ~ 490 km^{-2}	I	1969 - 2000	energy balance, Penman and Morton CRLE	1730 - 1880 mm $year^{-1}$	mean monthly values also given	Vallet-Coulomb *et al.*, 2001
Lake Volta, Ghana	~4953-8063 km^2	I	1972 - 1974	equilibrium temperature method	105 -172 mm $month^{-1}$	comparisons with long-term Penman averages	Hough, 2003
Lake Biwa, Japan	680 km^2	II	1985 - 1987	range of direct and indirect methods; emphasis on bulk transfer	'winter' 0.14 - 2.94 mm day^{-1}; 'summer' 0.49 - 6.13 mm day^{-1}	whole lake evaporation related to site measurements	Ikebuchi *et al.*, 1988
Lake Qinghai, China	4304 km^2 (1986)	II	1958 - 1988 1958 - 1984	pan thermodynamic model	1459 mm $year^{-1}$ 753-938 mm $year^{-1}$	shrinking lake; sensitivity to climate change evaluated	Qin and Huang, 1998
Caspian Sea, (Russia, Iran, Kazakhstan, Turkmenistan, Azerbaijan)	379000 km^{-2}	II	1900 - 1990	water balance	mean 377 km^{-3} $year^{-1}$	potential of isotopic tracer approaches evaluated	Froehlich, 2000
Lake Ahung Co, Tibet, China	3.6 km^2	II	1995 - 2001	lake energy balance model	mean monthly values between -30 and +160 mm $month^{-1}$	annual anomalies given	Morrill, 2004

Lake Serra Azul, Minas Gerais State, Brazil	8.8 km^2	III	1993 - mid 1995	energy budget and Morton CRLE	1.7 – 5.6 mm day^{-1}	reservoir	dos Reis and Dias, 1998
Lake Poopó, Bolivia	up to 3000 km^2	III	1990 -1995	adjusted pan observations	110 - 170 mm $month^{-1}$	lake dries at times	Zola and Bengtsson, 2006
Lake Titicaca, Peru and Bolvia	8560 km^2	III	various between 1956 and 1987	bulk transfer, energy budget and water budget	1350 -1900 mm $year^{-1}$	relationship to rainfall; comparison with pan data	Declaux *et al.*, 2007
			1964 - 1978	pan-lake transfer coefficient	130 -160 mm $month^{-1}$		
				radiative and atmospheric forcing variable models	50 -210 mm $month^{-1}$	8 models compared	
United States: 30 lakes and reservoirs	0.2 - 19400 km^2	IV	various between 1906 and 1974	water budget, energy budget, pan, mass transfer, Morton CRAE	505 - 2930 mm $year^{-1}$	comparisons of measured and modelled determinations; modelled monthly values; contoured map of US annual lake evaporation	Andersen and Jobson, 1982
Lake Victoria (East Africa); United States and	various	IV (and I)	variously 1960 - 65, 1964 -1969	complementary relationship lake evaporation model (CRLE)	600 - 2000 mm $year^{-1}$	contoured maps of annual evaporation for Canadian (east of	Morton, 1983 a,b

Canada lakes and reservoirs: Salton Sea, Silver, Hefner, Pyramid, Winnemucca, Ontario, Last Mountain and Dauphin						Pacific divide) and southern United States lakes, and for southern US reservoirs	
‘K-6’ lake, Lupin, NWT, Canada	0.06 km^2	IV	1992, 1993 ice-free periods only	isotope mass balance	mean 1.9 - 3.4 mm day^{-1}		Gibson *et al.*, 1996
Lake Frome, South Australia	maximum ~2700 km^2	V	‘several years’ prior to 1985	depth deuterium profiles	90 - 230 mm $year^{-1}$	a drying salt lake	Allison and Barnes, 1985
Lake Toba, Sumatra, Indonesia	1100 km^2	V	Jan - Feb 1989	eddy correlation measurements	mean 0.22 mm $hour^{-1}$; max 0.64, min -0.01 mean 5.1 mm day^{-1}; max 7.6, min 3.0		Sene *at al.*, 1991
Five reservoirs, Victoria, Australia	‘small’	V	1973 - 1976	Morton CRLE	monthly values between 13.2 and 144.5 mm $month^{-1}$	‘net reservoir evaporation’ ie. open water evaporation less original evapotranspiration from the reservoir site	Gan *et al.*, 1991

Australia: Lake Eucumbene; Cataract, Mundarin and Manton reservoirs	4.4 – 145 km^2	V	two years, prior to 1977	net heat models and measurements	986 - 2149 mm $year^{-1}$; mean monthly values between 1.5 and 6.5 mm day^{-1}	deep and shallow; alpine to semi-arid tropical	Vardavas and Fountoulakis, 1996
Kempton Park reservoir, London, UK	0.17 km^2	VI	1956 - 1962	pan, tank, Penman and Walker method	12.5 – 140 mm $month^{-1}$		Lapworth, 1965
Lake Kinneret (Sea of Galilee), Israel	166 km^2	VI	May-October 1990	eddy correlation and energy budget	-0.3 - 1.1 mm $hour^{-1}$ 2 - 12 mm day^{-1}	large differences noted between measured and estimated rates; comment on hot dry Sharav conditions	Assouline and Mahrer, 1993
Lake Tämnaren, south Sweden	35 km^2	VI	June-September 1994	water and isotope mass balances	0.6 - 6.5 mm day^{-1}	shallow lake	Saxena, 1996
characteristic case		~30° latitude		simplified Penman	rate of change of -4 mm $year^{-1}$	generalised evaporation trend study	Linacre, 2004

11 SUMMARY OVERVIEW

Pan evaporation and empirical factors can be considered as similar methods as they rely on the use of factors (ideally time varying) to convert 'standard' estimates of evaporation to those of the water body. The difference between the methods is the source of the reference evaporation: measurements from an evaporation pan or estimates of evaporation calculated using meteorological data.

The development of physically based models, such as the Penman-Monteith combination equation, has resulted in reliable estimates of evaporation being readily derived from meteorological data where sufficient such data are available.

The difficulty and expense of measuring all the elements that are required for the mass balance means that this method has only been applied in a few, exceptional circumstances. These tended to be in the 1950s and 60s. Since then, developments in instrumentation have meant that the energy budget method has become a more practical proposition. However, both these methods rely on calculating a balance, so that the errors accumulate in the estimate of evaporation. The result is that, unless the evaporation losses are comparable in magnitude to the other changes in the budget, the errors are likely to be large. Nevertheless, the energy balance method is considered to give the most accurate estimates of evaporation. For both methods, the estimates of evaporation are specific to the site where the measurements are made and cannot be transferred to other water bodies. The advantage is that local factors, such as thermal stratification, are taken into account.

The bulk transfer method seems initially very attractive as it makes use of data that are easily measured, namely the meteorological variables and the water body's surface temperature. In practice, the sensitivity to vapour pressure measurements combined with the difficulty of defining the wind function reduce the accuracy of this method. Because of the need for measurements of the surface temperature of the water body, the estimates of evaporation are specific to the site. However, this ensures that local conditions, such as thermal stratification, are handled.

Combination equations are arguably the most widely used method of estimating evaporation. They are particularly attractive because they make use of readily available meteorological data. The major limitation is that they do not take the heat storage of the water body into account if driven by net radiation data. This can be remedied by carrying out periodic thermal surveys of the water body and inputting the available energy, rather than the net radiation, into the equation: this does, however, make the methods site-specific.

The equilibrium temperature method is a relatively new method, which might explain why there are few references to it in the literature. It is an attractive method because it is physically based, uses readily available meteorological data and takes the heat storage of the water body into account. The only major limitation is that it assumes that the water body is uniformly mixed and thus it does not consider thermal stratification.

In practice, the availability and quality of data have a major impact on the method chosen to quantify lake evaporation. It is important to bear in mind the associated level of accuracy achievable from the selected method.

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REFERENCES

Allen, R. G., Pereira, L. S., Raes, R. and Smith, M. 1998 Crop evapotranspiration - Guidelines for computing crop water requirements. Irrigation and Drainage Paper 56, 300 pp, Rome, UN Food and Agriculture Organization.

Allison, G. B. and Barnes, C. J. 1985 Estimation of evaporation from the normally 'dry' Lake Frome in South Australia. Journal of Hydrology, 78, 229-242.

Aly, A. I. M, Froelich, K., Nada, A., Awad, M., Hamza, M. and Salem, W. M. 1993 Study of environmental isotope distribution in the Aswan High Dam Lake (Egypt) for estimation of evaporation of lake water and its recharge to adjacent groundwater. Environmental Geochemistry and Health, 15, 37-49.

Anderson, E. R. 1954 Energy-budget studies. In: Water Loss Investigations: Lake Hefner Studies. U.S. Geological Survey Professional Paper 269, 71-119.

Andersen, M. E. and Jobson, H. E. 1982 Comparison of techniques for estimating annual lake evaporation using climatological data. Water Resources Research, 18, 630-636.

Assouline, S. and Mahrer, Y. 1993 Evaporation from Lake Kinneret: 1 Eddy correlation system measurements and energy budget estimates. Water Resources Research 29, 901-910.

Brutsaert, W. 1982 Evaporation into the Atmosphere - Theory, History and Applications. Dordrecht, Holland, D Reidel Publishing Company.

Brutsaert, W. and Yu, S. L. 1968 Mass transfer aspects of pan evaporation. Journal of Applied Meteorology, 7, 563-566.

Davies, J. A. 1972 Surface albedo and emissivity for Lake Ontario. Climatological Bulletin, 12**,** 12-22.

de Bruin, H. A. R. 1978 A simple model for shallow lake evaporation. Journal of Applied Meteorology, 17, 1132-1134.

de Bruin, H. A. R. 1982 Temperature and energy balance of a water reservoir determined from standard weather data of a land station. J. Hydrol. 59, 261-274.

de Bruin, H. A. R. and Keijman, J. Q. 1979 The Priestley-Taylor evaporation model applied to a large shallow lake in the Netherlands. Journal of Applied Meteorology, 18, 898-903.

Delclaux, F., Coudrain, A. and Condom, T. 2007 Evaporation estimation on lake Titicaca: a synthesis review and modelling. Hydrological Processes, 21, 1664-1677.

Doorenbos, J. and Pruitt, W. O. 1984 Guidelines for predicting crop water requirements, Rome, UN Food and Agriculture Organization Irrigation and Drainage Paper 24.

dos Reis, R. J. and Dias, N. L. 1998 Multi-season lake evaporation: energy-budget estimates and CRLE model assessment with limited meteorological observations. Journal of Hydrology, 208, 135-147.

Edinger, J. E., Duttweiler, D. W. and Geyer, J. C. 1968 The response of water temperature to meteorological conditions. Water Resources Research, 4, 1137-1143.

Finch, J. W. 2003 Empirical factors for estimating open-water evaporation from potential evaporation. Journal of the Chartered Institution of Water and Environmental Management, 17, 51-53,

Fraedrich, K., Erath, B. G. and Weber, G. 1977 A simple model for estimating the evaporation from a shallow water reservoir. Tellus, 29, 428-434.

Froehlich, K. 2000 Evaluating the water balance of inland seas using isotopic tracers: the Caspian Sea experience. Hydrological processes, 14, 1371-1383.

Gan, K. C., McMahon, T. A. and O'Neill, I. C. 1996 Sensitivity of reservoir sizing to evaporation estimates. Journal of Irrigation and Drainage Engineering, 117, 324-335.

Gangopaghaya, M., Harbeck, G. E., Nordenson, T. J., Omar, M.H. and Uryvaev, V. A. 1966 Measurement and estimation of evaporation and evapotranspiration. Technical Note 83, 121pp. World Meteorological Organization.

Gibson, J. J., Edwards, T. W. D. and Prowse, T. D. 1996 Development and validation of an isotopic method for estimating lake evaporation. Hydrological processes, 10, 1369-1382.

Harbeck, G. E. 1962 A practical field technique for measuring reservoir evaporation utilizing mass-transfer theory. USGS Professional Paper 272-E, pp.101-105. US Geological Survey.

Harbeck, G. E., Kohler, M. A. and Koberg, G. E. 1954 Water-loss investigations: Lake Hefner Studies. USGS Professional Paper 269, pp.1-158. US Geological Survey.

Harbeck, G. E., Kohler, M. A. and Koberg, G. E. 1958 Water-loss investigations: Lake Mead studies. US Geological Survey Professional Paper 298, 99pp, US Geological Survey.

Henderson-Sellers, B. 1986 Calculating the surface-energy balance for lake and reservoir modeling - a review. Reviews of Geophysics, 24, 625-649.

Hicks, B. B. and Hess, G. D. 1977 On the Bowen ratio and surface temperature at sea. Journal of Physical Oceanography, 7**,** 141-145.

Hough, M. N. 2003 Evaporation from Lake Volta: the equilibrium temperature method. Project report to the Volta River Authority, Met Office, Bracknell, UK.

Hough, M. N. and Jones, R. J. A. 1997 The United Kingdom Meteorological Office rainfall and evaporation calculation system: MORECS version 2.0 - an overview. Hydrology and Earth System Science 1, 227-239.

Hounam, C. E. 1973 Comparison between pan and lake evaporation. Technical Note 126, World Meteorological Organization.

Hoy, R. D. and Stephens, S. K. 1977 Field study of evaporation, Research Project 6815. Canberra, A.C.T., Australian Water Resources Council, Department of Natural Resources.

Ikebuchi, S., Seki, M. and Ohtoh, A. 1988 Evaporation from Lake Biwa. Journal of Hydrology, 427-449.

Keijman, J.Q. 1974 The estimation of the energy balance of a lake from simple weather data. Boundary-Layer Meteorology, 7, 399-407.

Lapworth, C. F. 1965 Evaporation from a reservoir near London. Journal of the Institution of Water Engineers, 19, 163-181.

Likens, G.E. and Johnson, N. M. 1969 Measurement and analysis of the annual heat budget for the sediments in two Wisconsin lakes. Limnology and Oceanography, 14, 115-135.

Linacre, E .T. 1993 Data-sparse estimation of lake evaporation, using a simplified Penman equation. Agricultural and Forest Meteorology, 64, 237-256.

Linacre, E. T. 2004 Evaporation trends. Theoretical and Applied Climatology, 79, 11-21.

Makkink, G. F. 1957 Ekzameno de la formulo de Penman. Netherlands Journal of Agricultural Science 5, 290-305.

McMillan, W. 1973 Cooling from open water surfaces. Final Report: Part 1; Lake Trawsfynydd cooling investigation. NW/SSD/RR/1204/73, pp.1-55.

Mironov, D., G. Kirillin, E. Heise, S. Golosov, A. Terzhevik, and I. Zverev, 2003 Parameterization of lakes in numerical models for environmental applications. Proc. of the 7th Workshop on Physical Processes in Natural Waters, A. Yu. Terzhevik, Ed., Northern Water Problems Institute, Russian Academy of Sciences, Petrozavodsk, Karelia, Russia, 135-143

Monteith, J. 1965 Evaporation and the environment. In: 19th Symposium of the Society of Experimental Biology. Cambridge, UK Cambridge University Press, 205-234.

Morrill, C. 2004 The influence of Asian summer monsoon variability on the water balance of a Tibetan lake. Journal of Paleolimnology, 32, 273-286.

Morton, F. I. 1983a Operational estimates of areal evapotranspiration and their significane to the science and practice of hydrology. Journal of Hydrology, 66, 1-76.

Morton, F. I. 1983b Operational estimates of lake evaporation. Journal of Hydrology, 66, 77-100.

Ogawa, K., Schmugge, T., Jacob, F. and French. A. 2002 Estimation of broadband land surface emissivity from multi- spectral thermal infrared remote sensing. Agronomie, 22, 695-696.

Penman, H. L. 1948 Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society, A193, 120-145.

Priestley, C. H. B. and Taylor, R. J. 1972 On the assessment of surface heat flux and evaporation using large-scale parameters. Monthly Weather Review, 100, 81-92.

Qin, B. and Huang, Q. 1998 Evaluation of the climatic change impacts on the inland lake - a case stdy of Lake Qinghai, China. Climatic Change, 39, 695-714.

Richter, D. 1988 Methods for the determination of the albedo and of the longwave outgoing radiation from a free-water surface and of the longwave incoming radiation of the atmosphere. Zeitschrift fur Meteorologie, 38, 219-233.

Rosenberry, D. O., Sturrock, A. M. and Winter, T. C. 1993 Evaluation of the energy budget method of determining evaporation at Williams Lake, Minnesota, using alternative instrumentation and study approaches. Water Resources Research, 29, 2473-2483.

Sacks, L. A., Lee, T. M. and Radell, M. J. 1994 Comparison of energy-budget evaporation losses from two morphometrically different Florida seepage lakes. Journal of Hydrology, 156, 311-334.

Saxena, R. K. 1996 Estimation of lake evaporation from a shallow lake in central Sweden by oxygen-18. Hydrological Processes, 10, 1273-1281.

Sene, K. J., Gash, J. H. . and McNeil, D. D. 1991 Evaporation from a tropical lake: comparison of theory with direct measurements. Journal of Hydrology, 127, 193-217.

Shuttleworth, W. J. 1993 Evaporation. In: Maidment, D.R., (ed.) Handbook of Hydrology, New York: McGraw-Hill, 4.1-4.53.

Simon, E. and Mero, F. 1985 A simplified procedure for the evaluation of the Lake Kinneret evaporation. Journal of Hydrology, 78, 291-304.

Singh, V. P. and Xu, C. 1997 Evaluation and generalization of 13 mass-transfer equations for determining free water evaporation. Hydrological Processes, 11, 311-323.

Stannard, D. I. and Rosenberry, D. O. 1991 A comparison of short-term measurements of lake evaporation using eddy correlation and energy budget methods. Journal of Hydrology, 122, 15-22.

Stauffer, R. E. 1991 Testing lake energy budget models under varying atmospheric stability conditions. Journal of Hydrology, 128, 115-135.

Stewart, R. B. and Rouse, W. R. 1976 A simple method for determining the evaporation from shallow lakes and ponds. Water Resources Research, 12, 623-628.

Sturrock, A. M., Winter, T. C. and Rosenberry, D. O. 1992 Energy budget evaporation from Williams Lake - a closed lake in North Central Minnesota. Water Resources Research, 28, 1605-1617.

Sweers, H. E. 1976 A nomogram to estimate the heat-exchange coefficient at the air-water interface as a function of wind speed and temperature; a critical survey of some literature. Journal of Hydrology, 30, 375-401.

Vallet-Coulomb, C., Legesse, D., Gasse, F., Travi, Y. and Chernet, T. 2001 Lake evaporation estimates in tropical Africa (Lake Ziway, Ethiopia). Journal of Hydrology, 245, 1-18.

Vardavas, I. M. and Fountoulakis, A. 1996 Estimation of lake evaporation from standard meteorlogical measurements: application to four Australian lakes in different climatic regions. Ecological Modelling, 84, 139-150.

Venalainen, A., Heikinheimo, M. and Tourula, T. 1998 Latent heat flux from small sheltered lakes. Boundary-Layer Meteorology 86, 355-377.

Winter, T. C. 1981 Uncertainties in estimating the water balance of lakes. Water Resources Bulletin, 17 , 82-115.

Winter, T. C. and Rosenberry, D. O. 1995 Evaluation of 11 equations for determining evaporation for a small lake in the north central United States. Water Resources Research, 31, 983-993.

Zola, R. P. and Bengtsson, L. 2006 Long-term and extreme water level variations of the shallow Lake Poopo, Bolivia. Hydrological Sciences Journal, 51, 98-114.