ESTIMATING EVAPORATION FROM WATER SURFACES[1]

By
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INTRODUCTION

Because of its nature, evaporation from water surfaces is rarely measured directly, except over relatively small spatial and temporal scales (Jones 1992). Evaporation from water is most commonly computed indirectly by one or more techniques. These include pan coefficients × measured pan evaporation, water balance, energy balance, mass transfer, and combination techniques. The selection of the "best" technique to use for a particular computation is largely a function of the data availability, type or size of the water body, and the required accuracy of the estimated evaporation. The most commonly used method in the US for estimating evaporation from small, shallow water bodies, is to measure evaporation from a standard pan and then multiply by a coefficient.

The purpose of this brief paper is to describe some principles involved in estimating evaporation with example data from large and small water bodies. Particular emphasis will be on practical procedures and techniques that professionals can use to estimate evaporation from shallow water bodies using pan evaporation and a combination equation using available weather data and/or new data collected specifically for estimating daily evaporation.

BACKGROUND

Theories of evaporation from water surfaces go back to at least the 8th century B.C., but measurement and experimentation go back to the 17th and 18th centuries. Brutsaert (1982) credits Dalton's 1802 paper (cited by Brutsaert 1982, p. 31) as a major event in the development of evaporation theory. He expressed Dalton's results in present day notation as $E = f(\bar{u})(e_o - e_a)$ where E is the rate of evaporation expressed as rate per unit time, $\bar{u}$ is mean wind speed, e_o is the saturation vapor pressure at the temperature of the water surface and e_a is the vapor pressure of the air.

Evaporation requires energy, the primary source being solar radiation. Bowen (1926) developed what is now known as the Bowen ratio (BR). It is the ratio of heat loss by conduction to that by evaporation, or $\Delta t/\Delta e$, the ratio of air temperature gradient to vapor pressure gradient above the surface. The BR has had a major impact on methods of estimating and measuring evaporation technologies. McEwen (1930) described evaporation as a transformation of energy based on earlier work of Bowen (1926) and Cummings and Richardson (1927).

Many US states have tabulated evaporation data measured with evaporation pans. Some states have computed state-wide estimates of evaporation from shallow ponds and reservoirs (Borrelli et al. 1998). Many early studies and reports have also addressed pan evaporation

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relative to large water bodies. More recently, most studies have incorporated energy balance concepts.

ENERGY CONSIDERATIONS

Incoming solar radiation is the principle source of energy for evaporation. Although calculation of net radiation for land and water surfaces is similar, there is one major difference. For land surfaces, the net solar radiation (incoming solar radiation minus reflected solar radiation) is converted to sensible and latent heat at the soil or plant surfaces. In contrast, net solar radiation is not all absorbed at the water surface. Part of net solar radiation may penetrate to great depths in clear water. The depth of penetration varies with wave length.

In pure water, about 70% of the solar radiation adsorption occurs in the top 5 m. Solar radiation adsorbed below the water surface is stored as energy and is not immediately available for evaporation or for sensible heat. Stored energy typically results in a lag of evaporation relative to net solar radiation, and there usually is less total annual evaporation than when there is no significant energy storage. As the solar intensity begins to decrease after about mid July, the stored energy gradually becomes available for evaporation and sensible heat loss. For shallow water bodies, the amount of energy stored during the rising and falling solar cycle may not be large, but usually is large enough to affect daily evaporation rates.

The fraction of daily solar radiation that is reflected from water surfaces (albedo) during summer months is about 6-7%. The amount of solar energy absorbed by water bodies, therefore, is larger than that absorbed on land surfaces. The albedo for green crops is about 23-25%. The reflectance from water varies with the solar angle, with water turbidity, wave height and the ratio of diffuse to total solar radiation as summarized in table 1 (Cogley 1979).

Table 1. Monthly mean Grishchenko albedo (percent) of open water for latitudes 0º to 70º N. (From Cogley 1979, table 5)

Lat	Month												
(deg)	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
70	30.1	33.8	22.9	14.8	11.6	11.2	11.4	13.4	20.2	31.3	30.1	- -	14.1
60	33.9	24.0	15.5	10.5	8.8	8.4	8.6	9.8	13.6	21.6	32.1	35.5	12.0
50	22.0	16.1	10.8	8.4	7.5	7.3	7.4	8.0	9.9	14.4	21.0	24.1	10.2
40	14.5	11.1	8.5	7.3	6.8	6.7	6.8	7.1	8.0	10.3	13.8	16.1	8.6
30	10.3	8.6	7.3	6.7	6.5	6.4	6.4	6.6	7.1	8.2	10.0	11.1	7.6
20	8.3	7.4	6.7	6.4	6.3	6.3	6.3	6.4	6.6	7.2	8.1	8.7	6.9
10	7.2	6.7	6.4	6.3	6.4	6.4	6.4	6.3	6.3	6.6	7.1	7.4	6.6
0	6.6	6.4	6.3	6.4	6.6	6.8	6.7	6.4	6.3	6.4	6.6	6.8	6.5

Examples of Energy Storage and Release

In a large reservoir study in California, water temperature vs. depth was measured several times during the season. Typical results during the increasing solar radiation cycle are shown in fig. 1. Results during the decreasing solar radiation cycle are shown in fig. 2. The shape of the water temperature profile during the decreasing cycle illustrates an important characteristic of water-temperature profiles which will be discussed later. Similar water temperature profiles have been observed in the reservoirs on the Colorado River below Hoover dam. At the end of the season, as cold water settles, the entire water profile becomes mixed as shown in fig. 3.

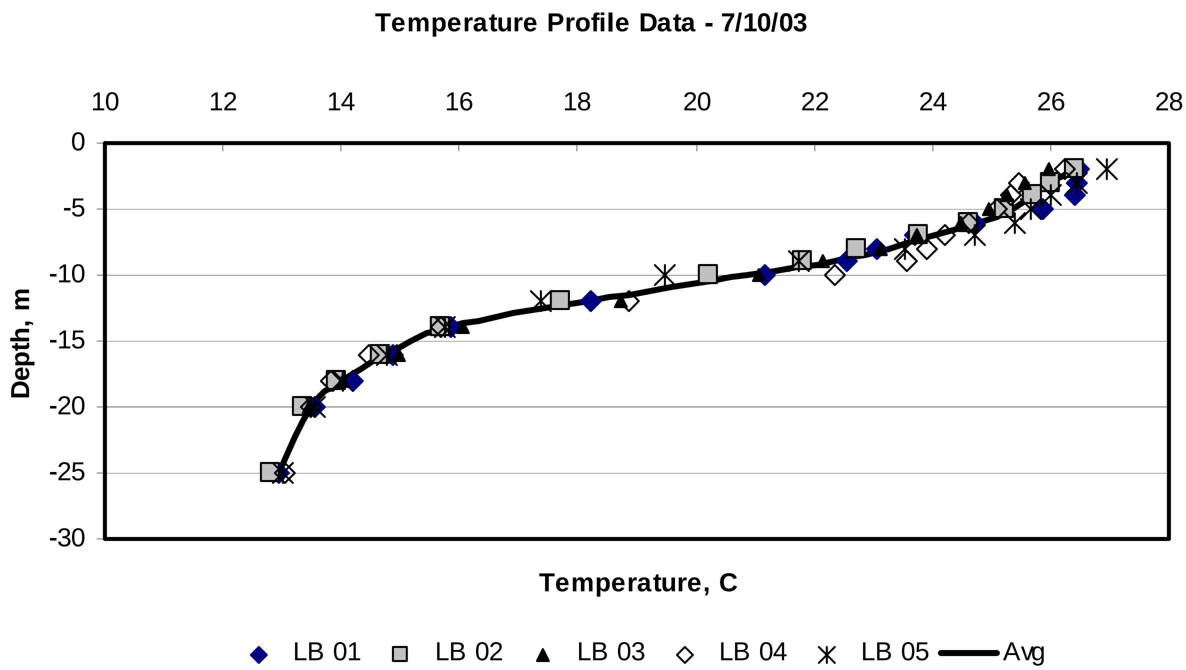

Fig. 1. Thermal profile data for Lake Berryessa in California on July 10, 2003. The individual points refer to locations within the reservoir. (Jensen et al. 2005).

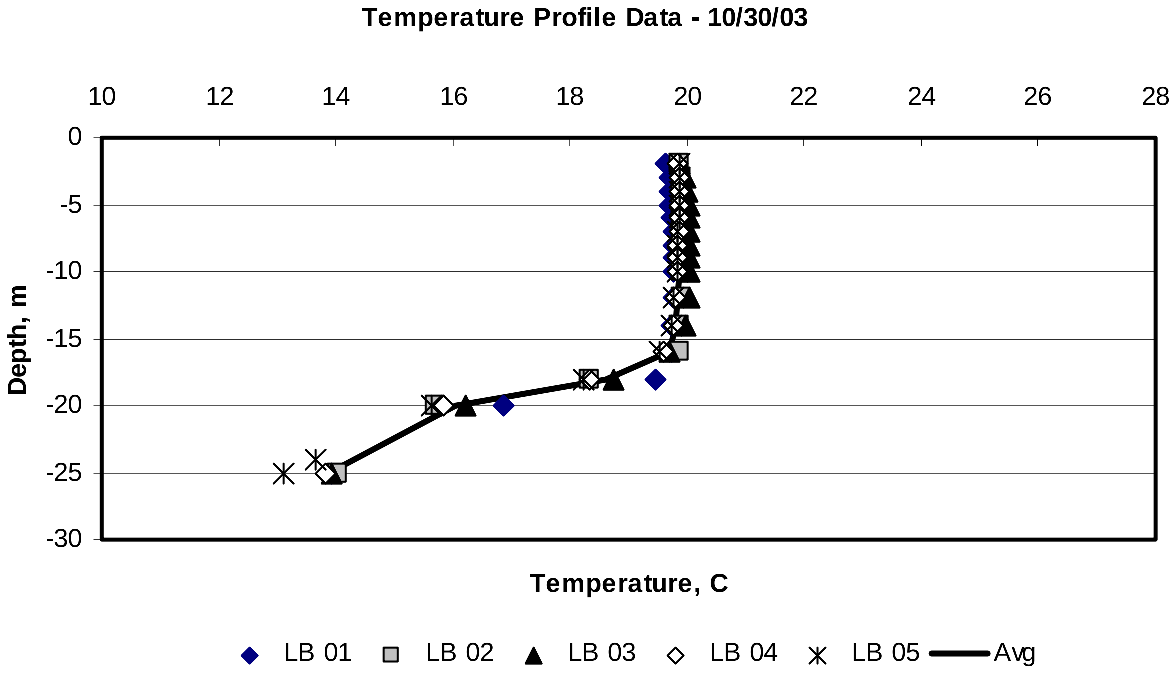

Fig. 2. Thermal profile data for Lake Berryessa, California on October 30, 2003. The individual points refer to locations within the reservoir. (Jensen et al. 2005).

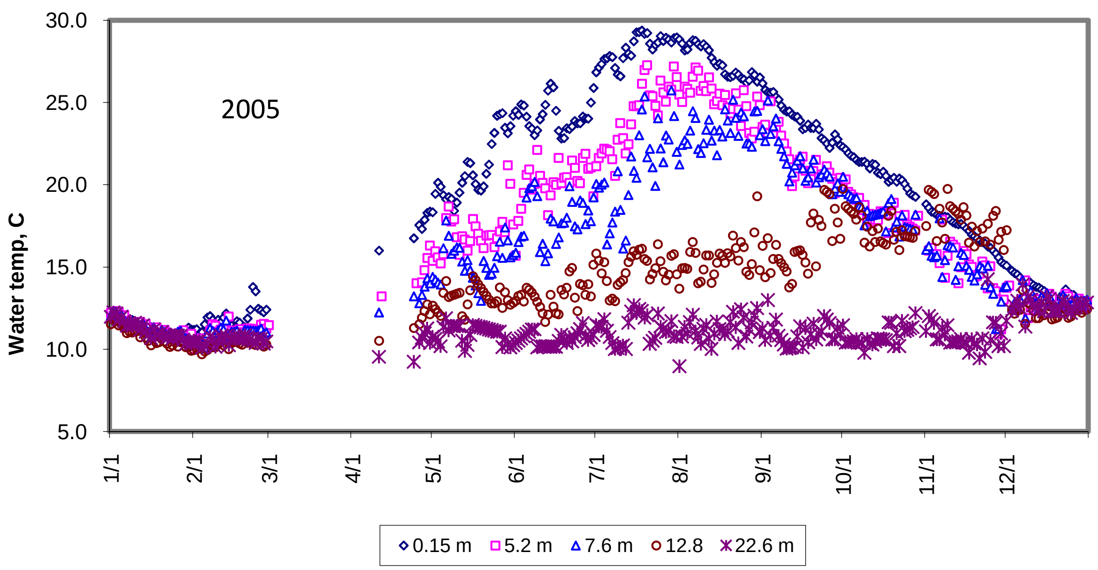

Fig. 3. Measured water temperatures (thermograph), Lake Berryessa, 2005.

Effect of Energy Storage on Estimated Evaporation

The storage and release of energy affects evaporation rates and thus the pan coefficients that are used for estimating evaporation from large water bodies. Maximum average energy storage rates of about 6 MJ m^{-2} d^{-1} (equivalent to about 2.5 ml d^{-1}, or 0.1 in. d^{-1}) was estimated for Lake

Berryessa using temperature data to a depth of 25 m. Energy storage rates of about 10 MJ m^{-2} d^{-1} were observed in Lake Mead[3] calculated using daily time steps and thermal profiles to a depth of 45 m. Surface water temperatures reached an average maximum of 27°C in July for Lake Berryessa and a maximum of about 28°C in August for Lake Mead. Energy storage results in a lag of surface water temperature and evaporation rates relative to net radiation. The lag is greater as energy storage increases in lakes or reservoirs of large depth, cleaner water, and greater depth of solar radiation penetration.

Advected Energy

Another energy factor that must be considered for reservoirs on river systems is the advection of energy in the water due to inflows and outflows. Using water temperature and flow data for the Colorado River I calculated the effect of advected energy on estimated average monthly evaporation between dams on the Colorado River as shown in fig. 4. The data between Hoover Dam and Davis Dam was not consistent because of the wide fluctuations in flow below Hoover Dam. The largest advected energy effect was on the reach between Davis Dam and Parker Dam. Lake Havasu lies between Davis Dam and Parker Dam. Advected energy is probably not a significant factor in shallow water bodies with limited inflows and outflows.

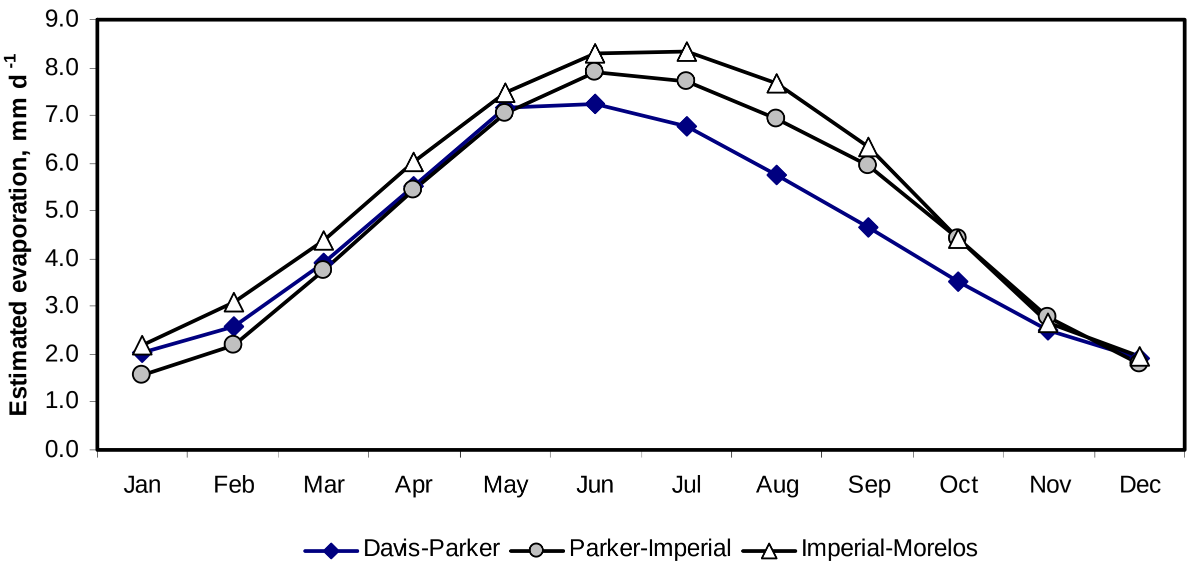

Fig. 4. Effect of advected energy on estimated mean monthly evaporation.[4]

ESTIMATING EVAPORATION FROM SHALLOW WATER BODIES

Equilibrium Temperature Approach

[3] Craig Westenburg (2006), personal communication.
[4] M.E. Jensen, personal communication.

Estimating evaporation from shallow water bodies using energy balance or combination methods also requires estimating energy storage and release. Edlinger et al. (1968) introduced the concept of equilibrium temperature for estimating energy storage, Q_t, (equivalent to the term G used for land studies) using weather data collected from one or more nearby stations. The model was developed further by Keijman (1974), Fraederich et al. (1977), de Bruin (1982) and evaluated by Finch (2001). The model calculates an equilibrium time constant, τ (days), and the equilibrium temperature, T_e (when the net rate of energy exchange is zero). The model has one key assumption and that is it assumes that the water is well-mixed and that heat flux at the bottom can be neglected.

Example Applications of the Equilibrium Approach

Keijman (1974) estimated evaporation for a lake of 46,000 ha and a depth of 3 m in the Netherlands. He used weather data from a station along the perimeter of the lake. He estimated that daily evaporation could be estimated with an error of 0.6 mm d^{-1} if the weather station was - downwind from the lake. He also used a weather data site in the center of the lake. The estimated average water temperature was 0.3°C below the measured value. Stewart and Rouse (1976) estimated evaporation from a shallow lake in the Hudson Bay lowlands of northern Ontario. The lake was very small, about 400 m in diameter, with an average depth of only 0.6 m. They reported that evaporation can be approximated by the equilibrium equation that is essentially the same as the Priestley and Taylor (1972) equation, LE =1.26[$\Delta/(\Delta + \gamma)$](R_n – G). They concluded that the Priestley-Taylor model can be used to estimate daily evaporation within 5% for this shallow lake. They also presented a model for shallow lakes and ponds of depths 0.5 to 2 m using temperature and incoming solar radiation. With that model, they indicated that evaporation can be estimated within 10% for periods of two weeks to a month.

Finch (2001) used the equilibrium model and estimated evaporation for two reservoirs near London over a 7-year period using a daily time step. Most measurements were made on the east reservoir with an area of 17 ha and a maximum depth of 7.2 m. Water temperatures that were measured weekly at depth intervals of 1 m near the center of the lake indicated uniform temperature during the winter, but water temperatures decreased with depth during the summer indicating energy storage. The modeled monthly evaporation followed the annual cycle very well with a tendency to over estimate evaporation during the summer and underestimate evaporation during the winter. The average estimated annual evaporation was 619 mm which was 6% lower than measured evaporation. The annual estimate taking estimated energy storage into account was about 16% lower than when assuming $Q_t = 0$ in the Penman equation. Finch indicated that the model can be used with considerable confidence.

Finite Difference Model

Finch and Gash (2002) developed a finite difference model that they indicated is an improvement over the equilibrium model when applied to the lake used by Finch (2001). The change in energy storage, Q_t, is calculated from the energy balance equation as

$$Q_t = R_n - \lambda E - H \tag{1}$$

The terms λE and H are calculated from flux-gradient equations for a water surface (Brutsaert, 1982)

$$\lambda E = f(u)(e_{sw} - e) \tag{2}$$

and

$$H = \gamma f(u)(T_w - T_a) \tag{3}$$

where Q_t = energy stored, R_n = net radiation, λ = the latent heat of vaporization, H = sensible heat flux, e_{sw} = the saturated vapor pressure at the water temperature, e = vapor pressure at the reference height and γ = the psychrometric constant. The change in surface temperature appeared to be based on estimated evaporation for the previous day and on the previous day's temperature.

I am not familiar with any estimates of evaporation made with the equilibrium model in the US or other studies using the finite difference model of Finch and Gash (2002).

Energy and Water Balance Studies

Most of the US lake studies found in the literature have involved energy and water balances. Examples include Lake Hefner in Oklahoma, Harbeck et al. 1958, U.S. Geological Survey 1954, 1958; Pretty Lake in Indiana, Ficke 1972; Harney-Malheur Lake in Oregon, Hosteler and Bartlein 1990; Lake Mendota in Wisconsin, Stauffer 1991; and Williams Lake in Minnesota, Sturrock et al. 1992.

Pan Evaporation and Pan Coefficients

One of the earliest studies of pan and lake evaporation was that by L.G. Carpenter at Fort Collins, Colorado (Carpenter 1898). He observed that water temperatures were higher in the shallow areas than in the deep areas, and that pan evaporation at night is almost the same as during the day. He reported 11 years of evaporation data measured in Colorado's 3 × 3 ft (0.91 × 0.91 m) sunken pan where the rim of the pan was within 5 cm of the ground surface. The Colorado pan was originally two feet deep and since 1889, it has been three feet deep. A significant observation was that during the day, the surface water warmed rapidly and the lower layers slowly. This thermal profile was caused by absorption of solar radiation with depth as illustrated in fig. 1. At night as the surface water layer cools by evaporation and outgoing long wave radiation. As the upper layer of water cools the colder, denser surface water settles until reaching the same temperature at a lower depth. This characteristic affects the shape of the thermal profile as solar radiation decreases after about mid-July. In shallow water bodies, 1 m or less, this characteristic tends to make water temperatures uniform in the morning and is important when estimating energy storage.

Example Pan Evaporation Studies

Rohwer 1931. In 1931, Rohwer (1931) published the results of a classic series of indoor and outdoor studies of evaporation from water surfaces conducted from 1923 to 1925, from 1926 to 1928, and from ice in 1928 and 1929. Several outdoor studies were conducted in 1927-1928 that included evaluating elevation effects on evaporation. He reviewed several Dalton-type equations expressing evaporation as a function of wind speed and the vapor pressure deficit (VPD). The most notable outdoor study involved an 85-ft diameter (25.9 m) copper-lined reservoir 6.75 ft (2.06 m) deep that was located near where the Lory Student Center now stands. He also evaluated several types of evaporation pans. Daily measurements were reported for Weather Bureau (WB) Class A land tank or pan and the reservoir for the periods of Oct.-Nov., 1926, Apr.-Nov., 1927, and Apr.-Nov. 1928. It included surface water and air temperatures, VPD, wind speed, and evaporation. The other pans or tanks included the Geologic Survey floating tank and the Colorado-type land tank. The floating tank was set in the middle of the reservoir. Rohwer contended that evaporation from a lake or reservoir is substantially the same as that from a 12 ft (3.66 m) square ground tank 3 ft (0.91 m) deep. However, a check of evaporation data from the 12-ft ground tank and a Class A pan study (Sleight, 1917) indicated that evaporation from the 12-ft tank is higher than from the large reservoir (mean ratio of Class A pan to 12-ft pan of 0.66 compared to a mean ratio of 0.70 for the Class A pan to the circular reservoir). Rohwer also measured air temperature 1 inch (25 mm) above the water surface and the surface water temperature using calibrated thermometers. Multiple thermometers were used for the air and surface water temperatures with the thermometer bulbs held just beneath the surface. The thermometers were shielded with a white metal plate. Multiple daily readings were made. The average monthly surface water temperatures were always higher than the air temperature one inch above the surface (fig. 7). Average reservoir-pan ratios for the 85-ft (m) reservoir and adjacent Class A pan in Fort Collins, CO are shown in fig. 6. Note the ratios for the 85-ft reservoir have a pattern similar to the coefficients for Lake Elsinore in California (Young 1947). This indicates that energy storage in a shallow (2 m) water body influences the resulting monthly pan coefficient.

Rohwer derived several evaporation formulas. Indoor controlled wind speed and evaporation measurements were made using a 3 × 3 ft square tank 10 inches deep (0.91 × 0.91 × 0.25 m) from which he derived his formula No. 6 of $E = (0.44 + 0.118\ W)\ (e_s - e_d)$ where W is "water-surface" wind speed in mi h^{-1} and vapor pressures of the air and the water surface are in inches of mercury. Penman (1948) extrapolated Rohwer's evaporation equation made at 5,000 ft (1,524 m) elevation to sea level and to wind speed measured at a height of 2 m to obtain his eq. No. 3. He then showed in his fig. 4 that Rohwer's equation was similar to what he had derived in England, although Penman perceived that a power wind function [$E = (aW^b)\ (e_s – e_d)$] was more theoretical than the linear wind function.

During a symposium addressing evaporation from water surfaces in 1933, Rohwer (1934) presented a detailed discussion of evaporation from different pans including a section by R. Follansbee who summarized national and international records of reservoir evaporation as related to pan evaporation. The papers were followed by nine discussions. A. Santos, Jr. from Brazil in his discussion stated that it was regrettable that not even passing mention was made by the authors of work done by I.S. Bowen, N.W. Cummings, G.F. McEwen, B. Richardson and C. Montgomery who have succeeded in placing the physics of evaporation on a firm basis. This observation illustrated the common tendency for engineers during that period, in particular, to only read and cite engineering sources. This tendency may have been a factor in the rate of developing improved methods of estimating evaporation in the US. Today, engineers typically cite sources from various disciplines.

Young 1947. Another classic study involving pan evaporation vs. lake or reservoirs was presented by Young (1947). He summarized pan records and calculated pan coefficients in California. Pan coefficients for the WB Class A pan and a screened pan for Lake Elsinore were reported in his table 10. Lake Elsinore is a small lake with an area of about 2,220 ha and an estimated maximum depth of about 6 m at that time. The annual lake evaporation to WB pan ratio was 0.77 for Lake Elsinore. Monthly Class A pan coefficients for evaporation from Lake Elsinore, show the effects of energy storage and release in this relatively shallow lake (fig. 5). Also shown are pan coefficients for Lake Okeechobee, FL, a shallow lake in a warm humid climate with an average depth of about 3 m (Kohler 1954).

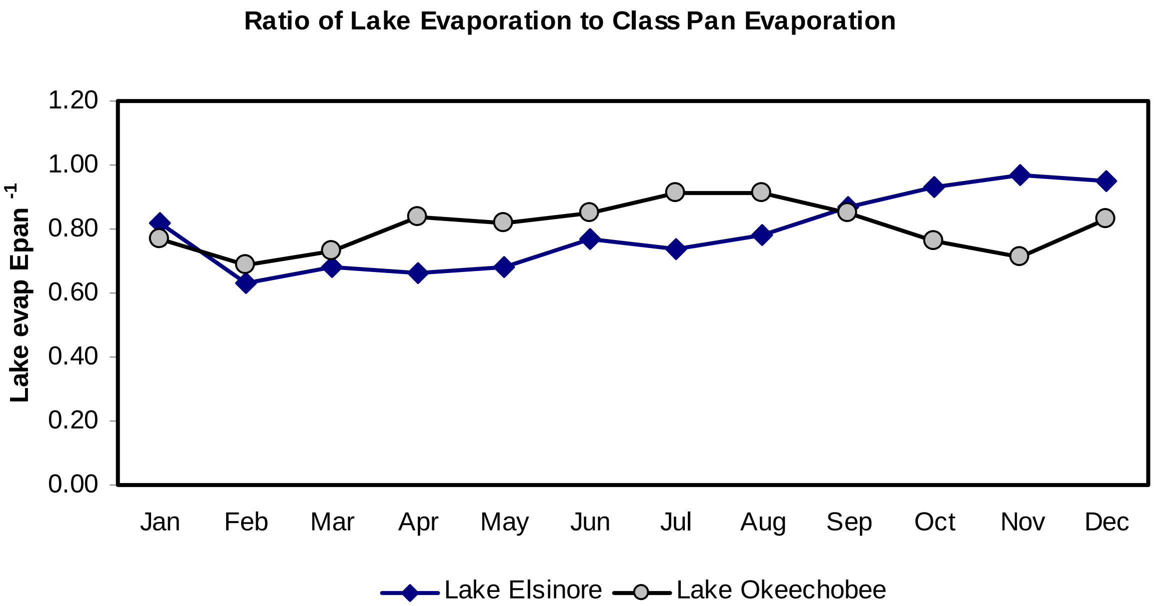

Fig. 5. Ratios of lake evaporation to Class A pan evaporation for Lake Elsinore, CA by Young (1947) and for Lake Okeechobee, FL by Kohler (1954).

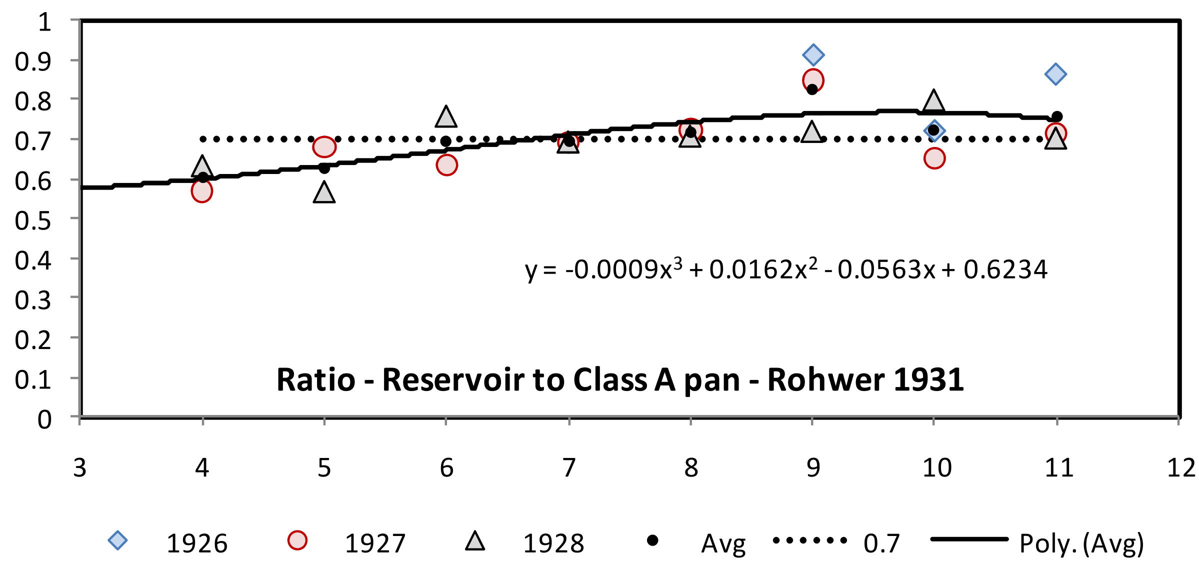

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Fig. 6. Ratios of evaporation from a 85-ft (27.9 m) diameter reservoir in Fort Collins to Class A pan evaporation located adjacent to the reservoir (Rohwer 1931).

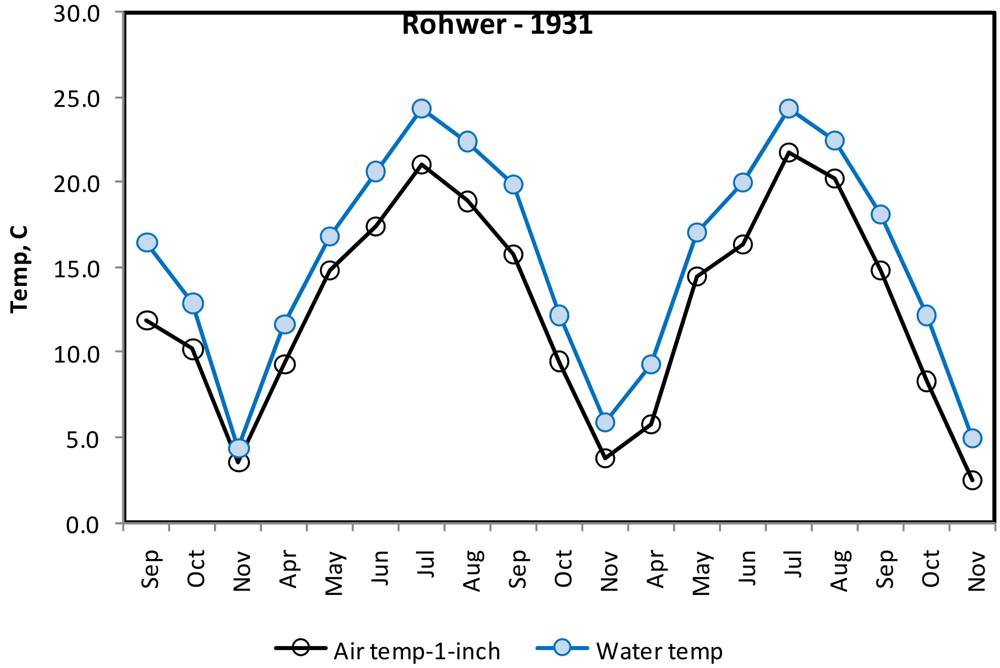

Fig. 7. Mean monthly air and surface water temperatures for the 85-ft (27.9 m) diameter reservoir in Fort Collins (Rohwer 1931).

Kohler 1954. Kohler (1954) used the results of the detailed U.S. Geologic Survey (USGS) Lake Hefner evaporation study combined with pan evaporation for estimating lake evaporation. One of his objectives was to derive a more reliable procedure for estimating lake evaporation from pan evaporation and related meteorological data normally collected by the Weather Bureau. The annual Class A pan coefficient derived for Lake Hefner was 0.69. Monthly coefficients varied because of the temperature lag in the lake due to differences in energy storage capacities of the two water bodies. Pan coefficients tended to be lower in spring months. Kohler concluded that annual lake evaporation could be estimated within 10-15% by applying the annual coefficient 0.70 to Class A pan evaporation.

Kohler 1959. Kohler et al. (1959) presented a series of excellent maps of average Class A pan evaporation, average annual lake evaporation in inches, average annual Class A pan coefficient in percent, and average annual May-October evaporation in percent of annual values. These maps still serve as a useful guide to expected annual evaporation in the USA although they have been replaced with free water surface evaporation estimates (Farnsworth et al. 1982).

Traditional Pan Coefficient Method. Traditional methods for estimating evaporation from lakes and reservoirs in the US have been based on evaporation measured from a network of evaporation pans. The standard evaporation pan in the US is the NWS Class A pan that is 1.21 m in diameter and 0.254 m deep placed 0.15 m above ground level on an open timber framework. Due to differing thermal characteristics between the Class A pan and large water bodies, evaporation from surface pans exceed the total amount of evaporation from large water bodies and distort the seasonal distribution. On a seasonal basis, pan evaporation usually peaks several months before the peak evaporation from deep lakes. Estimates based on pan evaporation are questionable during late fall and early winter periods when ice may form in the pan but not on large water bodies.

Annual lake evaporation estimates are usually obtained by multiplying the annual pan data by an appropriate coefficient. These coefficients have been computed for a number of water bodies and for a US NWS Class A pan and tend to range in value from 0.65-0.85 (U.S. Department of Commerce 1968; World Meteorological Organization 1973). The coefficient is higher under humid conditions and lower under arid or dry conditions (fig 5). A coefficient of 0.7 is applicable when water and air temperatures are approximately equal (Kohler et al. 1955, 1958).

Mean monthly, seasonal, and annual Class A pan evaporation for individual stations throughout the United States have been tabulated by Farnsworth and Thompson (1982) and Farnsworth et al. (1982). The second reference *Evaporation Atlas for the Contiguous 48 States* contains maps with isolines of pan coefficients recommended for use with the evaporation maps in estimating so-called *free water surface evaporation* (FWS) from shallow water bodies. Fig. 9 presents a portion of the Annual FWS Evaporation map for the 48 States from that publication. Seasonal maps for the period May-Oct are included in this NWS report. Farnsworth et al. (1982) defined FWS as evaporation from a thin film of water having no appreciable energy storage. They state that: "*Only when the change in heat storage is negligibly small will FWS be a good estimation of the evaporation from the lake*." Farnsworth et al. did not address the problem of solar radiation that penetrates beneath the 'thin film'. The most reliable map in this publication is Map 2 covering the period May-October, the warmer period when pan evaporation data were available. A map for November-April, not published, was developed and the two seasonal maps were *graphically* added to obtain annual values (Map 3). In the northern part of the US and especially in higher western areas, a minimum value of 1778 mm (7 inches) was estimated for the November through April FWS evaporation. Stauffer (1991) estimated vaporization and sublimation during ice cover for Lake Mendota in Wisconsin to be about 60 mm, but no time period was given.

Many reservoirs are operated using estimated evaporation based on measured pan evaporation from the U.S. Weather Service Class A pan. The guidelines for siting and maintenance of evaporation pans, however, have not always been followed closely. Some pans have had screens placed on them to keep birds and animals from drinking the water. Even heating devices have been used to keep ice from forming on the pan in the spring and fall. Some pan sites have been moved to new locations to facilitate observations. Some of these sites do not meet the criteria for a representative pan evaporation site for which the original coefficients were developed, and some sites are not well-maintained.

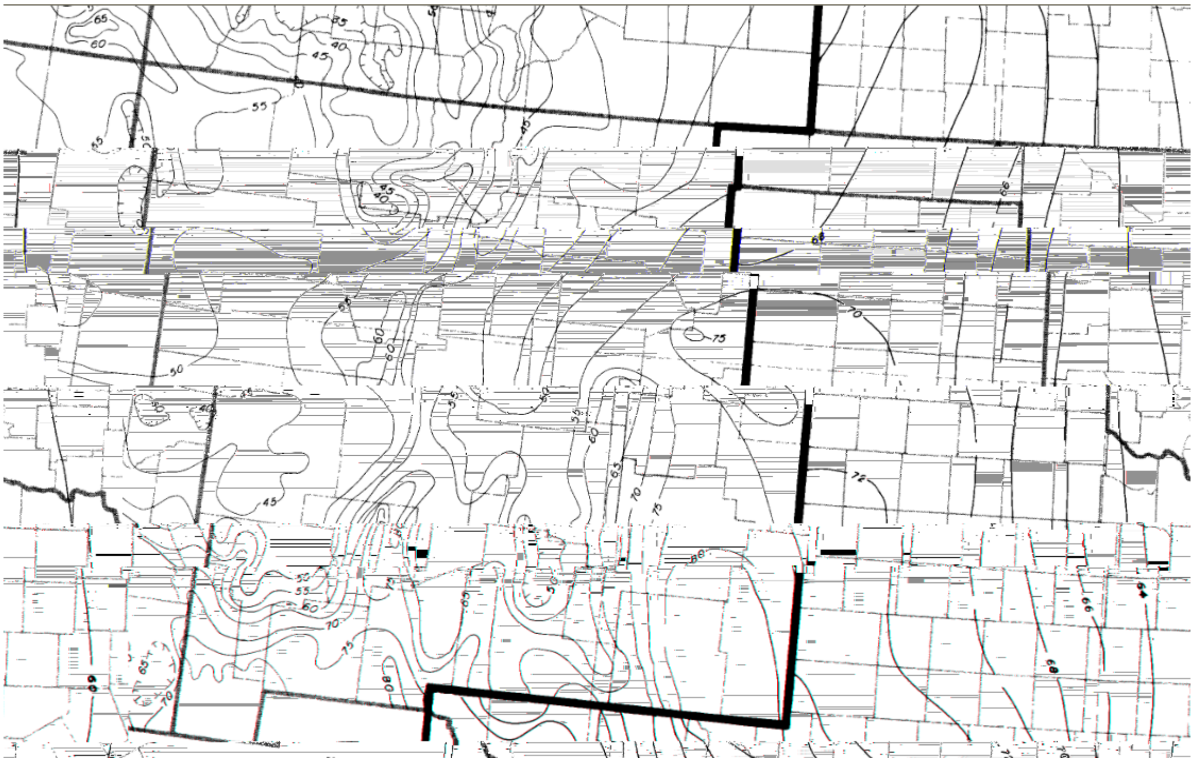

Fig. 9. A portion (NW USA), of a map of estimated free water surface (FWS) Annual Evaporation for the 48 States. Source: Farnsworth et al. (1982), http://www.weather.gov/oh/hdsc/studies/pmp.html

Pan Fetch Affects. Young (1947) discussed the problem of local pan environment in relation to estimating lake evaporation. Soon thereafter studies in India by Ramdas (1957) and studies by Pruitt at Prosser, Washington and in California (Pruitt 1960, 1966 and California, State of 1975) provided clear evidence that unless the local environment of the pan was taken into account, the estimation of lake evaporation was subject to errors of up to 35%. Fig. 8 summarizes some of the early results, which illustrate the problem associated with the environment just upwind of the pan. These results came from a study involving four Class A pans located within a large fallow field, with three of them placed within various-sized circular areas, flood irrigated frequently and planted to grass. The fifth pan, located within a 5-ha irrigated grass field, had a minimum upwind fetch of grass or irrigated pastureland of some 200 m. Data from the one pan in the fallow field, which had no surrounding grass, was plotted as if having a 0.3-m fetch of grass in order to use a log-linear plot. From this and other similar studies (e.g. Ramdas 1957, Pruitt 1960 and Stanhill 1962), Pruitt developed recommended pan coefficients (K_p) for estimating ET_o and in turn evaporation from shallow water bodies. Taken into account were the effects of upwind fetch (both dry and moist), mean relative humidity and total daily wind on K_p (Jensen 1974; Doorenbos and Pruitt 1975, 1977; Jensen et al. 1990)

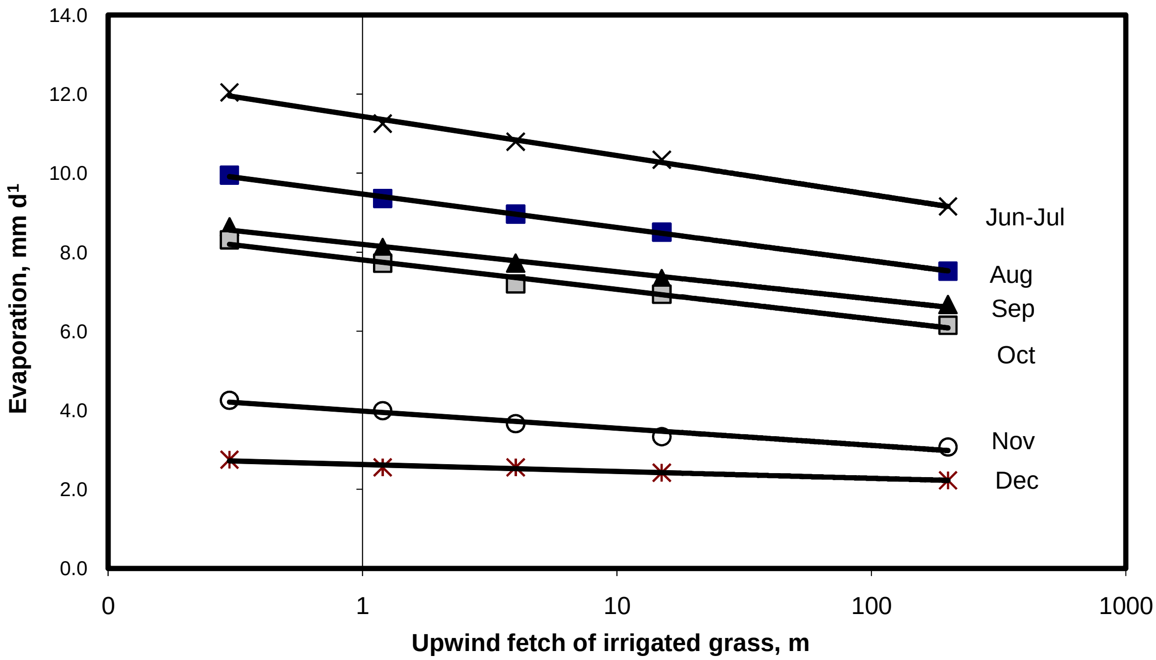

Fig. 8. Evaporation from NWS Class A Pans vs. upwind fetch of irrigated grass 0.07 to 0.15-m tall, Davis, California, 1959. The upwind fetch effect decreases from June to December. (Redrawn from Pruitt 1966).

Water Budget Procedure. The water budget procedure is a simple method for computing evaporation for monthly, seasonal, or annual time periods. All of the terms in the water balance, except for evaporation, are either measured or estimated with the evaporation computed as a residual. The water balance procedure is based upon the hydrologic equation

$$Inflow = Outflow + \Delta Storage \tag{4}$$

Inflow terms consist of precipitation on the water surface, runoff from the land basin, major channel inflow from outside the immediate drainage basin, groundwater inflow, and diversions into a free water body (lake, reservoir, pond, etc.) from outside its basin. Outflow terms consist of evaporation from the water surface, major channel flow out of the water body, diversions out of the water body, and groundwater flow from the water body. The water balance equation can be expressed in terms of depth or volume of water evaporated from the surface per unit time as:

$$E = P_r + R + Q_i + G_{gw} + D_i - Q_o - D_o - \Delta S \tag{5}$$

where E is the amount of water evaporated, P_r is the precipitation on the water surface, R is the runoff inflow from the drainage basin, Q_i and Q_o are major channel flows into and out of the water body, respectively, and ΔS is the change in water body storage. Units for all parameters in eq. 5 are the same and can be expressed in terms of volume or in units of depth (i.e., in mm) over the water body surface for a specified time interval. The computations are most accurate when the evaporation is of the same order of magnitude or larger than the other terms of the water balance. The groundwater component is usually the most difficult term to estimate. For most lakes the volume in storage and surface area, is required to extrapolate volumes to depths.

Water balance data sources are the National Climatic Data Center, NOAA regional data centers, and state climatologists for precipitation data, the U.S. Geological Survey, State and Provincial water agencies, and flood and conservation districts for runoff, major channel flows, and diversion data.

OTHER METHODS

Aerodynamic methods are among the most widely applied to calculate evaporation from large lakes and reservoirs. It will not be discussed in this paper.

The energy balance method as applied to water bodies is based upon conservation of energy principles. As with the water balance procedure, the evaporation is computed as a residual. This procedure is the most data intensive of the standard evaporation procedures, but it has wide applicability to many differing water bodies for time periods of minutes to years. The energy balance for a water body may be expressed as

$$Q_t = R_n - \frac{\lambda \rho_w E}{1000} - H + Q_V - Q_w \tag{6}$$

where Q_t is the change in energy stored in the water body in MJ m^{-2} t^{-1}, R_n is net radiative energy to the water body in MJ m^{-2} t^{-1}, $\lambda\rho_w E/1000$ is the energy utilized by evaporation in MJ m^{-2} t^{-1}, ρ_w is density of liquid water in kg m^{-3}, λ is the latent heat of vaporization in MJ kg^{-1}, E is the evaporation rate in mm t^{-1}, H is the energy convected from the water body as sensible heat in MJ m^{-2} t^{-1}, Q_v is the net energy advected into the water body by streamflow or ground-water in MJ m^{-2} t^{-1}, and Q_w is the energy advected by the evaporated water in MJ m^{-2} t^{-1}.

As mentioned, an important distinction between R_n for a water body and R_n from vegetation or soil is that with soil and vegetation, essentially all of the R_n quantity is captured at the 'opaque' surface and is immediately available for conversion to λE or H or conduction into the surface as G. For a water body, however, much of the solar radiation, R_s, penetrates to some depth in the water body, depending on the turbidity of the water, where it is converted to Q_t. Therefore, with water bodies, R_n is the net radiation captured by the water body, but is not necessarily available at the surface for immediate conversion to λE or H. This is why the energy storage change, Q_t, is used in the energy balance for water rather than the ground energy flux density term, G, that is used for soil. The term G for land areas is governed only by thermal conduction into the surface whereas Q_t is governed by both conduction and by penetration of solar radiation.

A common method for calculating R_n for water with commonly used nomenclature can be expressed as:

$$R_n = (1-\alpha) R_s \downarrow + (1-\alpha_l) R_L \downarrow - \varepsilon\, \sigma (T_s)^4 \uparrow \qquad (7)$$

for R_n, R_s and R_L in MJ m^{-2} t^{-1} where α is the solar albedo for water (0.04-0.15), dependent upon surface conditions (Brutsaert 1982; Bolsenga and Vanderploeg 1992), α_l is the long wave albedo, equivalent to (1 - ε) or 0.03 (WMO 1970), σ is the Stefan-Boltzmann constant in 4.901 $\times 10^{-9}$ MJ m^{-2} K^{-4} d^{-1} for daily time steps and 2.042 $\times$ 10^{-10} MJ m^{-2} K^{-4} h^{-1} for hourly time steps, ε is the long wave emissivity (0.92-0.97, with ε = 0.97 typically used (Anderson 1954) and T_s is the water surface temperature in K.

Albedo varies with water turbidity, wave height and the ratio of diffuse to total solar radiation. Cogley (1979) presented a table of albedo for water surface as a function of latitude based on an accurate weighting for radiation received at different elevation angles. These were based on the Fresnel equation for a plane surface of pure water under direct radiation and modified by measurements made by Grishencko (1959) for wave heights of 0.1 to 0.7 m and cloud cover of 0 to 25%. A summary of these values was presented in table 1.

One of the more difficult problems in applying the energy balance or the combination approach to estimating evaporation is that of estimating energy stored or released from the water body when no water temperature data are available. The data shown in fig. 1 and 2 illustrate typical thermal profiles encountered as solar radiation is increasing during the year and after solar radiation begins to decrease. About 50% of total solar radiation penetrates to a depth of 1 m and 20% to a depth of 10 m in pure water. Penetration is strongly wavelength dependant. The depth of penetration is also influenced by the turbidity of the water. Stauffer (1991) reported that 90% of the visible light flux was absorbed in the top 2 m of the water column in Lake Mendota. The thermal conductivity of water is not high and most of the increase in temperature is due to absorbed solar radiation that decreases exponentially with depth. At the end of the season as solar radiation has decreased and with the loss of energy due to evaporation, long-wave radiation emitted from the water surface and the transfer of sensible heat to the air, the surface of the water cools (fig. 2). Then, because of the higher density of the cooler water, it settles until it reaches the same temperature lower in the profile. The astute observation made in 1898 by Carpenter that the water in evaporation tanks is nearly uniform in the morning while a thermal gradient exists during the day illustrates this physical phenomenon of cooled water settling and mixing the water in the container.

For shallow lakes and reservoirs with depths of less than 1 m, measuring the water temperature near the surface in the mornings will likely produce a good estimate of average daily water temperature so as to calculate daily changes in energy storage. Daily measurements of near-surface temperature in the morning would enable calculating average rate of energy storage using a constant heat capacity of 4.18 MJ/m^3 in the range of 10 to 25ºC. For deeper water bodies, > 2-10 m, the trend in average daily change in energy storage measured in Pretty Lake Indiana and Williams Lake in Minnesota illustrates the magnitude of energy storage and release rates that can be expected (fig. 10). The values for Q_t shown in fig. 10, especially during early spring, late summer and fall are relatively large, averaging about ± 5-10 MJ m^{-2} d^{-1}, when R_n during these periods averages only 5 to 20 MJ m^{-2} d^{-1}.

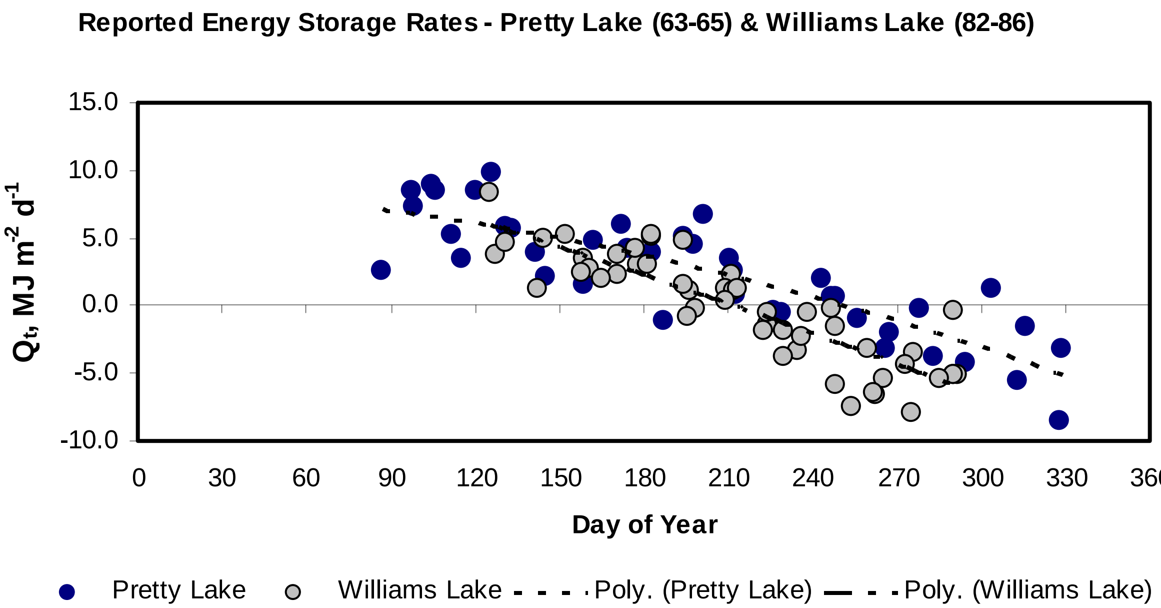

Fig. 10. Reported energy storage rates in Pretty Lake, Indiana (1963-1965) and Williams Lake, Minnesota (1982-1986) (Jensen et al. 2005).

Jensen et al. (2005) estimated daily energy storage based on Pretty Lake, Williams Lake and Lake Berryessa based on net short wave and long wave radiation:

$$Q_t = 0.5\,R_{sn} - 0.8\,R_{nl} \text{ , for Day of year <180} \tag{8a}$$

and

$$Q_t = 0.5\,R_{sn} - 1.3\,R_{nl} \text{ , for Day of year >180} \tag{8b}$$

where Q_t = average daily energy storage in the water body in MJ m^{-2} d^{-1}, R_{sn} = net solar radiation $(1 - \alpha)R_s$, and R_{nl} = net outgoing long-wave radiation [$R_{nl} = \varepsilon_l R_l\uparrow - (1 - \alpha_l)R_l\downarrow$]. Williams Lake is a small lake (36 ha) with a maximum depth of about 10 m. Pretty Lake has an area of about 75 ha, a maximum depth of about 25 m, and an average depth of about 8 m because of the large shallow areas. Lake Berryessa has an area of 8,320 ha when full, a maximum depth of 58 m near the dam, and an average depth of about 40 m. The r^2 for both equations combined was 0.63. A comparison of estimated Q_t from eq. 9 vs. reported or measured Q_t is shown in fig. 11.

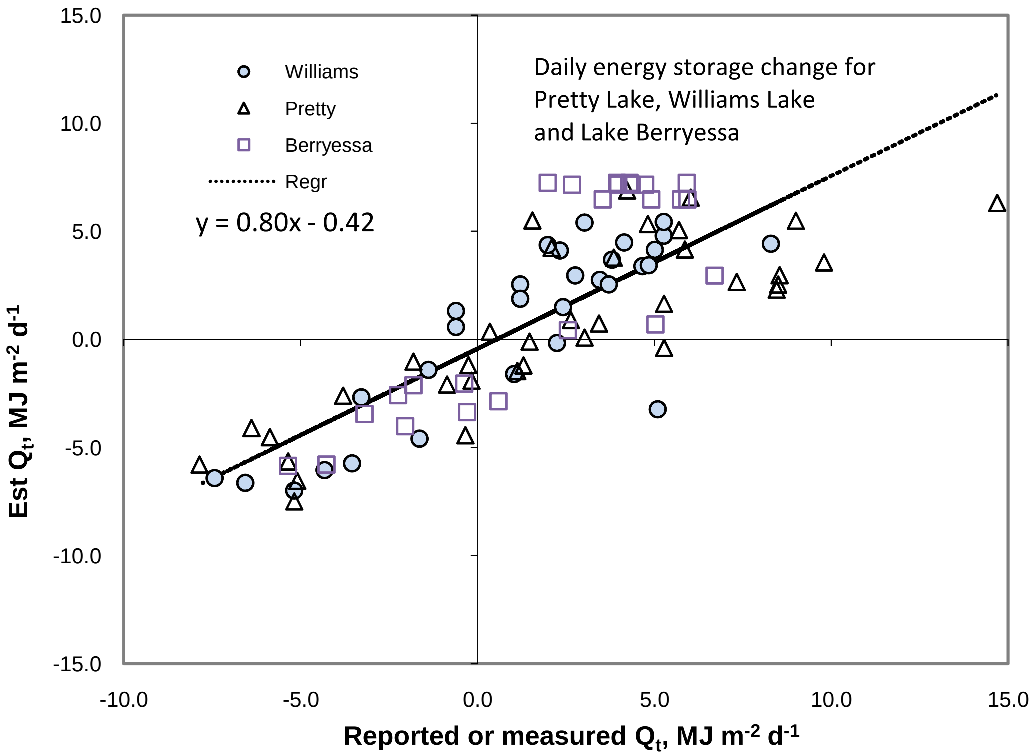

Fig. 11. Estimated daily Q_t from eq. 12 vs. reported or measured Q_t for Pretty Lake, Indiana (1963-1965), Williams Lake, Minnesota (1982-1986) and Lake Berryessa, California (2003-2004.[5]

Data needed for the energy balance method can be obtained from the National Climatic Data Center for radiation components, from satellite analysis, municipal water intakes or in situ measurements for water temperatures, and from thermistor strings or water intakes for the changes in energy storage. An example detailed energy budget for Lake Ontario is given by Aubert and Richards (1981) and includes an evaluation of energy terms.

Combination Methods

For many applications a combination of the aerodynamic and energy balance procedures is desirable. The most common of the combined equations was derived by Penman (1948, 1956) which in its general form is written as:

$$\lambda E = \frac{\Delta (R_n - Q_t) + \gamma E_a}{\Delta + \gamma} \tag{9}$$

[5] (Jensen et al. 2006, pers. comm.

where λE is the latent heat of evaporation in MJ m^{-2} t^{-1}, γ is the psychrometric constant in kPa $^{o}C^{-1}$, Δ is the slope of the saturation vapor pressure curve at temperature T_a in kPa $^{o}C^{-1}$, $(R_n - Q_t)$ is the net radiation minus the change in energy storage in MJ m^{-2} t^{-1}, and E_a is a bulk aerodynamic expression in MJ m^{-2} t^{-1} containing an empirical wind function.

$$E_a = 6.43\left(a_w + b_w u_z\right)\left(e_z^o - e_z\right) \tag{10}$$

where a_w and b_w are empirical wind function coefficients and u_z is wind speed at the z height, m s^{-1}. Variables e_z^o is the saturation vapor pressure and e_z is the actual vapor pressure at height z. The value 6.43 is for λE in MJ m^{-2} d^{-1}. For λE in MJ m^{-2} h^{-1}, the factor becomes 0.268. Various values for the empirical coefficients a_w and b_w have been proposed (Penman 1948). Penman (1948) initially proposed $a_w = 1.0$, but later revised it to 0.5 (Penman 1956, 1963), and $b_w = 0.54$ s m^{-1} for open water for $z = 2$ m and for e_z^o computed using mean daily air temperature.

The roughness height, z_{om}, for use with the PM equation can be taken from the range 0.0001 to 0.0006 m, with generally z_{oh} similar in value to z_{om} or even exceeding the value for z_{om} when applied to water (Brutsaert 1982). For detailed, theoretical aspects of relationships between z_{oh} and z_{om} for water, the reader is referred to Brutsaert (1982).

An important point to consider in use of combination equations (and others as well) is that the use of weather data collected over a typical weather station on land cannot be expected to give as reliable values as computations using data collected over the water surface. For one, the wind at a height z over aerodynamically smooth water may be significantly greater than wind at height z over a typical weather station site. Temperature and humidity may also be quite different over the two sites.

Simulation Studies

For many complex studies such as determining the impacts of potential climate change or variability on lake evaporation, it is necessary to combine the energy balance and mass transfer procedures (Croley 1989; Hostetler and Bartlein 1990). In this type of approach the mass or bulk transfer equations are substituted for λE and H in the energy balance and solved for the water temperature T_s and the resulting energy stored in the lake, Q_t. The evaporation is determined using the computed T_s in the bulk transfer equation. Depending upon the accuracy desired and the overlying assumptions, many simplifications may be employed in the procedure. This type of technique is currently being applied in many water resource and climate-change studies (Henderson-Sellers 1986; Henderson-Sellers and Davis 1989).

Because of the complexity in accounting for changes in energy storage when there is significant inflow and outflow to the water body such as reservoirs within river systems, computer models for calculating the reservoir thermal structure may be used. One such model, CE-THERM, was developed by the U.S. Corps of Engineers (USACE 1995). Use of the CE-THERM model to evaluate the effects of large flow releases in 1995 on the physiological rates of organisms in the Blue Mesa Reservoir located in southwest Colorado was described by Johnson et al. (2004). The model was calibrated based on several daily time steps by adjusting radiation and mixing parameters during the calibration period. Turbidity, or Sechi depths, and temperature profiles were measured biweekly during 24-May to 16-Sep period in 1994. Temperature measurements at three stations showed only minor differences in the surface layer of the stratified water body in 1995. Temperature profile measurements in front of the dam and at a mid-lake location in 1995 were nearly indistinguishable. Data from 1987 were used to estimate total dissolved solids (TDS). Meteorological data from the Gunnison Airport, about 25 km upstream were used. Based on daily wind speed measurements at the airport and reservoir in 1996, wind speeds were increased by a factor of 1.7 to account for local topographic conditions.

ESTIMATING EVAPORATION FROM SHALLOW LAKES

Applying a Combination Method

To apply a combination method, energy storage and water surface temperatures are needed. In many areas of the US, there are now many automated weather stations (AWS), especially in irrigated areas. Therefore, daily weather data, including solar radiation, humidity and wind speed, can be obtained for estimating evaporation. Data obtained from an AWS should be screened for quality before being used. For shallow lakes and reservoirs of less than 1 m, measuring the water temperature near the surface in the mornings will likely produce a good estimate of the average daily temperature of the water body because of redistribution of water at night. Daily measurements of near-surface temperature in the morning would enable calculating average rate of change in energy storage knowing the depth of the lake and using a constant heat capacity of 4.18 MJ/m^3 in the range of 10 to 25ºC. For depths of 2-3 m thermal profiles will develop and periodic temperature measurements at lower depths would be needed to enable estimating energy storage.

The surface temperature during sunny days will be higher than the morning value. For reliable estimates, use of a floating digital temperature-time recorder should be considered so as to estimate the saturated vapor pressure at the surface water temperature. With these data and albedo from table 1, net radiation can be calculated and the Penman-Monteith equation used to estimate daily evaporation. Periodic measurement of high surface water temperature in mid-afternoon along with the morning near-surface temperature could be used to estimate mean surface water temperature for calculating the saturated vapor pressure at the water surface.

Another alternative to estimate evaporation from a shallow water body is to use the ASCE-EWRI reference ET equation for daily, short crop, or grass, and a coefficient for open water.

$$ET_{ref} = \frac{0.408\,\Delta\,(R_n - G) + \gamma\,\dfrac{900}{T + 273}\,u_2\,(e_s - e_a)}{\Delta + \gamma\,(1 + 0.34\,u_2)} \tag{11}$$

where ET_{ref} applies to a clipped grass reference surface. ET_{ref} has units of mm d^{-1} for 24-hour time-steps. R_n and G are in MJ m^{-2} d^{-1}, T is mean daily (°C), u_2 is mean daily wind speed at 2-m height (m s^{-1}), e_s and e_a are in kPa, Δ and γ are in kPa °C^{-1}.

With calculated ET_{ref}, a coefficient for open water, K_w is then applied to estimate evaporation.

$$E = K_w \, ET_{ref} \tag{12}$$

where K_w is the coefficient for open water and ET_{ref} is reference ET for short Grass (Allen et al. 1998). For open water depths of less than 2 m, FAO-56 suggest a K_w value of 1.05. This a comparison with PM estimates for Home Lake indicates that a value of 1.10 may be better.

Evaporation Pan Method

The most common approach used in the US is to apply a coefficient to measured pan evaporation.

$$E = K_p \, E_{pan} \tag{13}$$

where K_p is the pan coefficient and E_{pan} is the evaporation from a Class A pan. Kohler et al. (1955) suggested that the 0.70 coefficient is applicable to the Class A pan when when average air and pan-water temperature are equal as did Farnsworth et al. (1982) for May-October. Obviously, pan evaporation data cannot be used during freezing temperatures. Farnsworth et al. (1982) assumed that a minimum value of 179 mm (7 inches) would hold everywhere and this value was as a minimum their Nov-Apr map. The two maps, May-Oct and Nov-Apr, were added graphically to obtain their annual free water surface map (FWS). They suggested that annual evaporation from their annual FWS map can be used when there is negligible heat (energy) storage and the energy content of inflow and out flow waters are essentially the same. They also stated that seasonal values cannot be used for estimating actual lake evaporation unless the chnages in energy storage and the difference in energy inflow and outflow are properly accounted for.

For Colorado, the best values to use for K_p are those derived from Rohwer (1931) because of very detailed measurements made on the adjacent Class A pan and the 85-ft diam reservoir. Using the polynomial equation in fig. 6, the suggested mean monthly values for K_p that adjust for differences between the pan and reservoir (energy storage):

Month	K_p
April	0.60
May	0.63
June	0.67
July	0.71
August	0.75
Sepember	0.78
October	0.77
Average	0.70

Example Application of the Combination Method of Evaporation from a Shallow Lake

Using daily climate data for 1996, estimates of May-October evaporation from Home Lake near Alamosa, Colorado were made using the PM equation[6]. Home lake is located about 2,297 m (7,536 ft.) above sea level. During 1996, the depth of Home Lake varied from 0.3 to 3 m and averaged about 1.5 m. Water temperature measurements were made at the 0.05-m, 0.3-m and 0.6-m depths at about 0800 and about 1630. The near-surface temperatures increased about 4 to 8ºC during the 8-hour daytime period. Some solar radiation likely penetrated to the bottom of the shallow lake. As surface water cooled at night due to evaporation and net long-wave radiation, the heavier cold water settled (Carpenter 1898), thereby mixing the water in the lake. Using a regression of the 0.3-m depth temperature and air temperature, the estimated daily average temperature at 0.3 m plus 3°C was assumed to represent daily average surface temperature. The average daily water temperature was used to estimate the change in daily energy storage. The mean air temperature and estimated average water temperature are shown in fig. 12. The lake surface was frozen by Oct. 29. Estimates of daily evaporation-sublimation was assumed when the estimated water or ice surface temperature was 0°C. During that period, albedo was assumed to be 0.70 and surface emittance was assumed to be 0.97. My estimate of evaporation-sublimation 284 mm (11.2 in.) for Nov-Apr and my estimate of evaporation was 894 mm (35.2 in.) for May-Oct. and for an annual total of 1,178 mm (46.4 in.). Map (Plate) 2 of Kohler et al. (1959) shows about 1,090 mm (43 inches) for Alamosa, CO. Map 2 of Farnsworth et al. (1982) indicates about 890 mm (35 inches) for May-Oct and map 3 indicates about 1,270 mm (~50 in.) for the San Luis Valley of Colorado which is about 380 mm (15 inches) more than the May-Oct. period.

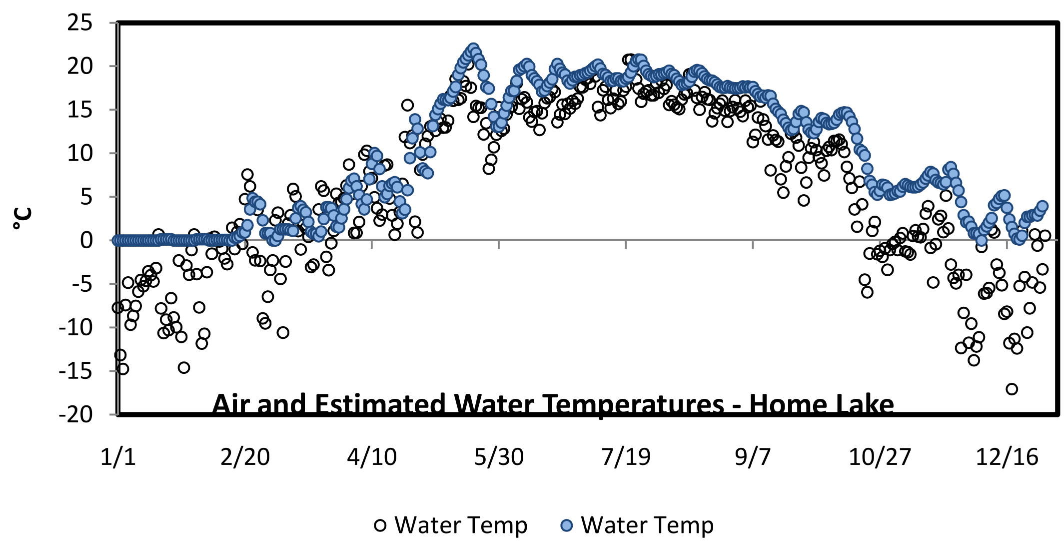

Fig. 12. Mean air and estimated water temperature in Home Lake in 1996.

The estimated albedo varied from 0.06 to 0.08. Net long wave radiation was estimated using estimated surface water temperature and Brutsaert's eq. 6.18 (Brutsaert 1982, p 139). The

[6] Jensen, M.E. 2006. Personal communication. Based on data from L. Salazar as summarized in a memo from Jensen to Salazar, 21-Jan-99.

variation in estimated daily energy storage change is shown in fig. 13 and was used to estimate (R_n - Q_t). These values for Q_t are much smaller than those shown in fig. 10 because of the shallowness of Home Lake. The resulting average monthly evaporation for April-Oct. is shown in fig. 14. The estimated total evaporation from May through October 1996 was 890 mm (35 in.), essentially the same as in NWS-33 (Farnsworth et al. 1982). The estimated evaporation using FAO-56 ET_o x 1.05 was about 5% lower for April-October. FAO-56 suggests using K_w = 1.05 for shallow water bodies. My data from Alamosa, Colorado indicates that a value of K_w = 1.10 agrees more closely with my PM estimates for April through November.

The resulting estimate of annual evaporation was 1,178 mm (46.4 inches) which may be high due to over-winter estimates. The USWS map 3 (Farnsworth et al., 1982) was about 1,270 mm (50 inches). The difference between the May-October evaporation of about 890 mm (35 inches) and the annual estimate is about 380 mm (15 inches) is about twice the November-April suggested minimum and may have resulted from elevation corrections that may have been applied because of the high elevation.

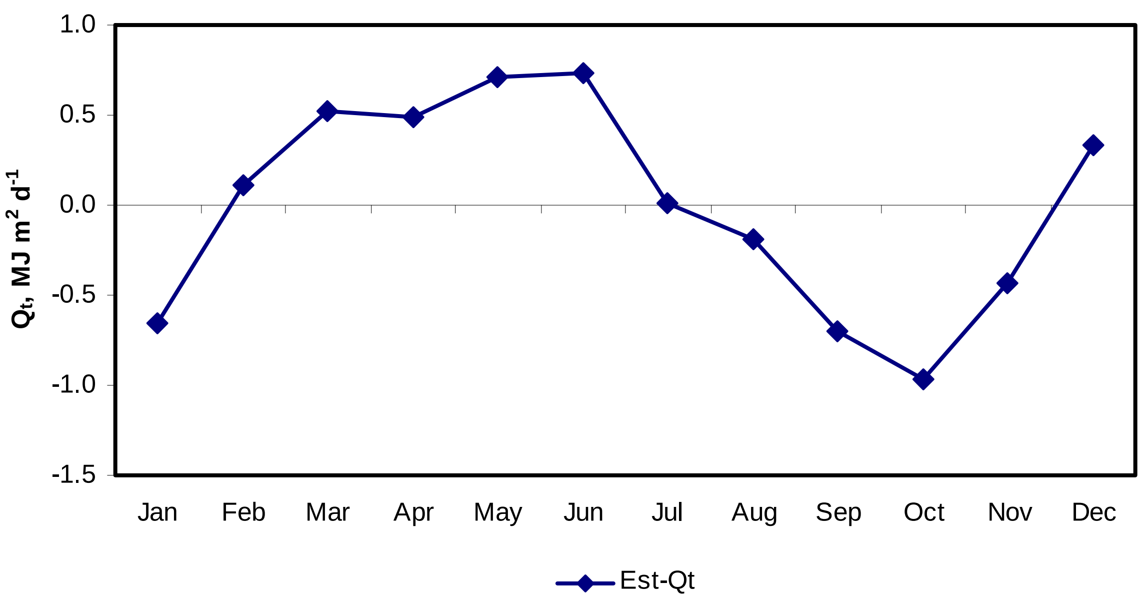

Fig. 13. Estimated energy storage rate for Home Lake assuming an average depth of 1.5 m.

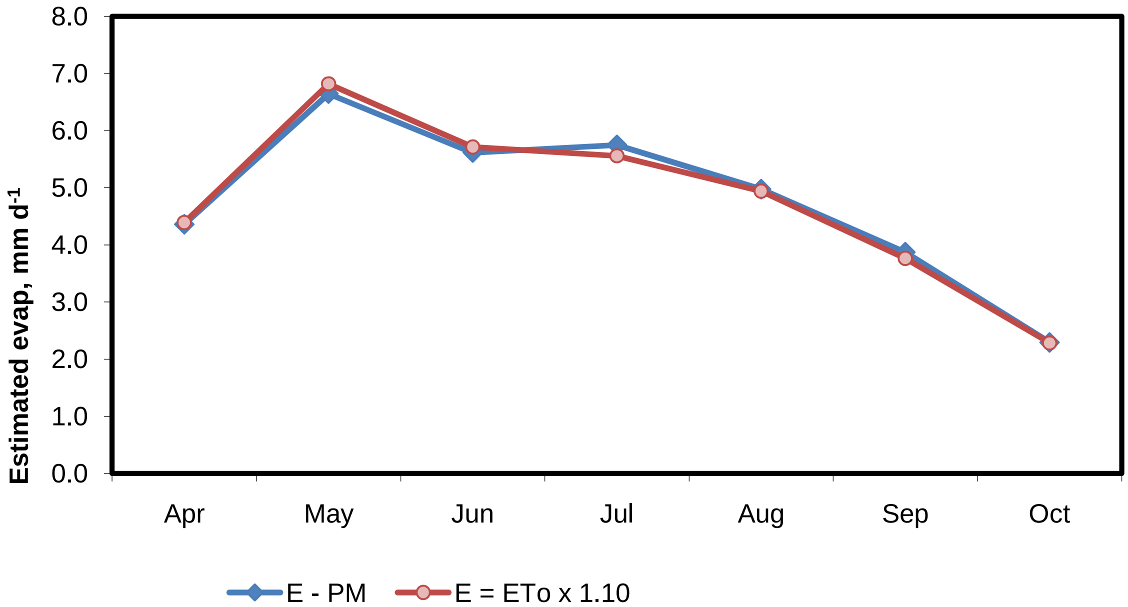

Fig. 14. Estimated evaporation from Home Lake near Alamosa, Colorado in 1996.

Wetlands

Estimates of evaporation and ET on wetlands in temperate climates can be obtained using the ET_{ref} for short grass and the suggested coefficients found in FAO-56, table 12, Allen et al 1998), and in table 8.2 of Design and Operation of Farm Irrigation Systems (Allen et al. 2007). The values are reproduced in table 2.

Table 2. Values of K_cs for the FAO-general crop curve (Allen et al. 1998, 2007).

	$K_{c\ ini}$	$K_{c\ mid}$	$K_{c\ end}$
Cattails, bulrushes (killing frost)	0.30	1.20	0.30
Cattails, bulrushes (no frost)	0.60	1.20	0.60
Short vegetation (no frost)	1.05	1.10	1.10
Reed swamp, standing water	1.00	1.20	1.00
Reed swamp, moist soil	0.90	1.20	0.70

Summary—Estimating Evaporation

General. Evaporation is rarely measured directly, even in small water bodies. It is usually estimated by association with measured evaporation from evaporation pans or calculated by water balance, energy balance, mass transfer or a combination of energy balance and aerodynamic techniques. The method selected depends on the depth of the water body and the availability of weather data or micrometeorological equipment.

Small Water Bodies. For small water bodies such as shallow lakes, the most widely used method is to multiply monthly coefficients by measured pan evaporation. The most commonly used pan in the USA is the Weather Service Class A pan. The accuracy of this method is related to the environment surrounding the pan as illustrated in fig. 8. Recommended monthly pan coefficients for shallow water bodies in Colorado are based on detailed Class A pan and reservoir evaporation studies conducted at Fort Collins, CO. For other areas, pan coefficients can be obtained from Map 4 for May-October in Farnsworth et al. (1982). Evaporation estimates for some states may be available such as that by Borrelli et al. (1998) for Texas. Local or regional calibration or verification of the coefficients used is recommended when adjusting for poor pan siting conditions.

Evaporation from a shallow water body can be estimated using a combination method following procedures similar to those described for a shallow lake or the Penman equation (Penman 1956, 1963), after adjusting for energy storage, with wind speed coefficients of $a_w = 0.5$ and $b_w = 0.54$ for wind speed in m s^{-1} measured at about 2 m above the water surface and saturation vapor pressure based on mean air temperature.

Evaporation from shallow water bodies can also be approximated by multiplying reference ET_{ref} for short grass by a coefficient of 1.05 (Allen et al. 1998). Monthly pan coefficients calculated using the high elevation shallow lake example in this chapter that were calculated using the PM equation increased linearly from 1.1 in May to 1.15 in October. The total May-Oct evaporation values for this example by different methods were similar, indicating that any or a combination of methods would have provided reasonable values:

Home Lake:	Estimating Method				
	PM	Penman	NWS-33	ET_o x 1.10	Kohler et al. 1959
Est. evaporation, mm	894	906	890	892	787
Est. evaporation, inches	(35.2)	(35.7)	(35.0)	(35.1)	(31)
Est, percentage of PM	100	101	100	99.8	88

Lake Berryessa:	Estimating Method			
	0.95 PM	Penman	P-T	USBR Orig
Est evaporation, mm	1325	1424	1277	955
Est. evaporation, inches	(52.2)	(56.1)	(50.3)	(37.7)
Est, percentage of PM	100	108	96	72

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