Chapter 8
Lake Evaporation

Abstract Evaporation from lakes depends on available energy, mechanism of momentum, mass, and energy transfers. The surface area of the lake determines the effect of advective energy from the surrounding environment. The depth of the lake is critically important for heat storage and release. Various approaches have been used to develop lake evaporation equations that address the factors that affect the rate. In this chapter, the evaporation process and methods of estimation are presented in detail. Actual measurements of evaporation are used to calibrate models from the simplest to the most complex. The pan method, water budget method, energy balance methods, mass transfer methods, Bowen ratio method, Penman method, and radiation-based and radiation–temperature-based methods are presented in detail.

Keywords Lake evaporation • Pan evaporation • Energy balance • Evaporation estimation methods

8.1 Introduction

Lakes and reservoirs are sources of water supply, fishing for food supply, recreation, hydropower, and transportation and maintain wetland and aquatic ecosystems. Lakes' and reservoirs' evaporation rates are of great interest for water resources management. A major cause of decline in water level for lakes and reservoirs is evaporation. With ever increasing water demand, measuring or estimating evaporation rates is important. Especially in arid and semiarid areas, evaporation is a major flux in the hydrologic cycle. Lake evaporation direct measurements are generally not easy nor are reliable data available. In most cases, lake evaporation is estimated using pan evaporation, water budget, or from empirical models that mostly use data collected on land. Evaporation from lakes depends on the availability of energy and the mechanisms of mass and energy transfer. The surrounding environment of the lake, the in-lake environment, depth, and surface area of the lake affect the rate of evaporation. The effect of depth of lake on evaporation has been discussed.

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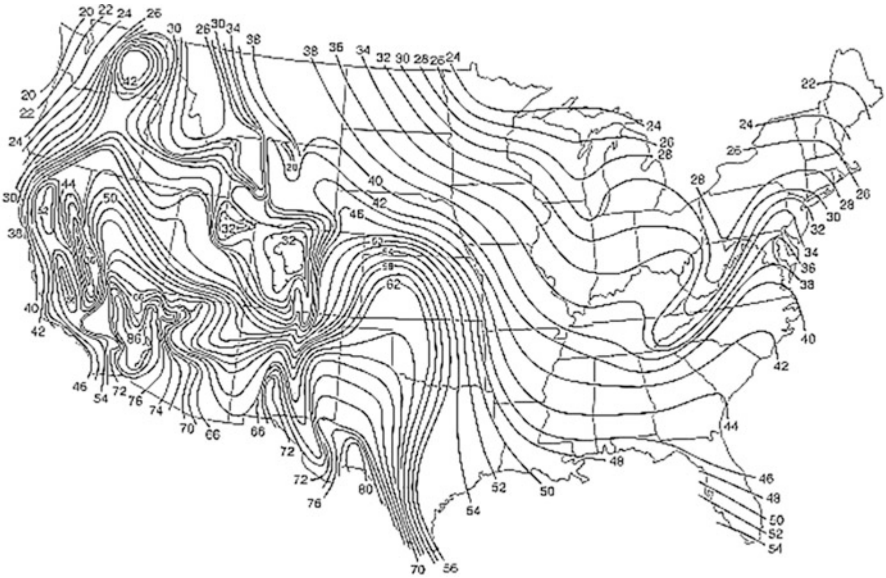

Fig. 8.1 Mean annual lake evaporation in inches over the USA (Kohler et al. 1959; U.S. Weather Bureau); 1 in. = 2.54 cm

Lake depth is suggested to be more important at higher latitudes where there is a more marked difference between summer and winter temperatures than in the tropics (Reis and Dias 1998). The importance of the vertical temperature profile and subsurface heat storage changes and depths of various lakes are presented stating that deeper depths reduce annual lake evaporation. Evaporation is a function of solar radiation, temperature, wind speed, vapor pressure deficit, atmospheric pressure, and advective and storage energy sources and sinks. Most of these variables are latitude and altitude dependent.

Annual lake evaporation in the United States varies from 51 cm in the northeast and 218 cm in southern California. A U.S. Weather Bureau 1959 technical paper provides a map, Fig. 8.1, with mean annual lake evaporation for the contiguous United States (Kohler et al. 1959). Generally, the south and southwest have higher evaporation than the north and northeast. Roberts and Stall (1967) produced a lake evaporation isohyetal map for Illinois showing a general increasing trend from the northeast to the southeast ranging from 76 to 97 cm. Based on lysimeter measurements and literature review, isohyetal lines for lake evaporation in south Florida are shown in Fig. 8.2 (Abtew et al. 2003).

Lake Okeechobee is located in the central region of south Florida at 26° 39′ and 80° 37′ longitude. The lake has a surface area of 1,732 km^2 and an average depth of 2.7 m (Jin et al. 1998). Based on 5 years of meteorological data applications in evaporation models and water budget analysis, annual average evaporation of 132 cm was reported for Lake Okeechobee (Abtew 2001). This estimate is close

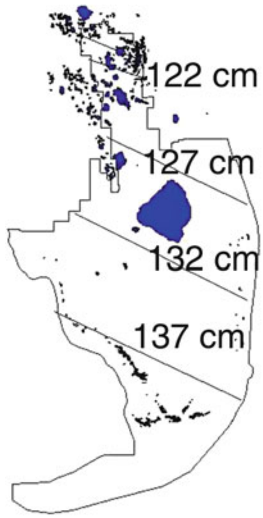

Fig. 8.2 Estimated isohyetal lines for open water evaporation, wetland evapotranspiration, and potential evapotranspiration for south Florida (Abtew et al. 2003)

to what is presented in Figs. 8.1 and 8.2. In a study on a temperate lake, Lake Sparkling in northern Wisconsin, evaporation and energy balance are presented by Lenters et al. (2005). It was reported that humidity (32%) followed by the difference in water and air temperature (22%) explains most of the interseasonal variation in lake evaporation. The results of this study show the differences in subtropical and tropical lakes' energy balances and evaporation variation. It is also reported that wind and interseasonal lake evaporation show little correspondence in that climate and mass transfer evaporation estimations were less accurate due to the inclusion of wind speed.

8.2 Lake Evaporation Estimation Methods

8.2.1 Pan Method

The most common lake evaporation estimation method is the pan method where evaporation from a small pan is related to evaporation from a lake through a pan coefficient. Various types of pans are used in different parts of the world. A common pan is the class A evaporation pan of the National Weather Service in the United States. The pan is 120.7 cm in diameter and 25 cm in depth. Water is added or removed to maintain water level at 5 cm from the rim. The pan is usually accompanied with a rain gauge to factor out the contribution of rainfall to the depth of water in the pan. The sunken Colorado pan is square in shape (100 cm × 100 cm), 50 cm deep, and buried in the ground to a depth of 45 cm. Variations between pans include setup, pan environment, measurement errors, and differences in operations

(Abtew et al. 2011). Details on evaporation pans are presented in Chap. 3. The process of acquiring evaporation estimates from a pan can be presented with a mass balance equation, Eq. 8.1:

$$E_{\text{pan}} = D_{t-1} - D_t + R_{\text{f}} - L \pm e \tag{8.1}$$

where D_t is current day depth of water in the pan and D_{t-1} is previous day depth of water measured from the top, R_{f} is rainfall, L is other losses such as bird or animal consumption, and e is errors. Sources of error in monitoring evaporation with an open outdoor pan include environmental factors such as location, wind flow obstruction, advective heat sources or losses in the area surrounding the pan, height of pan, bird guard, rate of windblown sediment accumulation, and frequency of cleanup, reading and measurement errors, and recording errors. Some pans have bird guards of meshed wire cover to deter birds from drinking or bathing. Bird guard was acknowledged for lowering evaporation rates. In an Australian case, a correction factor (7%) has been applied to correct for the effect of bird guard (Gifford et al. 2007).

Lake evaporation is estimated from pan evaporation based on Eq. 8.2, where E_{L} is lake evaporation, E_{pan} is pan evaporation, and K_{p} is pan coefficient:

$$E_{\text{L}} = E_{\text{pan}} \times K_{\text{p}} \tag{8.2}$$

Pan coefficients vary from area to area for multiple reasons. Geographical location is a factor in pan coefficient. Other reasons are variations in pan type, pan setup, pan environment, pan operator's skills, pan maintenance, rainfall measurement, losses, and errors. Wide ranges of pan coefficients have been reported. Abtew (2001) evaluated pan coefficients for Lake Okeechobee in south Florida from seven pan stations around the lake and vicinity areas. An average coefficient of 0.76 was produced from all pan stations with varying coefficients of a low of 0.64 to a high of 0.95 on an annual basis. Boyd (1985) after 1 year of fully controlled experiment in Auburn, Alabama, reported a range of pond to pan evaporation coefficients (0.72–0.90) with an average of 0.81. Morton (1986) applied the CRLE (Complementary Relationship Lake Evaporation) model for 16 lakes in North America and one lake in East Africa. Extracting pan coefficients from the reported pan evaporation and the CRLE lake evaporation estimates result in a range of pan coefficients for the 17 lakes (Table 8.1). Coefficients range from 0.59 to 0.84 with a mean of 0.69. Due to variation in thermal inertia between a lake and a pan, pan coefficients incur error in evaporation estimation for shorter periods as less than a season (Webb 1966).

8.2.2 *Water Balance Method*

Water balance is one of the simplest methods of lake evaporation estimation. As long as there are data on surface water inflows (I), outflows (O), rainfall (R_{f}), water level, and storage (S), evaporation from a lake (E_{L}) can be estimated based on Eq. 8.3:

Table 8.1 Pan coefficients (K_p) derived from published pan and lake evaporation (mm year^{-1})

Lake	E_{pan} (Morton 1986)	Lake evaporation (CRLE, Morton 1986)	K_p
Dauphin, Manitoba, Canada	859	665	0.77
Last Mountain Lake, Saskatchewan, Canada	1,005	695	0.69
Lake Ontario, North America	913	709	0.78
Utah Lake, Utah	1,945	1,235	0.63
Lake Winnemucca, Nevada	2,076	1,319	0.64
Pyramid Lake, Nevada	2,123	1,249	0.59
Lake Hefner, Oklahoma	1,778	1,286	0.72
Silver Lake, California	2,631	1,920	0.73
Salton Sea, California	3,006	1,765	0.59
Lake Victoria, East Africa	1,940	1,624	0.84
Lake Superior, North America	801	528	0.66
Great Salt Lake, Utah	1,438	1,005	0.70
Walker Lake, Nevada	1,930	1,277	0.66
Tulare Lake, California	2,237	1,464	0.65
Buena Vista Lake, California	2,535	1,535	0.61
Elsinore Lake, California	1,800	1,348	0.75
Lake Okeechobee, Florida	2,070	1,624	0.78

$$E_L = I + R_f - O - \Delta S \pm e \pm S_p \tag{8.3}$$

where ΔS is change in storage, e is errors, and S_p is seepage or groundwater movement into or out of the lake. The application of this method is limited by the completeness and quality of both surface and subsurface inflows and outflows into the lake. Rainfall over the lake surface area is usually estimated from nearby rain gauges. The number of gauges and the location of gauges are factors that affect rainfall estimation. Storage in a lake is computed using stage–storage relationships. Based on bathymetry surveys and stage–storage curves or tables, the volume of water in a lake can be estimated from the water level or water surface elevation readings. Stage–storage relationship for Lake Okeechobee in south Florida is shown in Fig. 8.3. Langbein (1951) estimated Lake Okeechobee evaporation for the 1941 through 1947 period using water budget analysis. The annual evaporation estimate of 132.5 cm is comparable to the model estimate of 132 cm (Abtew 2001). Morton (1986) presented water budget evaporation estimate for the 17 lakes cited in Table 8.1 including Lake Okeechobee. Estimates for Lake Okeechobee were higher, 156.7 cm. Mostly, the results are comparable to the CRLE model estimates.

8.2.3 Energy Balance

Energy balance is a method that is applicable to estimate evaporation from lakes. As the water balance method accounts for inflows, outflows, and change in storage

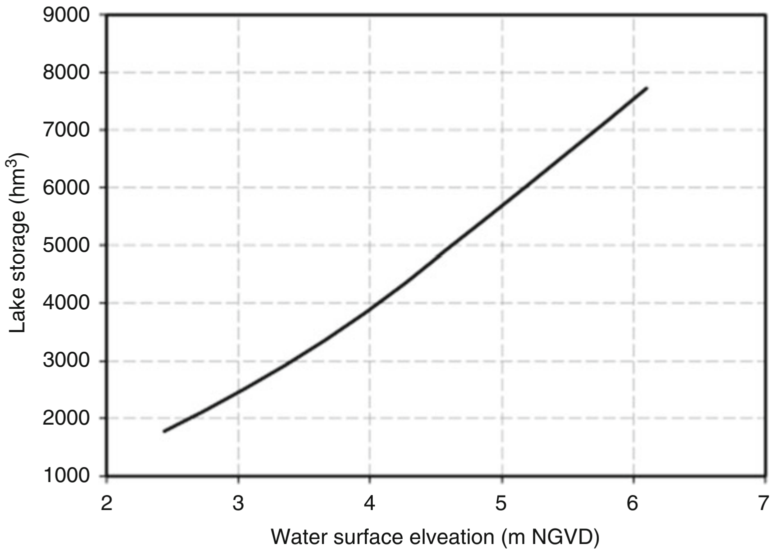

Fig. 8.3 Stage–storage curve for Lake Okeechobee

of water mass, the energy balance method accounts for energy input (Q_{in}) into the lake, energy leaving the lake (Q_{out}), and change in energy storage (ΔQ_s) in the lake. The general equation for energy balance is shown in Eq. 8.4a with error term *e*. Errors are from measurements of each mass balance component and completeness of source and sink accounting in the energy balance equation:

$$Q_{in} - Q_{out} = \Delta Q_s \pm e \tag{8.4a}$$

Energy inflows into the lake water body are shown in Eq. 8.4b:

$$Q_{in} = Q_{R_n} + Q_a + \Delta Q_s + Q_e + Q_h \quad \text{where } (Q_a, \Delta Q_s > 0 \text{ and } Q_e \text{ and } Q_h) \tag{8.4b}$$

where Q_{R_n} is net solar radiation, Q_a is positive net advective energy input, ΔQ_s is positive change in energy storage, Q_e is energy released from condensation at the surface of the lake, and Q_h is sensible heat lost by air at the surface of the lake. Net advective energy is energy input as a balance of energy inflows and outflows associated with surface and groundwater movement into and out of the lake. Energy outflows from the lake are shown in Eq. 8.4c:

$$Q_{out} = Q_e + Q_h + Q_a + \Delta Q_s \quad \text{where } (Q_a, \Delta Q_s < 0 \text{ and } Q_e > 0) \tag{8.4c}$$

where Q_e is energy used for evaporation, Q_h is sensible heat gained by air at the surface of the lake, Q_a is advective energy loss, and ΔQ_s is loss of stored energy. Energy lost by evaporation can be computed by Eq. 8.4d:

$$Q_e = Q_{R_n} - Q_h - Q_a - \Delta Q_s \tag{8.4d}$$

Net solar radiation is measured with instrumentation or estimated from incoming solar radiation. Energy balance of lakes is dependent on season, latitude, altitude, lake depth, surface area of lake, and surrounding environment. Salinity or dissolved solids concentrations are also cited to be a factor (Morton 1986). Winter and Rosenberry (1995) pointed out that one of the contentious issues in energy balance is the time interval of energy balance computation and the sampling of stored energy in the lake. There is uncertainty in the amount of heat stored in the lake at a certain time. Continuous measurement of representative temperature profile will aid in reducing uncertainty in heat storage in the lake. The uncertainty in estimating energy storage in the whole lake could be reduced by monitoring water temperature at a few centimeters depth to compute the energy flux into the air at the surface or into the water surface. The vertical energy balance at the surface of the lake water can be expressed by Eq. 8.5 dropping the advective energy term:

$$\lambda E = R_n - H - G \tag{8.5}$$

where λE is latent heat flux, H is sensible heat flux, and G is heat gained or lost by the upper layer of the lake. λ is latent heat of vaporization of water (Eq. 8.6):

$$\lambda = 2.501 - 0.002361 T_s \tag{8.6}$$

where T_s is water temperature in °C at lake surface and λ is in MJ kg^{-1}. Net solar radiation (R_n) is measured using hemispherical net radiometers or estimated from solar radiation measurements (Jensen 1974), as shown in Eq. 8.7:

$$R_n = (1 - \alpha) R_s - R_b \tag{8.7}$$

where α is shortwave reflectance or albedo, R_s is solar radiation, and R_b is net back or outgoing thermal radiation. If measured solar radiation data are not available, there are formulas to estimate it from clear day solar radiation or extraterrestrial solar radiation (Jensen 1974; Linacre 1993; Allen et al. 2005). Figure 8.4 depicts extraterrestrial radiation (R_a), clear sky or cloudless solar radiation (R_{so}), incoming solar radiation (R_s), and net shortwave radiation (R_n) at south Florida. The source of data for R_a at 26° latitude is FAO 1977. Cloudless solar radiation (R_{so}) was estimated as 71% of R_a. Solar radiation (R_s) was measured over Lake Okeechobee in south Florida, and net solar radiation (R_n) was measured on land close to the lake (latitude 26° 39′ and longitude 80° 37′), averaged from 2001 to 2009.

To directly apply the energy balance equation (Eq. 8.5), the estimation of sensible heat (H), heat gained or lost by air at the lake surface, is difficult. Temperature

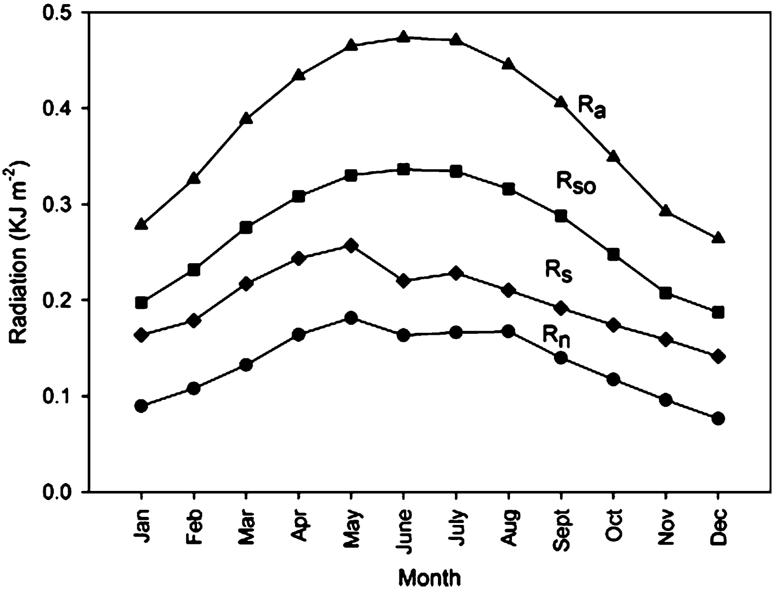

Fig. 8.4 Extraterrestrial (R_a), cloudless solar radiation (R_{so}), solar radiation (R_s), and net solar radiation (R_n) over south Florida

gradient and sensible heat transfer coefficient estimations are challenging. In modeling the evaporation and condensation processes, momentum, mass, and energy transfer mechanisms have to be accounted. The sensible heat (H) flux, shear stress (τ), and latent heat (λE) flux are presented in general form by Eqs. 8.8, 8.9, and 8.10:

$$H = \rho c_{\mathrm{p}} k_{\mathrm{h}} \frac{\mathrm{d}T}{\mathrm{d}z} \tag{8.8}$$

where ρ is air density, c_p is specific heat of air, k_h is sensible heat transfer coefficient, and $d\mathrm{T}/dz$ is change in temperature with height.

$$\tau = \rho k_{\mathrm{m}} \frac{\mathrm{d}u}{\mathrm{d}z} \tag{8.9}$$

where τ is shear stress, ρ is air density, k_m is transfer coefficient for shear stress, and $\mathrm{d}u/\mathrm{d}z$ represents the change in wind speed with height.

$$\lambda E = \frac{\lambda \varepsilon}{P} k_{\mathrm{w}} \frac{\mathrm{d}e}{\mathrm{d}z} \tag{8.10}$$

where λ is latent heat of vaporization, ε is the ratio of molecular weights of water to dry air, P is atmospheric pressure, k_w is coefficient for latent heat transfer, and $\mathrm{d}e/\mathrm{d}z$ is vapor pressure change with height.

The three transfer coefficients (k_m, k_w, k_h) are dependent on wind speed, vapor pressure, and temperature gradient with height. Surface conditions and atmospheric stability are also factors to be considered (Katul and Parlange 1992). For most applications, the three transfer coefficients are assumed to be equal (Federer 1970). The heat transfer coefficient (k_h) has been expressed in implicit and explicit forms. Explicit forms from various sources are presented by Eqs. 8.11, 8.12, and 8.13:

$$k_{\mathrm{h}} = u_*^2 \frac{\mathrm{d}z}{\mathrm{d}u} \tag{8.11}$$

where u_* is friction velocity and $\mathrm{d}z/\mathrm{d}u$ is the inverse of wind speed gradient (Monteith 1973).

$$k_{\mathrm{h}} = \frac{k u_* (z - d + z_{\mathrm{h}})}{\Phi_{\mathrm{h}}} \tag{8.12}$$

where k is the von Karman constant (0.41), z is height, d is displacement height, z_{h} is roughness length for heat transfer, and Φ_{h} is a stability correction factor, a function of the Monin–Obukhov length (Stannard 1993).

$$k_h = u_* \theta_* \frac{\mathrm{d}z}{\mathrm{d}T} \tag{8.13}$$

where θ_* is temperature scale and is computed by Eq. 8.14 as the inverse of the temperature gradient (Jacovides et al. 1992).

$$\theta_* = \frac{\Delta T k}{\ln\left(\frac{z_2}{z_1}\right)} \tag{8.14}$$

where ΔT is temperature difference between the two heights of measurement (z_1 and z_2); Federer (1970) presented Eq. 8.15.

$$k_h = \frac{k u_* z}{\Phi_h} \tag{8.15}$$

In this chapter, Lake Okeechobee in south Florida is selected to show the application of different evaporation estimation methods as meteorological data is available measured in the lake on platforms. Figure 8.5 depicts south Florida and Lake Okeechobee with weather monitoring sites. Site L006 was used in this analysis.

The platform with monitoring stations and close-up of the sensors is shown in Fig. 8.6a, b. Monitoring parameters in Lake Okeechobee at weather station L006 are shown in Table 8.2.

To demonstrate the application of the energy balance method, net solar radiation is borrowed from a land-based weather station in the vicinity of the lake (ENR308).

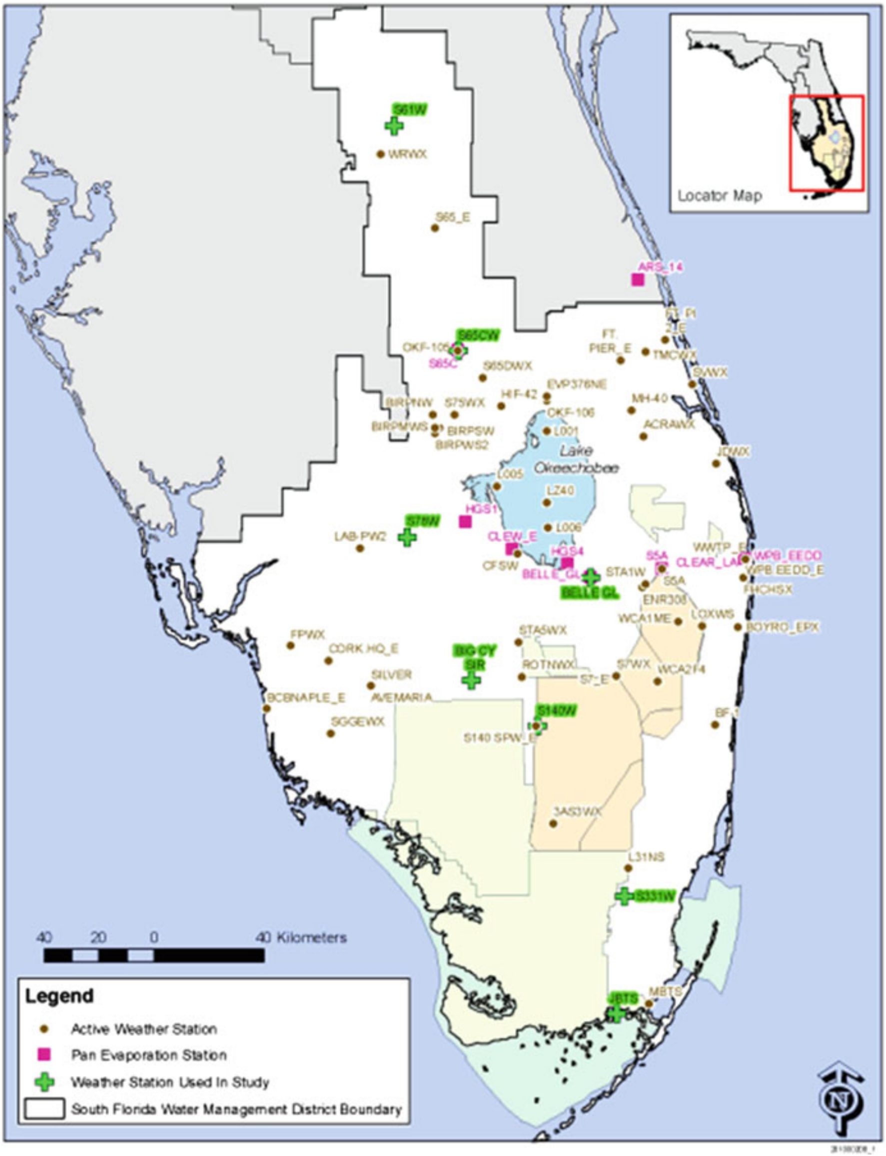

Fig. 8.5 South Florida and Lake Okeechobee with weather-monitoring sites (Abtew et al. 2011)

Equations 8.11 and 8.13 were applied for estimating the heat transfer coefficient using the meteorological parameter observations in Lake Okeechobee. Sensible heat is computed using Eq. 8.8. Two parameters to be estimated are the friction velocity (u_*) and the inverse of the wind speed gradient (dz/du). In order to estimate u_*, many more parameters have to be estimated. Table 8.3 depicts equations used to estimate parameters required to derive u_*, k_h, and H.

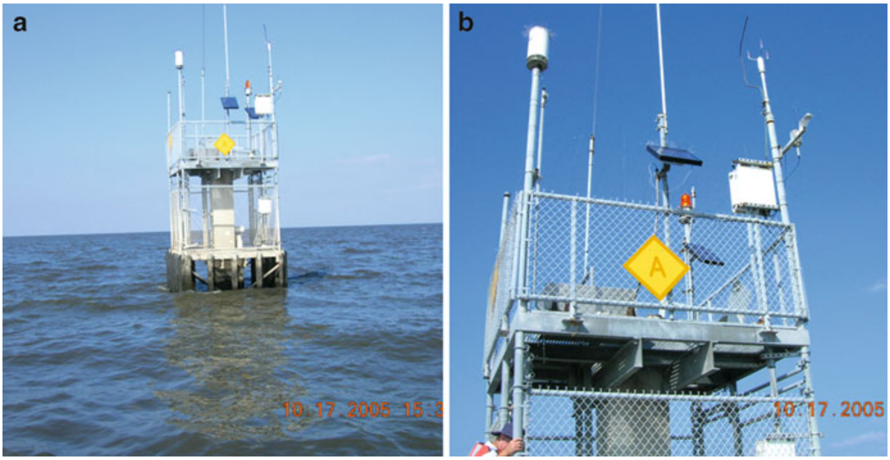

Fig. 8.6 (**a**) Weather station tower in Lake Okeechobee (Photograph provided by South Florida Water Management District), (**b**) Close-up of weather station tower sensors in Lake Okeechobee (Photograph provided by South Florida Water Management District)

Table 8.2 Monitoring parameters at station L006 in Lake Okeechobee (average water surface elevation is 4.7 m NGVD29)

Parameter	Unit	Height above water level	Frequency
Water temperature	°C	15 cm below water level	15 min
Water temperature	°C	1.5 m above lake bottom	15 min
Water temperature	°C	30 cm above lake bottom	15 min
Air temperature	°C	At 9.8 m − lake elevation	15 min
Humidity	%	At 9.8 m − lake elevation	15 min
Wind speed	mph	At 11.7 m − lake elevation	15 min
Wind direction	Degrees	At 11.7 m − lake elevation	15 min
Wind gust (maximum)	mph	At 11.7 m − lake elevation	10 s
Atmospheric pressure	kPa	At 9.8 m − lake elevation	15 min
Solar radiation	kw m^{-2}	At 9.8 m − lake elevation	15 min
Rain	Inches	At 11.7 m − lake elevation	15 min

The energy balance method was applied for 1 day, May 1, 1998, with 15 min of meteorological data measured inside the lake except net solar radiation (Table 8.2). Water heat flux (G in kJ m^{-2}) is computed by Eq. 8.16. Results of the analysis are presented in Table 8.4 as an hourly average from 6:00 a.m. to 7:00 p.m.:

$$G = c_s d_w (T_n - T_{n-1})^* \frac{1,000}{86,400} \tag{8.16}$$

where c_s is water heat capacity (4.18 MJ m^{-3} $°C^{-1}$) and d_w is water depth where the top 30-cm water depth was used for change in storage computation with water

Table 8.3 Supplementary parameters for energy balance evaporation estimation

Parameter	Equation	Remarks
u_*	$u_* = \frac{uk}{\ln\left((z-d)/z_o\right)}$	u = wind speed (m s^{-1}) at 2-m height
		k = von Karman constant (0.41)
		z = height of wind speed measurement (m)
		d = displacement height (m)
		z_o = roughness height (m)
d (Abtew et al. 1989)	$d = 0.5$ h	h = average wave height (m)
z_o (Abtew et al. 1989)	$Z_o = 0.13\ (h–d)$	–
h (Linsley and Franzini 1979)	$h = 0.005\ u^{1.06}\ F^{0.47}$	u = wind speed (km h^{-1}) at 6.8-m height
		F = fetch (km)
dz	Average of two heights of winds speed measurement	In m
du	Average of change in wind speed from water surface to 2 m and from 2 to 6.8 m	In m s^{-1}
dT	Change in temperature between water temperature at 15-cm depth and air temperature at 4.7 m	°C
θ (Federer 1970)	$\theta_* = \frac{\Delta Tk}{\ln(z_2/z_1)}$	$z_2 = 5.2$ m and $z_1 = 0.5$ m
		Height raised by 0.5 m to match k_h computed by Eq. 8.11 and to avoid dividing by zero or small height at the surface

temperature measured at 15-cm depth in the lake. T_n and T_{n-1} are water temperature on day $n-1$ and n. Advection energy is assumed negligible with the assumption that inflow and outflow temperature is the same as lake surface water temperature.

Based on the energy balance, lake evaporation for 24 h on May 1, 1998, was 4.04 mm. Energy balance components are shown for May 1, 1998, from 6:00 a.m. to 7:00 p.m. (Fig. 8.7a). The 15-min lake evaporation in mm is shown in Fig. 8.7b. A limitation of the energy balance method is that the ability of the environment to transfer or hold water vapor is not accounted. These limitations are shown by the vapor pressure deficit pattern over Lake Okeechobee from the 1-day (May 1, 1998) energy balance analysis (Fig. 8.7c).

8.2.4 Mass Transfer Method

Mass transfer models are based on estimating the net transport of water vapor from the lake surface to the atmosphere (Ikebuchi et al. 1988). The methods are based on Dalton's law where vapor transfer from an evaporating surface is proportional to

Table 8.4 Hourly average energy flux and other parameters on May 1, 1998

Time	u @ 6.61 mm s^{-1}	u @ 2 mm s^{-1}	Wave height m	ΔT °C	u_* m s^{-1}	k_h (Eq. 8.11) m^2 s^{-1}	R_n kJ m^{-2}	G kJ m^{-2}	H kJ m^{-2}	E kJ m^{-2}
6:00	6.93	4.91	0.856	0.895	0.603	0.248	0.004	0	0.058	−0.054
7:00	6.15	4.45	0.755	1.035	0.522	0.207	0.154	0	0.056	0.098
8:00	5.48	4.02	0.668	0.518	0.452	0.16	0.255	−0.002	0.024	0.232
9:00	3.48	2.67	0.412	0.089	0.26	0.09	0.441	0	0.002	0.439
10:00	4.82	3.58	0.583	−1.34	0.386	0.144	0.354	0	−0.051	0.404
11:00	4.11	3.11	0.493	−1.1	0.32	0.115	0.432	0.006	−0.034	0.461
12:00	2.05	1.62	0.236	−0.84	0.14	0.044	0.621	0.005	−0.008	0.624
13:00	1.42	1.15	0.16	−2.38	0.091	0.026	0.54	−0.009	−0.015	0.563
14:00	0.43	0.36	0.045	−4.685	0.023	0.006	0.413	−0.001	−0.007	0.421
15:00	2.97	2.26	0.352	−3.601	0.225	0.08	0.19	0.001	−0.075	0.264
16:00	7.37	5.15	0.916	−1.608	0.655	0.278	0.079	0.004	0.113	0.187
17:00	9	6.06	1.13	−0.825	0.838	0.373	0.013	−0.001	−0.077	0.09
18:00	11.68	7.26	1.49	1.143	1.169	0.576	−0.02	−0.001	0.163	0.252

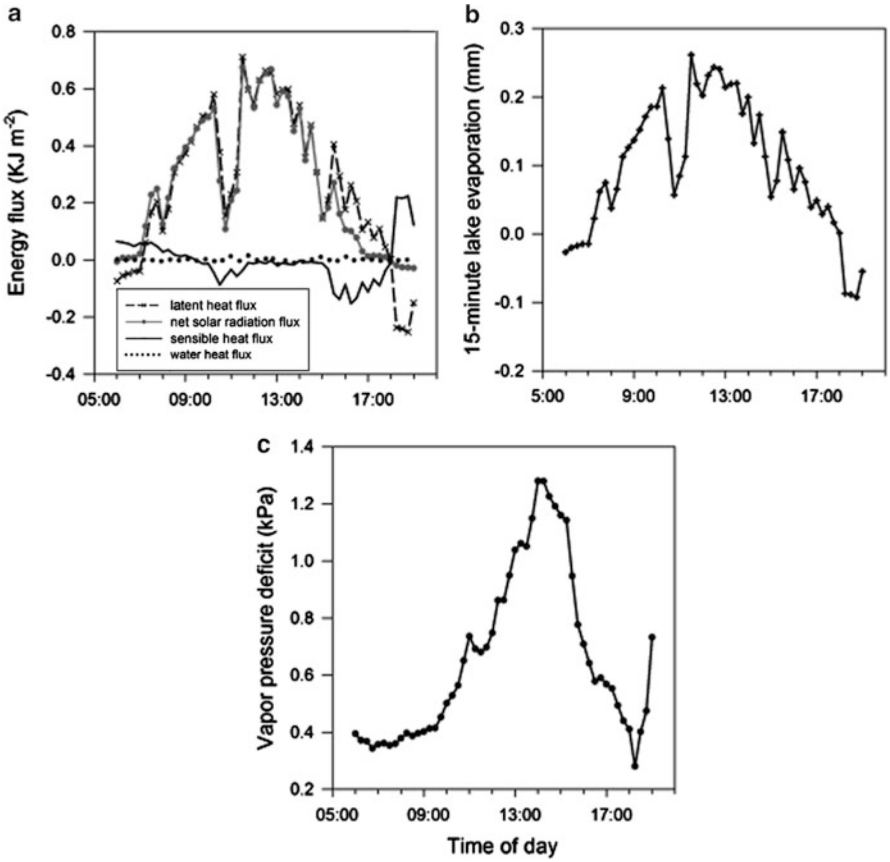

Fig. 8.7 (**a**) Energy balance components, (**b**) 15-min lake evaporation, and (**c**) vapor pressure deficit (May 1, 1998)

wind velocity and vapor pressure deficit over the surface. By combining the sheer stress and latent heat flux, the mass and momentum transfer equations produce a mass transfer equation (Eq. 8.17) for estimating evaporation (E) (Singh 1989):

$$E = \rho u_*^2 \frac{k_w(q_2 - q_1)}{k_m(u_2 - u_1)} \tag{8.17}$$

where $(q_2 - q_1)$ is difference in specific humidity (dimensionless) at heights z_2 and z_1 above the water surface and $(u_2 - u_1)$ is wind speed difference between the heights z_2 and z_1. A mass transfer model that was originally developed by Harbeck (1962) was applied by Hostetler and Bartlein (1990) to estimate evaporation for Harney–Malheur Lake in Oregon (Eq. 8.18).

$$E = N_o u_2 (e_o - e_a) \tag{8.18}$$

where E is in mm, N_o is an empirically determined mass transfer coefficient (mm s m^{-1} kPa^{-1}), u_2 is wind speed at 2-m height above the lake surface, e_o is saturation vapor pressure at the lake surface (kPa), and e_a is ambient vapor pressure of the air (kPa). The mass transfer coefficient N_o is computed for large lakes from lake surface area, A (km^2), by Eq. 8.19 (Shuttleworth 1993).

$$N_o = 2.909\ A^{-0.05} \tag{8.19}$$

This method was applied for the same day the energy balance method was applied for Lake Okeechobee (May 1, 1998). The lake surface area is 1,732 km^2. The computed mass transfer coefficient N_o is 2.00, and the average 2-m height wind speed above the lake surface was 4.18 m s^{-1}. Evaluating the performance of the mass transfer method for evaporation estimation in a semiarid region of India, the coefficient, N_o, was found to be 2.35 (Ali et al. 2007). The average vapor pressure difference calculated as described above is 0.453 kPa. Using the mass transfer method (Eq. 8.18), the estimated evaporation for Lake Okeechobee on May 1, 1998, is 3.79 mm. The limitation with the mass transfer method is that energy required for evaporation is not considered. Increase in wind speed and vapor pressure deficit result in extremely high evaporation. Both the energy balance and mass transfer methods were applied for the full month of May 1998 using daily average meteorological data. The monthly mean lake evaporation by the energy balance and mass transfer methods was 4.77 and 4.86 mm day^{-1}, respectively. When daily evaporation estimates are compared, the limitations of both methods are clearly shown. Figure 8.8a depicts wind speed at 2-m height and vapor pressure deficit over Lake Okeechobee for the month of May 1998. Figure 8.8b depicts daily lake evaporation estimates by the energy balance and the mass transfer methods.

8.2.5 The Penman Method

Penman in 1948 derived a combination equation to estimate evaporation. The method combines the energy required to cause evaporation and the mechanisms required to remove vapor from the evaporating surface (Jensen et al. 1990). Vapor pressure deficit combined with wind speed creates the condition for vapor movement from higher saturation to lower saturation zones and the resistance of vapor movement from water to air. The energy balance between net radiation, sensible heat flux, and change in heat storage results in energy available for evaporation. The Penman combination equation (Eq. 8.20) that is applied to compute potential evapotranspiration or open water evaporation is given as follows (Shuttleworth 1993; Valiantzas 2006):

$$\mathrm{ET} = \frac{1}{\lambda}\frac{\Delta(R_n - G) + \gamma 6.43(f(u))(e_a - e_d)}{(\Delta + \gamma)} \tag{8.20}$$

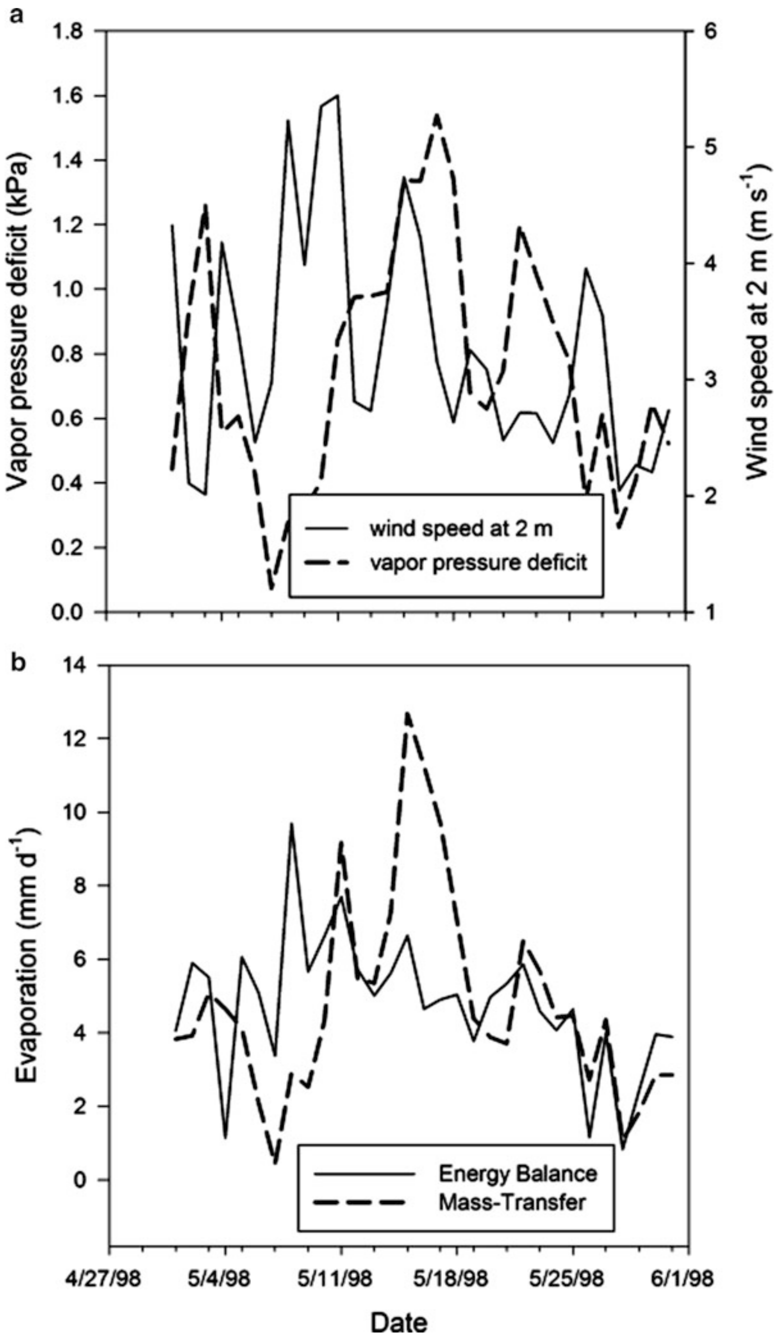

Fig. 8.8 (**a**) Wind speed at 2-m height and vapor pressure deficit, (**b**) daily evaporation estimates by energy balance and mass transfer methods

where E is evaporation in mm day^{-1}, R_n is net radiation (MJ m^2 day^{-1}), G is water heat flux (MJ m^2 day^{-1}), is slope of vapor pressure curve (kPa °C^{-1}), γ is psychrometric constant (kPa °C^{-1}), e_s is saturation vapor pressure, e_d is actual vapor pressure, $(e_s - e_d)$ is vapor pressure deficit computed using air temperature, and $f(u)$ is wind function expressed by Eq. 8.21:

$$f(u) = a_w + b_w u_2 \tag{8.21}$$

where a_w and b_w are wind function coefficients and u_2 is wind speed at 2-m height (m s^{-1}). In the original Penman equation, $a_w = 1$ and $b_w = 0.536$ (Valiantzas 2006). Realizing that these coefficients are site dependent, J.W. Wright, USDA, Kimberly, Idaho, developed equations to estimate wind coefficients based on a normal probability density function to reflect seasonal variation with the general form shown by Eqs. 8.22 and 8.23 for northern latitudes (Allen et al. 1989; Shuttleworth 1993):

$$a_w = c_1 + c_2 \exp\left\{-\left[\frac{J - 173}{58}\right]^2\right\} \tag{8.22}$$

$$b_w = c_3 + c_4 \exp\left\{-\left[\frac{J - 243}{80}\right]^2\right\} \tag{8.23}$$

where c_1, c_2, c_3, and c_4 are site-dependent coefficients. For Kimberly, Idaho, c_1, c_2, c_3, and c_4 are 0.4, 1.4, 0.605, and 0.345, respectively. Coefficients of 0.1, 3.0, 0.04, and 0.2 were developed for south Florida based on least square fit of lysimeter measured evaporation and model estimated data (Abtew 1996). The Penman method was also applied to estimate lake evaporation for May 1998. The average daily evaporation was 4.47 mm day^{-1}.

8.2.6 The Simple Abtew Method

Comparison of lysimeter measurements of open water evaporation and wetland vegetation evapotranspiration showed that there is no significant difference between evaporation from shallow lakes, evapotranspiration from wetland vegetation, and potential evapotranspiration in south Florida. It was also shown that most of the variance (73%) in evaporation and evapotranspiration in south Florida is explained by variation in solar radiation. A simple equation was calibrated to estimate daily lake evaporation, wetland evapotranspiration, or potential evapotranspiration in south Florida (Abtew 1996). Equation 8.24 is also cited as the Abtew equation and simple Abtew equation in published literature (Abtew 1996; Xu and Singh 2000; Abtew et al. 2003; Delclaux and Coudrain 2005; Oudin et al. 2005; Shoemaker and Sumner 2006; Melesse et al. 2009; Zhai et al. 2009; Enku et al. 2011; Setegn et al. 2011). The advantage of the simple equation is that it uses only a single input parameter, R_s:

$$\mathrm{ET} = K_1 \frac{R_s}{\lambda} \tag{8.24}$$

where ET is daily evapotranspiration from wetland or shallow open water or potential evapotranspiration (mm day^{-1}), R_s is solar radiation (MJ m^{-2} day^{-1}), λ is latent heat of vaporization (MJ kg^{-1}), and K_1 is a dimensionless coefficient (0.53). The mm day^{-1} unit is derived from the fact that a kilogram of water is 1,000 cc (10^6 mm^3) and a square meter is 10^6 mm^2. The simple Abtew method was also applied to estimate lake evaporation for May 1998. The average daily lake evaporation estimate was 4.68 mm day^{-1}. The simple Abtew method is currently applied to acquire daily evaporation data for Lake Okeechobee in south Florida and stored in a widely accessed hydrometeorological database, DBHYDRO. It was also successfully applied to estimate evaporation from Lake Ziway in the Ethiopian Rift Valley (Melesse et al. 2009). Compared to eight evaporation models, the Simple Method provided the best result in estimating evaporation from Lake Titicaca (Delclaux and Coudrain 2005). Lake Titicaca is located in the Southern Hemisphere, between Bolivia and Peru at 3,810-m elevation (16°S 69°W).

8.2.7 Solar Radiation–Maximum Temperature Method

In order to include air temperature in evaporation estimation, an equation was developed and calibrated using lysimeter measurements (Abtew 1996). Maximum air temperature in °C (T_{max}) is added to Eq. 8.24 with a calibration coefficient, K_3 (°C), Eq. 8.25. This method was also applied to estimate lake evaporation for May 1998. The average daily lake evaporation estimate was 4.63 mm day^{-1}:

$$\mathrm{ET} = \frac{1}{k_3} \frac{R_s}{\lambda} T_{max} \tag{8.25}$$

8.2.8 Modified Turc Equation

The original Turc equation for potential evapotranspiration estimation for humid regions is similar to Eq. 8.26 but uses daily average temperature. The Turc equation was modified by using daily maximum air temperature (°C) as this gave better fit to measured lysimeter data in south Florida, a humid and warm subtropical region (Abtew 1996, 2001). The coefficient K_2 is similar to the original Turc value of 0.013, which was originally recommended for humid region:

$$\mathrm{ET_P} = K_2 \frac{(23.89 R_\mathrm{s} + 50) T_\mathrm{max}}{(T_\mathrm{max} + 15)} \tag{8.26}$$

where E is evaporation in mm and R_s is solar radiation in MJ m^{-2} day^{-1}. K_2 has unit mm MJ^{-1} m^2 day in order to balance the units. The modified Turc equation was also applied to estimate lake evaporation for May 1998. The average daily lake evaporation estimate was 4.87 mm day^{-1}.

8.2.9 Priestley–Taylor Method

The Priestley–Taylor equation is a simplified form of the Penman equation, where the aerodynamic component is left out of the equation. A coefficient (α), with a value of greater than 1 (1.26), is included to make up for the loss of the aerodynamic component of evaporation. This method is widely used to estimate evaporation sometimes with modified coefficient (α) (Reis and Dias 1998; Abtew 1996). The Priestley–Taylor equation is presented by Eq. 8.27:

$$\mathrm{ET} = \frac{\alpha}{\lambda} \frac{\Delta R_\mathrm{n}}{(\Delta + \gamma)} (R_\mathrm{n} - G) \tag{8.27}$$

The Priestley–Taylor equation was also applied to estimate lake evaporation for May 1998 with α value of 1.26. The average daily lake evaporation estimation was 4.26 mm day^{-1}. The average daily lake evaporations estimated by the Penman, simple Abtew, solar radiation–maximum temperature, modified Turc, and Priestley–Taylor methods are 4.47, 4.68, 4.63, 4.87, and 4.26 mm day^{-1}, respectively. Figure 8.9 depicts daily evapotranspiration measurement by the Penman, simple Abtew, solar radiation–maximum temperature, modified Turc, and Priestley–Taylor methods.

8.2.10 Energy Balance–Bowen Ratio Method (EBBR)

Estimation of sensible heat, H, in the energy balance method is challenging as shown in Sect. 8.2.3. As shown in Eq. 8.8, temperature change with height and a transfer coefficient is required to estimate H. The EBBR method substitutes the Bowen ratio in the energy equation in place of H (Eq. 8.28). The Bowen ratio (β) is the ratio of sensible heat to latent heat flux (Eq. 8.29):

$$\lambda E = \frac{R_\mathrm{n} - G}{1 + \beta} \tag{8.28}$$

$$\beta = \frac{H}{\lambda E} = \gamma \frac{\Delta T}{\Delta e} \tag{8.29}$$

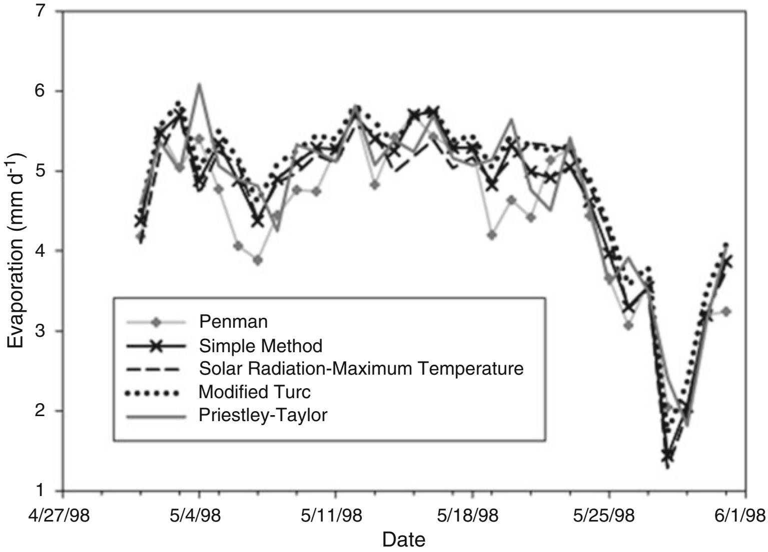

Fig. 8.9 Daily lake evaporation estimation by the Penman, simple Abtew, solar radiation–maximum temperature, modified Turc, and Priestley–Taylor methods

The Bowen ratio estimation requires temperature and vapor pressure measurements at two heights over the water surface. Different approaches have been presented to avoid measurements at two heights: replacing with water and air temperature measurements and associated saturation vapor, actual vapor pressure, and air pressure. Referring to studies at Lake Mead and Lake Eucumbene, Omar and El-Bakry (1981) applied a different format (Eq. 8.30) in their estimation of evaporation from Lake Nasser, Aswan Dam. Stannard and Rosenberry (1991) credited the Bowen ratio equation to E.R. Anderson and Lake Hefner, Oklahoma, evaporation study. Both formats use a constant and air pressure in place of γ. The Bowen ratio estimation equation (Eq. 8.30) with analysis is presented by Reis and Dias (1998):

$$\beta = \gamma \frac{(T_s - T_a)}{(e_s - e_d)} \tag{8.30}$$

where T_s is lake surface water temperature (°C), T_a is air temperature over the lake (°C), e_s is saturation vapor pressure corresponding to T_s (kPa), and e_d is the air actual vapor pressure corresponding to T_a (kPA). Lake Okeechobee temperature, pressure, and vapor pressure data was used to compute daily evaporation for May 1998. The average daily evaporation with the EBBR method (Eqs. 8.28 and 8.30) is 4.47 mm. Daily evaporation estimates are shown in Fig. 8.10 comparing EBBR, Penman, and the simple Abtew methods. Except at the beginning of the month, the EBBR method has given comparable estimates to the Penman, simple Abtew, and other methods.

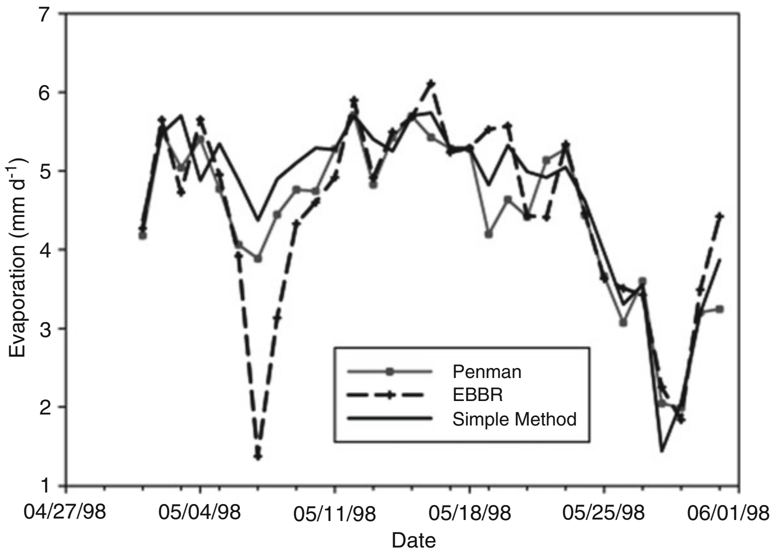

Fig. 8.10 Daily lake evaporation estimation by the Penman, EBBR, and simple Abtew methods

The EBBR method fails to produce reasonable estimates of evaporation when there is a sudden drop in air temperature compared to water temperature or vice versa.

8.3 Summary

Lake evaporation estimation or measurement has been challenging and various methods have been applied for different lakes. Method selection should be dependent on location or environment of the lake and available input data. For lakes located in arid areas, advection energy needs to be accounted for. For tropical and subtropical lakes, the most dominant parameter, solar radiation, may be enough to estimate evaporation. Adding temperature average or maximum into the equation could improve the estimate. The mass transfer method has the potential to produce out of range estimates due to high influence of wind speed on the equation. Relatively, energy balance methods produce within range values, as the available energy limits the maximum evaporation that could occur. The energy balance–Bowen ratio (EBBR) method is susceptible to error of estimation when daily difference in lake surface water and air temperatures is large. The simple Abtew method and the solar radiation–maximum temperature methods use minimum input and produce competitive results for south Florida and other locations where applied. In tropical and subtropical areas, such methods can be tested and successfully applied. Methods that do not use net radiation bypass the technical challenges of acquiring good quality net radiation data for longer time periods. To compare the

Table 8.5 Mean daily evaporation, standard deviation, and range for Lake Okeechobee for the month of May 1998

Method	Mean (mm)	Stdev (mm)	Range
Energy balance	4.77	1	0.83–9.69
Mass transfer	4.86	2.81	0.4–12.68
Penman	4.47	1	1.99–5.71
Simple Abtew method	4.68	1.05	1.44–5.74
Solar radiation–maximum temperature method	4.63	1.04	1.28–5.70
Modified Turc equation	4.87	0.99	1.73–5.87
Priestley–Taylor equation	4.26	0.88	1.65–5.41
Energy balance–Bowen ratio method	4.47	1.19	1.37–6.11

difference between evapotranspiration methods, good quality shorter period data should be preferred than lower quality but longer period data. Otherwise, error in data will bias the results. Table 8.5 summarizes evaporation estimation for Lake Okeechobee in south Florida for the month of May 1998. The energy balance and mass transfer methods have wider ranges and out of range values reflecting the inherent deficiencies of the two methods.

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