NOTES AND CORRESPONDENCE

Quantifying Surface Energy Fluxes and Evaporation over a Significant Expanding Endorheic Lake in the Central Tibetan Plateau

Yanhong GUO
Key Laboratory of Tibetan Plateau Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, China
University of Chinese Academy of Sciences, China

Yinsheng ZHANG
Key Laboratory of Tibetan Plateau Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, China
CAS Center for Excellence in Tibetan Plateau Earth Sciences, China

Ning MA
Key Laboratory of Tibetan Plateau Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, China
University of Chinese Academy of Sciences, China

Hongtao SONG, and Haifeng GAO
Key Laboratory of Tibetan Plateau Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, China

(Manuscript received 17 March 2016, in final form 23 June 2016)

Abstract

In this study, energy and water vapor exchange between the lake and atmosphere over the largest lake in Tibet, Lake Serling Co, was measured by an eddy covariance system from April 26 until September 26, 2014. The results demonstrated that the diurnal variations of the sensible heat flux (H) and latent heat flux (LE) were different from that of the net radiation (Rn). Rn reached its peak value at the local noon, whereas H peaked in the morning and LE peaked in the afternoon. On a seasonal scale, H and LE were also different from Rn. The maximum value of Rn occurred in June, while the maxima of H and LE were observed in September. Lake evaporation was quantified with a daily mean value of 2.7 mm d⁻¹ and a total amount of 417.0 mm during the study period. In addition, evaporation from Lake Serling Co was compared with two types of pan evaporation (D20 pan and E601B pan) and potential evaporation on the land surface. The variability of conversion coefficients between lake evaporation, D20/E601B pan and potential evaporation indicate that coefficients varied depending on the month and could not be defined as a single experimental value.

Keywords energy fluxes; eddy covariance method; lake evaporation; pan evaporation; Tibetan Plateau
1. Introduction

With a mean altitude of 4,000 m above sea level (a.s.l.), the Tibetan Plateau (TP) plays important roles in the global water cycle and the Asian monsoon system because of its thermodynamic influence on atmospheric circulation patterns (Ma et al. 2015a; Wu et al. 2012; Xu and Haginoya 2001). The TP has the greatest number (1055) and area (41831.7 km²) of lakes in China (Ma et al. 2011). Heat and water exchange between such large water bodies and the atmosphere strongly influence the local climate (Lv et al. 2007; Yang et al. 2015). Observations of lake surface energy balance and evaporation by the eddy covariance (EC) method, which is typically considered to be the most reliable and accurate method, have been reported in several studies (Allen and Tasumi 2005; Liu et al. 2014; Spank and Bernhofer 2008; Tanny et al. 2008). Previous observed results have shown that the characteristics of the surface energy balance and turbulent exchange vary and depend on various factors, including lake sizes, depths, and regional climate conditions. For instance, shallow subtropical lakes, e.g., Lake Taihu (Wang et al. 2014) and Logan Dam (McGloin et al. 2015), have small heat capacities and quickly respond to changes in the atmospheric forcing, leading to short time lags between net radiation and turbulent fluxes. In contrast, deep cold lakes, e.g., Great Slave Lake (Oswald and Rouse 2004; Rouse et al. 2003) and Superior Lake (Blanken et al. 2011), have larger heat capacities and slowly respond to changes in the atmospheric forcing, leading to longer time lags (e.g., several months). However, the characteristics of the heat and water exchange over the lakes in the TP remain unclear since the harsh environment of the TP has resulted in extremely sparse in situ observation of its lakes. Li et al. (2015) analyzed two-year surface energy budgets during the ice-free period over Lake Ngoring in the northeast of the TP. Wang et al. (2015) observed and modeled the interactions between a lake and the air during the summer over a small and shallow lake in the Nam Co basin of the central TP. Moreover, remote-sensing-based investigations have suggested that most lakes in the central TP have experienced significant expansions in area and increases in water level during the past few decades (Lei et al. 2013; Wang et al. 2013; Zhang et al. 2011, 2013). This is especially true for the largest lake in Tibet, Lake Serling Co (88°33′–89°21′E, 31°34′–31°57′N) (Fig. 1), which has expanded from 1617 km² in the mid-1970s to 2341 km² in the 2010s (Zhang et al. 2011). To reveal the reason for such fast endorheic lake expansion, an accurate estimation of the lake water balance components, including precipitation, runoff, and evaporation from the lake surface is needed (Lenters et al. 2005). As Lake Serling Co is an endorheic lake, evaporation from its surface is the only output component because there is no outlet runoff. Therefore, precise quantification of the evaporation from Lake Serling Co is crucial for elucidating the mechanism of lake expansion.

Based on in situ observations of turbulent fluxes measured using the EC method from April 26 until September 26, 2014, the objectives of the present study were (1) to investigate the characteristics of the energy fluxes over largest endorheic lake in the central TP, Lake Serling Co, and (2) to quantify the evaporation over the surface of Lake Serling Co and its relation with two types of pan evaporation as well potential evaporation.

2. Methods

2.1 Site description and measurements

The elevation of Lake Serling Co is about 4543 m (a.s.l.), and its maximum depth is about 50 m (Chen et al. 2001; Lei et al. 2013). Our field observations showed that Lake Serling Co is ice-covered from the end of December until mid-April of the following year. A flux tower was installed on an island in the west part of Lake Serling Co (88°37′, 31°48′, Fig. 1b, c). An EC system consisting of a three-dimensional sonic anemometer (CSAT3, Campbell Scientific Inc., USA) and an open-path carbon dioxide/water vapor (CO₂/H₂O) infrared gas analyzer (Li 7500, Li-Cor USA) and an open-path carbon dioxide/water vapor (CO₂/H₂O) infrared gas analyzer (Li 7500, Li-Cor Inc., USA) was mounted on the tower at a height of approximately 3.2 m above the water surface. Raw data were recorded by a data logger (CR1000, Campbell Scientific Inc., USA) at a frequency of 10 Hz. Additionally, a four-component net radiometer (CNR4, Kipp & Zonen, Netherlands) was installed on the tower approximately 1.5 m above the water surface. The air temperature (Ta) and relative humidity (RH) were measured using temperature and humidity probes (HMP45C, Vaisala Inc., Finland) located 3.0 m above the water surface. The water temperature was also measured using temperature sensors (109, Campbell Scientific Inc., USA) near the tower at the depths of 0.25 m, 0.45 m, and 0.85 m, respectively. A double tipping-bucket rain gauge (SM3-1, Shanghai Meteorological Instrument Factory Co. Ltd., China) was installed near the tower. A China D20 pan with a diameter of 0.2 m was installed at the shore side (about 3 km from the lake side) to observe the pan
evaporation (Fig. 1b). In addition, the lake surface temperature ($T_w$) was obtained by using the water temperature at a depth of 0.45 m, which is reasonable because the water is well mixed by high-speed wind (Nordbo et al. 2011; Zhang and Liu 2013). The water surface vapor pressure ($e_w$) was calculated according to $T_w$ in this study.

2.2 Data processing and footprint analysis

The 10 Hz raw EC data were processed using EddyPro software (Burba 2013). All of the necessary data corrections, including the deletion of outliers, the double rotation of the coordinate axes, and the correction of the Webb–Pearman–Leuning density were performed (Kaimal and Finnigan 1994; Vickers and Mahrt 1997; Webb et al. 1980). Finally, the 30 min mean values of the sensible heat flux ($H$) and latent heat flux ($LE$) were calculated.

To perform turbulent flux measurements, it is necessary to quantify the sampling area, also called the flux footprint, which is the upwind area or distance contributing to the turbulent fluxes (Blanken et al. 2000). Here we estimated the upwind distance using a footprint model developed by Kljun et al. (2004). We also analyzed the wind rose during the study period, and the results showed that the vast majority of the winds flowed from the southwest and northeast (Fig. 2b). The distances from the observation tower providing the 90 % cumulative contributions to the turbulent fluxes from the southwest and northeast were approximately 225 m and 254 m, respectively (Fig. 2c). Based on the Google Earth satellite image (Fig. 2a), we found that more than 90 % of the flux measurements originated from the lake surface. The 30 min mean $H$ and $LE$ data were filtered out during (1) periods in which the wind direction was 0°–25° or 270°–360°, which indicated fluxes that were affected by the island, and (2) rainfall events. These resulted in 18.9 % and 23.7 % gaps for $H$ and $LE$, respectively, during the study period. All of the gaps were filled using the bulk transfer relation, which has been widely used in a variety of previous studies (e.g., McGloin et al. 2015; Wang et al. 2014).
3. Results

3.1 Seasonal and diurnal variation of lake surface energy fluxes

The seasonal variations of the surface energy budget components and the meteorological variables over the lake surface are illustrated in Fig. 3 and Table 1. It can be seen that $R_n$ peaks in June and decreases from July with a fluctuating pattern due to the increases in cloudiness and precipitation after the monsoon onset (Figs. 3a, f). The seasonal variations of $LE$ are different from those of $R_n$ with some weekly scale sporadic increases (Fig. 3b), which are in agreement with the wind speed variations (Fig. 3f). The larger vapor pressure deficit between the lake surface and the overlying atmosphere caused by the decreased water vapor pressure in the atmosphere in high wind conditions might also enhance $LE$ (Figs. 3f, h). The seasonal variation of $H$ is also different from that of $R_n$, with an increasing trend during the observation.
Fig. 3. Daily means of meteorological variables and surface energy budget components: (a) net radiation ($R_n$); (b) latent heat flux ($LE$); (c) sensible heat flux ($H$); (d) Bowen ratio ($Bo$); (e) lake evaporation ($E_{Lake}$); (f) wind speed ($u$) and precipitation; (g) air temperature ($T_a$), lake surface temperature ($T_w$), and vertical temperature difference between the water surface and overlying atmosphere ($\Delta T$); (h) vapor pressure in the atmosphere ($e_a$), vapor pressure at the lake surface ($e_w$), and vertical vapor pressure difference between the water surface and overlying atmosphere ($\Delta e$); (i) product of wind speed and temperature difference between the water surface and air ($u'\Delta T$); and (j) product of wind speed and difference in vapor pressure between the water surface and overlying air ($u'\Delta e$).
Table 1. Monthly means of the components of the surface energy budgets and meteorological variables in 2014 over Lake Serling Co.

<table>
<thead>
<tr>
<th>Date</th>
<th>$R_n$ W m$^{-2}$</th>
<th>$H$ W m$^{-2}$</th>
<th>$LE$ W m$^{-2}$</th>
<th>$S$ W m$^{-2}$</th>
<th>$u$ m s$^{-1}$</th>
<th>$T_a$ °C</th>
<th>$T_w$ °C</th>
<th>$\Delta T$ °C</th>
<th>$e_a$ hPa</th>
<th>$e_w$ hPa</th>
<th>$\Delta e$ hPa</th>
<th>RH</th>
<th>Bo</th>
</tr>
</thead>
<tbody>
<tr>
<td>May 2014</td>
<td>210.2</td>
<td>-4.4</td>
<td>82.7</td>
<td>131.8</td>
<td>3.9</td>
<td>5.3</td>
<td>4.3</td>
<td>-1.0</td>
<td>3.8</td>
<td>8.3</td>
<td>4.5</td>
<td>0.45</td>
<td>-0.05</td>
</tr>
<tr>
<td>Jun 2014</td>
<td>234.7</td>
<td>-1.8</td>
<td>71.9</td>
<td>164.6</td>
<td>3.0</td>
<td>10.2</td>
<td>8.9</td>
<td>-1.3</td>
<td>6.2</td>
<td>14.0</td>
<td>5.2</td>
<td>0.52</td>
<td>-0.02</td>
</tr>
<tr>
<td>Jul 2014</td>
<td>188.0</td>
<td>13.6</td>
<td>72.2</td>
<td>102.1</td>
<td>3.4</td>
<td>12.1</td>
<td>13.0</td>
<td>0.9</td>
<td>9.1</td>
<td>14.9</td>
<td>5.8</td>
<td>0.68</td>
<td>0.19</td>
</tr>
<tr>
<td>Aug 2014</td>
<td>153.0</td>
<td>24.3</td>
<td>71.0</td>
<td>57.6</td>
<td>2.9</td>
<td>10.7</td>
<td>13.8</td>
<td>3.1</td>
<td>8.6</td>
<td>15.8</td>
<td>7.2</td>
<td>0.71</td>
<td>0.34</td>
</tr>
<tr>
<td>*Sep 2014</td>
<td>136.1</td>
<td>25.4</td>
<td>96.1</td>
<td>14.6</td>
<td>3.4</td>
<td>9.7</td>
<td>13.0</td>
<td>3.3</td>
<td>7.3</td>
<td>15.0</td>
<td>7.7</td>
<td>0.62</td>
<td>0.26</td>
</tr>
<tr>
<td>Mean</td>
<td>184.4</td>
<td>11.4</td>
<td>78.8</td>
<td>94.2</td>
<td>3.3</td>
<td>9.6</td>
<td>10.6</td>
<td>1.0</td>
<td>7.0</td>
<td>13.1</td>
<td>6.1</td>
<td>0.60</td>
<td>0.14</td>
</tr>
</tbody>
</table>

*Sep 2014 (from 1 Sep to 26 Sep), $R_n$: net radiation, $H$: sensible heat flux, $LE$: latent heat flux, $S$: energy balance residual ($S = R_n - H - LE$), $u$: wind speed, $T_a$: air temperature, $T_w$: water temperature at the depth of 0.45 m, $\Delta T$: vertical temperature difference between the water surface and the overlying atmosphere, $e_a$: actual vapor pressure in the atmosphere, $e_w$: vapor pressure at the evaporating surface calculated using $T_w$, $\Delta e$: vertical vapor pressure difference between the water surface and the overlying atmosphere, RH: relative humidity, Bo = $(H/LE)$: Bowen ratio.

period due to the increase in the vertical temperature gradient ($T_w - T_a$) (Figs. 3c, g). Due to the different specific heat capacities of water and air, the warming rate of the air over the lake surface is faster than that of the water in May and June, and the cooling rate of the air over the lake surface is also faster than that of water in August and September. Therefore, $H$ is negative in May and June and positive in August and September. $H$ shows better accordance with ($T_w - T_a$) than with $u$, while $LE$ shows better accordance with $u$ than with ($e_w - e_a$) (Figs. 3b, c and Figs. 3f, h). As demonstrated by Figs. 3b, c and Figs. 3i, j, $H$ shows the best accordance with $u'(T_w - T_a)$, while $LE$ shows the best accordance with $u'(e_w - e_a)$. The Bowen ratio (Bo = $H/LE$) ranges from -0.38 to 0.67, with a daily mean value of 0.14 during the observation period (Fig. 3d), demonstrating that the majority of the available energy contributed to water loss through evaporation.

Figure 4 shows the diurnal cycles of $R_n$, $u$, $T_a$, $e_a$, $T_w$, $e_w$, ($T_w - T_a$), ($e_w - e_a$), $H$, and $LE$ using the averages of the half-hourly values in May–September 2014. $R_n$ shows an ideal diurnal variation pattern with a peak at local noon (14:00) (Fig. 4a). However, $H$ reaches its maximum in the morning (08:00–10:00) and reaches its minimum in the early evening (18:00–20:00) (Fig. 4k). The diurnal variation of $LE$ is opposite to that of $H$. Specifically, the daily maximum values of $LE$ are observable in the afternoon (16:00–20:00), whereas the minimum occurs in the early morning (08:00–10:00) (Fig. 4i). It is apparent that the diurnal courses of both $H$ and $LE$ are different from that of $R_n$. The time at which $H$ peaks is earlier than at which $R_n$ peaks by about 4–6 h, whereas the time at which $LE$ peaks is later than that at which $R_n$ peaks by about 2–6 h. It is clear that the diurnal course of $H$ is mainly controlled by the diurnal course of $T_w - T_a$ (Figs. 4g, k). $T_w - T_a$ peaks in the morning and reaches its minimum in the afternoon due to the different diurnal courses of $T_w$ and $T_a$, which are caused by the different specific heat capacities of water and air (Figs. 4c, e). Similarly, the diurnal course of $H$ shows better accordance with ($T_w - T_a$) than with $u$, and $H$ shows the best accordance with $u'(T_w - T_a)$ (Figs. 4b, g, i, k). Meanwhile, the diurnal course of $LE$ shows better accordance with $u$ than with ($e_w - e_a$), and $LE$ shows the best accordance with $u'(e_w - e_a)$ (Figs. 4b, h, j, l). Moreover, the month-to-month differences of $H$ and $LE$ are mainly caused by the month-to-month differences of $T_w - T_a$ and $u$.

3.2 Comparison of lake evaporation with pan and potential evaporation over land

The daily amount of evaporation from Lake Serling Co ($E_{Lake}$) is shown in Fig. 3e. The daily values of $E_{Lake}$ range from 1.1 mm d$^{-1}$ to 5.6 mm d$^{-1}$, with a mean of 2.7 mm d$^{-1}$ and a total value of 417.0 mm between April 26 and September 26, 2014.

Pan evaporation and potential evaporation have usually been employed to estimate the lake evaporation on the TP (Li et al. 2001; Shi et al. 2010; Yang et al. 2010). For instance, an empirical pan coefficient (the ratio of the lake evaporation to the pan evaporation) of 0.614 based on the D20 pan has been used to estimate the evaporation over Qinghai Lake (Qu 1994). The potential evaporation calculated with the meteorological data from the adjacent meteorological station has also been used to estimate lake evap-
Fig. 4. Monthly mean diurnal course of (a) net radiation ($R_n$), (b) wind speed ($u$), (c) air temperature ($T_a$), (d) vapor pressure in the atmosphere ($e_a$), (e) lake surface temperature ($T_w$), (f) vapor pressure at the lake surface ($e_w$), (g) vertical temperature difference between the water surface and overlying atmosphere ($\Delta T$), (h) vertical vapor pressure difference between the water surface and overlying atmosphere ($\Delta e$), (i) product of wind speed and temperature difference between the water surface and air ($u\Delta T$), (j) product of wind speed and difference in vapor pressure between the water surface and overlying air ($u\Delta e$), (k) sensible heat flux ($H$), and (l) latent heat flux ($LE$) in May–September 2014.
oration on the TP (Li et al. 2001; Yu et al. 2011). It is therefore necessary to compare the current lake evaporation observations with the pan evaporation and potential evaporation. Figure 5a shows a comparison of the monthly values of \( E_{Lake} \) with the pan evaporation from two kinds of pan and potential evaporation. The first pan is the China D20 pan, which was mentioned above, and the other pan is the China E601B pan made of fiberglass with a depth of 0.687 m and a diameter of 0.618 m (Ma et al. 2015b), which was installed at the Xainza China Meteorological Administration meteorological station (Fig. 1a). The potential evaporation was calculated using the method given by Xu et al. (2005) based on the meteorological data from Xainza station. \( E_{Lake} \) exhibits seasonal patterns different from those of the evaporation from the D20 pan (\( E_{D20} \)), E601B pan (\( E_{E601B} \)), and potential evaporation (\( E_{p(Xainza)} \)). \( E_{D20} \), \( E_{E601B} \), and \( E_{p(Xainza)} \) reach peak values in June, but seasonal variations of \( E_{Lake} \) are not obvious. It can also be seen that \( E_{D20} \), \( E_{E601B} \), and \( E_{p(Xainza)} \) display similar seasonal variations during the observation period, but their amounts are obviously different. \( E_{D20} \) shows the largest value, followed by \( E_{p(Xainza)} \) and \( E_{E601B} \). Additionally, \( E_{D20} \), \( E_{E601B} \), and \( E_{p(Xainza)} \) are always larger than \( E_{Lake} \) during this period. The differences between the lake evaporation and pan evaporation are mainly caused by different thermal characteristics of the pan and large water bodies. The heat storage in the D20 and E601B pans is too small to cause any lag effect. Moreover, the horizontal wind from the dry surfaces around the pans could enhance the pan evaporation (Ma et al. 2015b; Shuttleworth and Maidment 1992). Conversely, a lake with a larger water body and depth could store and release a great amount of heat. Furthermore, thermal stratification occurs in large and deep water bodies and may accentuate the time lag between the net radiation and the lake evaporation (Finch and Calver 2008). Therefore, the seasonal patterns of the pan evaporation closely follow the energy input, but the seasonal patterns of the lake evaporation lag behind the energy input by several months (Brutsaert 1982, 2005). It should be noted that potential evaporation calculated using the meteorological variables from the land surface (i.e., Xainza station) actually reflected a non-potential environment maximum evaporation (Brutsaert 2005). For instance, in the potential evaporation results calculated using the observed data obtained from Lake Serling Co (\( E_{p(Lake)} \)), it can be found that \( E_{p(Lake)} \) is smaller than \( E_{p(Xainza)} \) and that \( E_{p(Lake)} \) agrees well with \( E_{Lake} \) (Fig. 5c). It is also evident that the larger wind speed and drier air in Xainza station are the main reasons for \( E_{p(Xainza)} \) being larger than \( E_{p(Lake)} \) and \( E_{Lake} \) (Figs. 5d, f). Meanwhile, it can also be concluded that lake evaporation can be well estimated using the potential method if meteorological variables near the lake (i.e., the meteorological variables are able to reflect the lake environment to a large extent) are available.

Research related to the ratio of lake evaporation to pan evaporation using a US Class A pan has been reported in many studies (Lapworth 1965; Boyd 1985). However, research into the ratio of lake evaporation to pan evaporation is scarce because of the lack of lake evaporation observation (Ma et al. 2016). The monthly ratios of lake evaporation to D20 pan evaporation (\( E_{Lake}/E_{D20} \)) and lake evaporation to E601B pan evaporation (\( E_{Lake}/E_{E601B} \)) are given in Fig. 5b. These results show that \( E_{Lake}/E_{D20} \) and \( E_{Lake}/E_{E601B} \) have seasonal variations similar to those of \( E_{D20} \) and \( E_{E601B} \), with the minimum values in June. The values of \( E_{Lake}/E_{E601B} \) and \( E_{Lake}/E_{D20} \) obtained during the observation period are 0.37 and 0.54, respectively. Fig. 5b shows that the seasonal variations of \( E_{Lake}/E_{p(Xainza)} \) are similar to those of \( E_{Lake}/E_{D20} \) and \( E_{Lake}/E_{E601B} \). During the observation period, \( E_{Lake}/E_{p(Xainza)} \) was determined to be 0.49. It is also evident that the conversion coefficients vary by month and cannot be defined as single experimental values.

4. Discussion

As pointed out in the descriptions of Fig. 3 and Table 1, the seasonal variations of \( LE \) and \( H \) show patterns different from that of \( R_{n} \). This phenomenon is likely due to the changes in heat storage in the water body. The residual of the energy balance (\( S - R_{n} - H - LE \)) was assumed to be heat storage, although estimating \( S \) as a residual incorporates all of the errors in the other energy balance terms (Blanken et al. 2000, 2011; Li et al. 2015). As shown in Fig. 6a, \( S \) is positive during most of the observation period, and negative \( S \) is observable only for several days in mid- and late September. The variation of \( S \) suggests that the majority of \( R_{n} \) absorbed by the lake was stored into the water body in May–August and was then released in September during the observation period. The cumulative \( S \) reaches its peak in the middle of September. The observed daily mean surface energy budget components between October 11 and November 10, 2012, using the EC method are also given in Fig. 6b. These results also show that \( S \) is always negative during this period, suggesting that the release of heat stored in the lake contributed to the increases of \( H \) and \( LE \) in October and November.
seasonal variations of $LE$ and $H$ over Lake Serling Co might lag $R_n$ by approximately 4 months, based on the two periods of observational data.

The heat storage capacity of a lake has a remarkable influence on the phases of the seasonal variations of the lake surface turbulent fluxes, as has frequently been reported (Blanken et al. 2011; Schertz et al. 2003; Tasumi 2005). Table 2 provides a comparison of the phase delay between $LE$ and $R_n$ over Lake Serling Co and those of 10 different inland water bodies in different climatic zones and with different water areas and depths. The climate classifications in
Fig. 6. (a) Daily means of energy balance residual ($S = R_n - H - LE$) (red line) and cumulative $S$ (blue line) from April 26 until September 26, 2014 and (b) daily means of surface energy budget components ($R_n$: net radiation, $H$: sensible heat flux, $LE$: latent heat flux, $S = R_n - H - LE$: energy balance residual) from October 11 to November 10, 2012.

Table 2. Comparison of phase delays for different inland water bodies.

<table>
<thead>
<tr>
<th>Lake Name</th>
<th>Location</th>
<th>Area (km²)</th>
<th>Mean Depth (m)</th>
<th>Maximum Depth (m)</th>
<th>Climate¹</th>
<th>Data Period</th>
<th>Delay time² (month)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lake Serling Co, China</td>
<td>88.3°37′N, 31°48′E</td>
<td>2341.2</td>
<td>–</td>
<td>50</td>
<td>Dwb</td>
<td>26 April-26 September 2014</td>
<td>4</td>
<td>This study</td>
</tr>
<tr>
<td>Lake Superior, USA and Canada</td>
<td>47°11′N, 87°16′W</td>
<td>82,100</td>
<td>148</td>
<td>406</td>
<td>Dfc</td>
<td>2.5 years from through 2010</td>
<td>5</td>
<td>Blanken et al. (2011)</td>
</tr>
<tr>
<td>Thau Lagoon, France</td>
<td>43°24′N, 3°36′E</td>
<td>75</td>
<td>4</td>
<td>11</td>
<td>Csa</td>
<td>2.5 year from mid-2008 through 2010</td>
<td>0</td>
<td>Bouin et al. (2012)</td>
</tr>
<tr>
<td>Lake Nam Co, China</td>
<td>30°46′N, 90°59′E</td>
<td>1980</td>
<td>–</td>
<td>95</td>
<td>Dwb</td>
<td>3 years from through 2008</td>
<td>5</td>
<td>Haginoya et al. (2009)</td>
</tr>
<tr>
<td>Lake Erhai, China</td>
<td>25°46′N, 100°09′E</td>
<td>256.5</td>
<td>10</td>
<td>20.7</td>
<td>Cfa</td>
<td>9 years from through 20080</td>
<td>1</td>
<td>Haginoya et al. (2012)</td>
</tr>
<tr>
<td>Lake Ngoring, China</td>
<td>N35°01′N, 107°39′E</td>
<td>610</td>
<td>17</td>
<td>32</td>
<td>Dwb</td>
<td>2 ice free seasons between 2011 and 2012</td>
<td>2</td>
<td>Li et al. (2015)</td>
</tr>
<tr>
<td>Ross Barnett Reservoir, USA</td>
<td>32°26′N, 90°02′W</td>
<td>130</td>
<td>5</td>
<td>8</td>
<td>Cfa</td>
<td>2 years from through 2008</td>
<td>1</td>
<td>Liu et al. (2009)</td>
</tr>
<tr>
<td>Logan’s Dam, Australia</td>
<td>27°34′S, 152°20′E</td>
<td>0.17</td>
<td>–</td>
<td>6</td>
<td>Cwa</td>
<td>September 2009-August 2011</td>
<td>0</td>
<td>McGloin et al. (2014)</td>
</tr>
<tr>
<td>Lake Ikeda, Japan</td>
<td>31°44′N, 130°55′E</td>
<td>10.62</td>
<td>125</td>
<td>233</td>
<td>Cfa</td>
<td>25 years from through 2005</td>
<td>3</td>
<td>Momii and Ito (2008)</td>
</tr>
<tr>
<td>Lake Valkea-Kotinen, Finland</td>
<td>61°14′N, 25°03′E</td>
<td>0.041</td>
<td>2.5</td>
<td>6.5</td>
<td>Dfc</td>
<td>4 ice free seasons between 2005 and 2008</td>
<td>0</td>
<td>Nordbo et al. (2011)</td>
</tr>
<tr>
<td>Lake Taihu, China</td>
<td>31°25′N, 120°13′E</td>
<td>2400</td>
<td>1.9</td>
<td>3.1</td>
<td>Cfa</td>
<td>2 years from through 2012</td>
<td>1</td>
<td>Wang et al. (2014)</td>
</tr>
</tbody>
</table>

¹Koppen-Geiger climate classification symbols: A, tropical; B, arid; C, temperate; D, cold; f, no dry season; s, dry summer; S, steppe; W, desert; w, dry winter; a, hot summer; b, warm summer; c, cold summer; h, hot.²Delay time: the delay time between the peak $LE$ and peak $R_n$ in season.
mean value 2.7 mm d\(^{-1}\) and a total value of 417.0 mm from April 26 to September 26, 2014. Comparing these results with those of two kinds of pan evaporation (D20 pan and E601B pan) and potential evaporation, it was found that the conversion coefficients varied depending on the month and could not be defined as single experimental values.

Monthly conversion coefficients were obtained for May–September in 2014. However, further observations are needed to acquire long-term data to improve the conversion coefficients parameterization. In addition, the daily range of the air temperature was larger than that of the water temperature, suggesting that the turbulent observations might have been affected by the land. Therefore, the average evaporation over the whole lake might be smaller than that reported herein. Overall, the present study provides the seasonal and diurnal variations of the energy fluxes between the largest lake and atmosphere; however, long-term in situ observations are still needed to elucidate how the lake-atmosphere energy and water exchange respond to climate changes.

Acknowledgments

This study was supported by the Natural Science Foundation of China (Grant No. 41430748) and the Strategic Priority Research Program (B) of the Chinese Academy of Sciences (Grant No. XDB03030206). This study was also supported by the Key Projects of Foreign Cooperation of Bureau of International Co-operation Chinese Academy of Sciences (Grant No. 131C11KYSB20150006) and the Natural Science Foundation of China (Grant No. 41190080).

References

Bouin, M. N., G. Caniaux, O. Traulle, D. Legain, and P.


