Environmental Controls on the Surface Energy Budget over a Large Southern Inland Water in the United States: An Analysis of One-Year Eddy Covariance Flux Data

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ABSTRACT

The authors analyzed the surface energy fluxes that were measured by an eddy covariance system over the Ross Barnett Reservoir in Mississippi for a 1-yr period in 2008. On a monthly basis over the course of the year, positive vertical temperature and vapor pressure differences were observed, though negative vertical temperature differences occurred occasionally during some short periods when overwater air masses were warmer than the water surface. Consequently, the unstable atmospheric surface layer (ASL) and sufficient mechanical mixing led to positive sensible $H$ and latent $\lambda E$ heat fluxes. The quantities $H$ and $\lambda E$ were distinctly out of phase with the net radiation $R_n$. The $H$ and $\lambda E$ from the water to the ASL was still substantial on nights with a negative $R_n$ and in winter when $R_n$ was very small. From February to August, approximately 60%–91% of the $R_n$ was used for $H$ and $\lambda E$, with the remainder being stored in the water. Fueled by the previously stored heat in the water, $H$ and $\lambda E$ exceeded $R_n$ by almost 3 times from September to January. Nighttime evaporation represented a large loss of water (i.e., $\lambda E = 82.8 \text{ W m}^{-2}$ at night versus $91.4 \text{ W m}^{-2}$ during the daytime). Intraseasonal and seasonal variations in $H$ and $\lambda E$ were strongly affected by frequent passages of large-scale air masses that were brought in by different synoptic weather systems (e.g., cyclones or anticyclones). The authors’ analysis suggested that this southern reservoir responded to atmospheric forcings on both diurnal and seasonal scales in the same ways as northern lakes of comparable sizes and depths.

1. Introduction

Global hydrologic cycles have intensified as a result of climate change (Huntington 2006). However, the ways in which climate change affects evaporation and the surface energy budget over inland waters remains unclear. With quick responses to climate variability and seasonal perturbations, large inland waters provide a robust indicator about whether climate change has the potential to accelerate the energy and water vapor (and thus evaporation) exchange between inland waters and the atmosphere (Blanken et al. 2000; Oswald and Rouse 2004). The implications of changes in evaporation in response to climate change are important for inland water resource management and hydrologic cycles.

Inland waters (e.g., lakes, reservoirs, wetlands, etc.) act differently than surrounding lands in the exchanges of radiation, energy, water vapor, and trace gases between the water and the overlying atmosphere, thus providing unique spots in terms of their influence on the local, regional, and even global climate (Bates et al. 1993, 1995; Bonan 1995; Eaton et al. 2001; Eugster et al. 2003; Cole et al. 2007; Rouse et al. 2005; Long et al. 2007; Jonsson et al. 2008). Shortwave radiation that is not reflected by the water surface penetrates to deeper layers and is directly absorbed by the water from the surface to a certain depth. The amount of solar radiation absorbed by water decreases exponentially with the water depth and is dependent on water turbidity, which is influenced by suspended organic and inorganic materials (Henderson-Sellers 1986; Hostetler and Bartlein 1990). In addition, water eddy diffusion leads to efficient heat transfer in water layers below the surface (Henderson-Sellers 1986). As a consequence, water surfaces respond differently than land surfaces to diurnal and seasonal changes in the solar radiation forcing, in terms of temporal variations in the surface temperature, though these two types of surfaces have similar surface albedos (e.g., about 0.06 for open water and 0.08 for vegetated lands) (Henderson-Sellers 1986; Bonan 1995). It is known that sensible heat flux $H$ is primarily determined by the air temperature...
difference between the water surface and the overlying atmosphere as well as the turbulent exchange coefficient. Latent heat flux $\lambda E$ is dependent upon vapor pressure differences between the water–atmosphere interface and the overlying atmosphere as well as the turbulent mixing intensity (Henderson-Sellers 1986; Hostetler and Bartlein 1990; Garratt 1994; Bonan 1995). The water–air interface is typically at its saturation point and vapor pressure at this interface is a function of the water surface temperature (Hostetler and Bartlein 1990), thus making the water surface temperature a key element in determining a water–atmosphere exchange. Consequently, temporal variations in $H$ and $\lambda E$ are largely governed by variations in the water surface temperature and the meteorological properties (e.g., wind speeds, air temperature, and humidity) of overwater air masses (Blanken et al. 2003; Schertzer et al. 2003; Blanken et al. 2008, 2011; Spence et al. 2011).

Studies of a large high-latitude lake indicate that the water stores a large amount of solar energy in the spring and summer, leading to a gradual increase in the water temperature (Rouse et al. 2003). Since the water temperature increases much more slowly than the overlying air temperature, there is a thermal inversion in the atmospheric surface layer (ASL), or small vertical temperature difference between the water surface and the overlying atmosphere. During this heating stage, $H$ and $\lambda E$ are usually dampened because of the presence of the stably stratified ASL (Oswald and Rouse 2004). As the seasons progress into fall and winter, when solar radiation becomes low, the overlying air cools faster than the water and leads to a positive vertical temperature difference in the ASL (hereafter, “positive” means that the water surface temperature is higher than the overlying air temperature). It is expected that evaporation usually occurs during this cooling stage, since continental air masses in the ASL are typically drier than the water–air interface, producing positive vertical vapor pressure differences in the ASL (Blanken et al. 2003; Lenters et al. 2005). In some cases, however, condensation may occur when warm and humid air masses pass over the cold water surface, making the vapor pressure of the air masses greater than the saturation pressure in the water–air interface (Blanken et al. 2000; Liu et al. 2009). As long as the mechanical turbulence mixing is sufficient, $H$ and $\lambda E$ and, thus, evaporation are promoted when the ASL is unstable, and dampened when the ASL is stable. In the wintertime, when net radiation $R_n$ is small, $H$ and $\lambda E$ are still substantial. This energetic exchange of sensible and latent heat between the water surface and the atmosphere is fueled by the energy that was previously stored in the water, which acts as an energy source (Rouse et al. 2005).

It is well known that variations in the water surface energy balance reflect the response of the surface to changes in atmospheric forcings. Different regions that are dominated by different synoptic weather patterns are expected to experience distinct seasonal and diurnal variations in the meteorological properties of overlake air masses (i.e., atmospheric forcings). It remains unclear whether the observed physical processes that govern energy and moisture exchanges over large northern high-latitude lakes are applicable to mid- and low-latitude inland waters (Sacks et al. 1994; Vallet-Coulomb et al. 2001; Verburg and Antenucci 2010). For example, a study of the surface energy budget over a midlatitude reservoir in Israel showed that the diurnal course of $\lambda E$ was not in phase with $R_n$ and was more in correlation with the diurnal variations of mean wind speed, whereas diurnal variations of $H$ were more correlated with vertical temperature differences (Assouline et al. 2008). However, almost all of the previous long-term studies over lakes were concentrated in high latitudes (Fig. 1; Nordbo et al. 2011). There is still a lack of direct, long-term measurements of the water surface energy budget over mid- to low-latitude regions.

This study reports eddy covariance measurements and analyzes the surface energy budget over the Ross Barnett Reservoir, a large inland body of water in central Mississippi, during 2008. This is one of the first studies to yield longer-term eddy covariance measurements of the surface energy budget over water (Blanken et al. 2000; Vesala et al. 2006; Blanken et al. 2008; Rouse et al. 2008; Liu et al. 2009; Nordbo et al. 2011; Blanken et al. 2011). The purpose of the present research is to determine the processes that regulate different components of the water surface energy budget, including evaporation, and to study their diurnal, intraseasonal, and seasonal variations in a 1-yr period in 2008. To our best knowledge, this study is the first to conduct the long-term surface energy budget and the surface energy balance closure over a large midlatitude inland body of water, which enables us to compare our results with those that were derived from high latitudes.

2. Site description, instruments, and methods

We measured surface energy fluxes by using an eddy covariance system that was installed in the Ross Barnett Reservoir, central Mississippi (“the reservoir”; 32°26’N, 90°02’W; about 117.5 m MSL; Fig. 1). The reservoir has an area of 130 km² and a water depth of 4–8 m. Since this man-made body of water has relatively uniform water depths, it is expected that there should be no significant spatial variations in vertical water temperature distributions. The eddy covariance system was
mounted approximately 4 m above the water surface on a 5-m aluminum tower (Climatronics Corp.). The tower stood on a stationary square wooden platform attached to four wooden columns that were anchored in the bed of the reservoir. In 2008, the water level varied approximately 0.6 m. The distances from the tower to the shore ranged from about 2 to 14 km and the water depth around the tower was approximately 5 m. We estimated the turbulent flux footprint (i.e., the source area of the fluxes) by using the solution of diffusion equations that were employed by Schuepp et al. (1990). The equations were corrected to account for the influence of atmospheric stability based on the median value of the Obukhov stability length (Blanken et al. 2011). Our calculations indicated that 95% of the total flux was obtained over an upwind distance of 1.8 km. Given the distance of approximately 2–14 km from the tower to the shore, the measurements ensured that the footprint was within the reservoir.

The eddy covariance system that was used in this study consisted of a three-dimensional sonic anemometer (CSAT3; Campbell Scientific, Inc.) and an open-path carbon dioxide/water vapor (CO₂/H₂O) infrared gas analyzer (IRGA; model LI 7500, LI-COR, Inc.). Sensor signals from the eddy covariance system were recorded by a datalogger (model CR5000, Campbell Scientific, Inc.) sampling at 10 Hz. The 10-Hz raw time series data collected in this study were processed and corrected to obtain eddy covariance fluxes by using a postfield data processing program. This program participated in the intercomparison study of the Energy Balance Experiment (EBEX; Mauder et al. 2006) and has been updated since then (e.g., Liu et al. 2009). Briefly, the virtual air temperature was converted to air temperature following the procedure suggested by Campbell Scientific, Inc.’s instruction manual (Campbell Scientific, Inc. 2006). The raw 10-Hz time series data were checked for spikes/noise in which data with automatic gain control (AGC) values (i.e., output from the LI-COR 7500 sensor) that were greater than 70 were removed. Data points were replaced through linear interpolation when their magnitudes exceeded 5 times the half-hour mean standard deviations. Double rotation was used for revolving the coordinate system for each averaging period that, in this study, was a total of 30 min. The fluxes were computed by using a block average rather than a linear detrending method. The quantities $H$ and $\lambda E$ were obtained via 30-min mean covariance between vertical velocity and the respective air temperature and water vapor density fluctuations. Because of air density effects, $\lambda E$ was corrected according to Liu (2005), which is close to that after the Webb, Pearman, and Leuning (WPL) corrections (Webb et al. 1980). Results were removed when they did not pass a quality check (Foken et al. 2004). Data

![Fig. 1. Overview of eddy covariance measurements of the water surface fluxes: 1. Toolik Lake, Alaska, United States (Eugster et al. 2003); 2. Great Bear Lake, Canada (Rouse et al. 2008); 3. Great Slave Lake (Blanken et al. 2000); 4. Ross Barnett Reservoir, Mississippi, United States (Liu et al. 2009; this study; the site is denoted by a triangle in the left inset, and a photo of the study site is shown in the right inset); 5. Alqueva Reservoir, Portugal (Salgado; Le Moigne 2010); 6. Lake Geneva, Switzerland (Assouline et al. 2008); 7. Soppensee, Switzerland (Eugster et al. 2003); 8. Lake Tämnaren, Sweden (Elo 2007; Heikinheimo et al. 1999; Venäläinen et al. 1998); 9. Scharmutzelsee, Germany (Beyrich et al. 2006); 10. Großer Kossenblatter See, Germany (Panin et al. 2006); 11. Lake Merasjärvi, Sweden (Jonsson et al. 2008); 12. Lake Valkea-Kotinen, Finland (Huotari et al. 2011; Vesala et al. 2006); 13. Eshkol Reservoir, Israel (Assouline et al. 2008); 14. Williams Lake, Minnesota, United States (Anderson et al. 1999); and 15. Lake Superior (Blanken et al. 2011; Spence et al. 2011).]
were reported for a 1-yr period in 2008. Considering all of the missing and rejected data, the percent availabilities of the 30-min mean flux observations were 92%, 97%, 95%, 93%, 96%, 89%, 92%, 97%, 95%, 93%, 96%, and 89% for January, February, March, April, May, June, July, August, September, October, November, and December, respectively.

Several micrometeorological variables were measured as 30-min averages of 1-s readings (Fig. 1). Rn was measured with a net radiometer (model Q-7.1, Radiation and Energy Balance Systems, Inc.). Incoming solar radiation was measured by using a silicon pyranometer (model IRR-P, Apogee Inc.) that was mounted on a horizontal boom approximately 1 m above the water surface. Air temperature and relative humidity profiles were measured at four different heights (i.e., approximately 1.9, 3.0, 4.0, and 5.46 m above the water surface) by using four temperature/humidity probes (model HMP45C, Vaisala, Inc.). A wind Sentry unit (model 03001, RM Young, Inc.) was mounted at the top of the tower to measure the wind speed and direction, whereas another three wind speed sensors (model 03101, RM Young, Inc.) were mounted at the same heights as the HMP45C probes. In this study, we reported air temperature $T_a$, atmospheric vapor pressure $e_a$, and wind speed $U$ measured at 5.46 m (Fig. 1). Water surface temperatures $T_s$ were measured by using an infrared temperature sensor (model IR-10, Apogee Inc.) that was mounted on a horizontal boom approximately 1 m above the water surface. This instrument is reported to have an accuracy of $\pm 0.2^\circ$C after factory calibration. Vapor pressure on the water surface $e_w$ was calculated as the saturation vapor pressure at the infrared-determined $T_s$ (Buck 1981; Blanken et al. 2000). Eight water temperature probes (model 107-L, Campbell Scientific, Inc.), all attached to a buoy under the stable platform, were placed at 0.10-, 0.25-, 0.5-, 1.0-, 1.5-, 2.5-, 3.5-, and 4.5-m depth below the water surface. Precipitation totals were measured at half-hour intervals by using an automated tipping-bucket rain gauge (model TE525, Texas Instruments, Inc.). Sensor signals from all slow-response (i.e., 1 s) sensors were also recorded by using the CR5000 data logger at 30-min intervals. All instruments were powered by six deep cycle marine batteries that were charged by two 65-W solar panels (model SP65, Campbell Scientific, Inc.).

### 3. Results

#### a. Climatic and meteorological conditions

Mississippi experiences a humid, subtropical climate with long summers and short, mild winters. Based on the data measured at the Jackson Airport climate monitoring station (http://www.sercc.com/), which is located approximately 15 km southeast of our flux station, the annual mean air temperature from 1963 to 2009 was 18.1$^\circ$C. The monthly average air temperature from 1963 to 2009 ranged from 7.7$^\circ$C in January to 27.6$^\circ$C in July. The datasets indicated that, in 2008, the monthly average air temperature at our flux station increased from 6.9$^\circ$C in January to 27.4$^\circ$C in July and decreased to 8.2$^\circ$C in December ($T_a$ in Table 1). The annual average precipitation from 1963 to 2009 was 1393 mm. Precipitation
in this region is fairly unevenly distributed over the course of a year. The monthly average precipitation from 1963 to 2009 ranged from 87 to 142 mm. Compared with the 47-yr mean, 2008 was a slightly warm and wet year, with the annual mean temperature of 18.3°C and the annual mean precipitation of 1514 mm.

This region is frequently subject to influences of anticyclonic and cyclonic weather systems, which bring in air masses of different meteorological properties. In general, warm and humid air masses are predominant as a result of the influence of southerly winds from the Gulf of Mexico. These southerly winds are associated with occasional incursions of tropical cyclones and the influence of the clockwise circulation of airstreams around the Bermuda high (i.e., a semipermanent, subtropical area of high pressure in the North Atlantic Ocean off the East Coast of North America) that drives warm, humid air masses northward into the United States from the Gulf of Mexico. Northerly and northwesterly winds bring in continental dry and cool/cold air masses from the north to affect this region, particularly during cool months characterized by the passage of cold fronts (Liu et al. 2009). Over the course of the year in 2008, this region was alternately predominated by northerly and southerly winds (Fig. 2). Seasonal patterns in wind directions were also present, with dominant southerly and northerly winds in spring (i.e., March, April, and May), southerly winds in summer (i.e., June, July, and August), easterly and northerly winds in fall (i.e., September, October, and November), and northerly winds in winter (i.e., December, January, and February) (Fig. 2). Distinct meteorological properties associated with the different air masses provide temporal variations in atmospheric forcings. These forcings influence the exchange of heat and water vapor between the water surface and the atmosphere. In addition, the atmospheric forcings on local and regional scales, which are influenced by those on synoptic scales, provide external forcings to govern the water surface exchange processes and their diurnal, intraseasonal, and seasonal variations.

b. General characteristics of $R_n$, $H$, and $\lambda E$

The half-hourly time series data regarding $R_n$, $H$, $\lambda E$, and friction velocity ($u_*$) are shown in Fig. 3. The surface energy balance closure is often used as an indicator of the flux data quality (Wilson et al. 2002). In our study, monthly values of the energy balance closure (EBC), [as described by $EBC = (H + \lambda E)/(R_n - S)$, where $S$ denotes the heat storage change in the water; Table 1] over this open-water surface ranged from 0.85 to 1.17, with a mean of 0.97. The EBC values were greater than 1 (i.e., ranging from 1.03 to 1.17) from February to June and less than 1 for the rest of the months. Particularly, the lowest EBC values occurred in November (0.88), December (0.85), and January (0.87). It is noted that our EBC values observed over such a large, open inland water are close to those obtained over land surfaces (e.g., Wilson et al. 2002).

It is noted that $R_n$ exhibited clear diurnal cycles with positive values during the daytime and negative, or small positive, ones at night. The daytime maximum $R_n$ ranged from about 400 W m$^{-2}$ in January to approximately 800 W m$^{-2}$ in July (Fig. 3a), and the nighttime $R_n$ ranged from about $-150$ to 50 W m$^{-2}$. Passages of warm air masses over this region caused a warmer ASL than on the water surface and, thus, larger incoming longwave radiation than outgoing longwave radiation, leading to occasional positive $R_n$ at night. The quantities $H$ and $\lambda E$ also had diurnal variations in response to variations in atmospheric forcings (e.g., wind speeds, air temperature and humidity, etc., as shown in Fig. 4). The quantity $H$ was relatively small (50 W m$^{-2}$) during the warm months (i.e., April–September) and large during the cool months (i.e., October–March). On a few occasions, $H$ reached about 150 W m$^{-2}$ and declined to about $-100$ W m$^{-2}$. By contrast, $\lambda E$ increased from its daily maximum of about 100 W m$^{-2}$ in the cool months to more than 400 W m$^{-2}$ in the warm months. Negative $\lambda E$ occasionally occurred when dew formed.

It is noted that the diurnal variations of $H$ and $\lambda E$ were actually superimposed with large $H$ and $\lambda E$ pulses (“flux pulses”) that were evident throughout the year (Figs. 3b,c). Following the method in Blanken et al. (2000), we defined a “pulse” as occurring when both of the 24-h mean $H$ and $\lambda E$ are at least 1.5 times the values of the centered 10-day running mean. A total of 38 flux pulses were identified for 2008 under this definition, covering 58 days throughout the year. In general, flux pulses occurred more often in the cool months than in the warm months. We estimated that there were 28 and 10 flux pulses in the cool and warm months, corresponding to 41 and 17 days, respectively. Apparently, $H$ pulses had larger magnitudes in the cool months (55.7 W m$^{-2}$ as averaged for all pulses) than in the warm months (49.9 W m$^{-2}$ as averaged for all pulses), whereas $\lambda E$ pulses had smaller magnitudes in the cool months (129.6 W m$^{-2}$ as averaged for all pulses) than the warm months (218.3 W m$^{-2}$ as averaged for all pulses). These large flux pulses were the direct consequences of synoptic forcings that were associated with high-wind events (i.e., $U$ pulses in Fig. 4b) as reflected by large $u_*$ (Fig. 3d), similar to previous studies (Blanken et al. 2000; Lenters et al. 2005). Previous studies of this site for the cool months indicated that these high-wind events were usually associated with cold, dry air masses immediately behind cold fronts, which created large
Fig. 2. Wind rose diagrams for (a) the entire year of 2008, (b) spring, (c) summer, (d) fall, and (e) winter. The radius of the rose plot is the percentage that a wind speed value is within the given direction range. The different colors indicate the wind speed ranges as explained in the legend. In this study, an increment of 45° in wind directions and an increment of 2 m s⁻¹ in wind speed were used.
vertical differences in temperature as well as in vapor pressure (i.e., large $T_s - T_a$ and $e_v - e_a$ in Figs. 4c,d) in the ASL (Liu et al. 2009, 2011). High-wind events in the warm season were attributed to cool air mass bursts associated with the passages of extratropical cyclones, similar to the summer situations in high latitudes (Lenters et al. 2005). Strong mechanical mixing (as reflected by a large $u^*$) and the enhanced convective ASL associated with these high-wind events promoted turbulent exchanges of sensible and latent heat, generating these episodic $H$ and $\lambda E$ pulses (Blanken et al. 2000; Lenters et al. 2005; Liu et al. 2009, 2011).

c. Environmental controls on diurnal variations in the surface energy fluxes

1) DIURNAL VARIATIONS IN Rn

The general features of $Rn$, $H$, and $\lambda E$ can be further analyzed through their averaged diurnal cycles for each month (Fig. 5). $Rn$ presented a clear, bell-shaped diurnal cycle for each month, with maximum values (positive) in the early afternoons [around 1300 local time (LT)] and minimum values (negative) at night. The daily maximum $Rn$ was largest (757 W m$^{-2}$) in July and smallest (253 W m$^{-2}$) in December. On average, the water gained net radiative heating during the daytime and experienced a net loss of longwave radiation at night. As a consequence, the water surface temperature exhibited a diurnal variation that was similar to that of $Rn$ with approximately 3-h time lags, with the minimum values for water surface temperature occurring in the early morning (about 0900 LT) and the maximum values occurring in the late afternoon (about 1600 LT) (Fig. 6). Our data also indicated an exponential decrease with depths in the diurnal wave amplitudes of water temperatures and the progressive wave phase shift with depths (Fig. 6). However, the diurnal wave amplitudes were generally small. For example, the diurnal wave amplitudes for the water surface temperature were largest in July (5.1°C) and smallest in January (1.4°C). Small diurnal variations in the water temperature at a depth of 4.5 m were still observed, with the largest occurring in February (0.7°C) and smallest in August (0.1°C). Relatively small wave amplitudes for water temperatures at different depths, as compared with those for soil (e.g., Arya 2001), were attributed to the direct heating of the water by solar radiation, which penetrated through the deep layers, as well as the significant heat transfer and vertical mixing in the water layers via large eddy diffusion (Henderson-Sellers 1986).

2) DIURNAL VARIATIONS IN $H$ AND $\lambda E$

The monthly average diurnal cycles of $H$ and $\lambda E$ showed no correspondence to those of $Rn$ (Fig. 5).
outcome reflected that Rn was not the direct driving force for H and $\lambda E$ on diurnal time scales. The H exhibited diurnal sine waves, with maximums occurring in the early morning (i.e., about 0600–0700 LT) and minimums in the late afternoon (i.e., about 1700–1800 LT). The daily maximums for H were largest in December (39.9 W m$^{-2}$) and smallest in August (17.0 W m$^{-2}$), whereas the daily minimums for H ranged from −5.1 W m$^{-2}$ in February to 11.8 W m$^{-2}$ in October. Our results indicated that these diurnal variation patterns for H were best explained by similar patterns in $T_s - T_a$ (Fig. 7a). Though diurnal variations for both water surface temperature and air temperature behaved like sine waves, with minimums in early morning and maximums in late afternoon, water surface temperature amplitudes were consistently much smaller than the overwater air temperature amplitudes for each month (Fig. 8), leading $T_s - T_a$ to reach maximum in the early morning (about 0700–1000 LT) and minimum in the late afternoon (about 1800–2000 LT). During the winter months, $T_a$ became larger than $T_s$ for a certain period in the late afternoon or early evening. The daily $T_s - T_a$ was always greater than zero from June to October and its minimum became negative from January to May and again from November to December. Consequently, such diurnal variation patterns in $T_s - T_a$ created ASL stratifications (reflected by the stability parameter $\zeta = z/L$ where $z$ is the measurement height and $L$ is the Monin–Obukhov length; Stull 1988) that were stable for a certain amount of time in the late afternoon and unstable during the rest of the period (Fig. 9). The strongest unstable stratification ($\zeta = -0.2 \sim -0.7$) of the ASL occurred in the early morning (i.e., around 0900 LT) with the largest $T_s - T_a$ (positive). The weakest unstable stratification ($\zeta = -0.1 \sim -0.2$, which turned into a weak stable stratification with $\zeta$ of up to 0.2) occurred in the early evening (around 1900 LT) with the smallest $T_s - T_a$, which turned into a negative value.

The monthly average diurnal cycles of $\lambda E$ showed two kinds of patterns (Fig. 5). In January, February, November, and December, the diurnal variations in $\lambda E$ roughly followed those of H. During these months, the overwater air was very dry under the influence of continental air masses, and the vapor pressure for the water–air interface was also low because of the low water surface temperature. As a result, diurnal variations in $e_s - e_a$ were fairly small, with its magnitudes varying from about 0.2 to 0.5 kPa (Fig. 7b). However, the mechanical mixing was remarkably strong during those months, as indicated by the high $u_*$ (Table 1). Under these conditions, the
diurnal variations in $\lambda E$ were likely to be controlled more by the ASL stability than by $e_s - e_a$, leading to the close correspondence between $\lambda E$ and $\xi$. For the months from April to October, the diurnal variations in $\lambda E$ followed a bell shape that depicted its maximums in the late afternoon (e.g., around 1500–1700 LT). Because of the large vapor pressure difference (i.e., $e_s - e_a$ that varies from 1.2 to 2.8 kPa) and sufficient turbulent mixing ($u_*$), the variations of $\lambda E$ in the months from April to October were likely to be governed by diurnal variations of $e_s - e_a$ (Fig. 5 versus Fig. 7b). Note that there were no obvious diurnal variations in $\lambda E$ in March. No obvious diurnal variation patterns in $u_*$ were observed.

Assouline et al. (2008) reported the same diurnal variations in $H$ and $\lambda E$ over a midlatitude reservoir in Israel (Fig. 1). They also noticed that the diurnal course of $\lambda E$ from the reservoir was not in phase with Rn and exhibited a higher level of correlation with the diurnal variations of mean wind speed, whereas diurnal variations of $H$ were more correlated with those of $T_s - T_a$. However, $H$ and $\lambda E$ over a lake in Switzerland, which was reported in this same study, did not show clear diurnal variations; such a result is likely due to low wind speeds. It is interesting to note that the diurnal variations in $H$ and $\lambda E$ over high-latitude lakes from May to September, as reported by Nordbo et al.

![Figure 5](https://example.com/figure5.png) Average monthly diurnal cycles of net radiation (Rn) and sensible ($H$) and latent ($\lambda E$) heat fluxes. Note that different scales are used for the y axis in different figures.
(2011) and Venäläinen et al. (1999), also exhibited the same behaviors as those observed over midlatitude lakes. However, in this study, $H$ was slightly larger than that reported in Nordbo et al. (2011), but smaller than that reported in Venäläinen et al. (1999). However, $\lambda E$ in this study was about 2 to 3 times larger than that reported in Venäläinen et al. (1999) and Nordbo et al. (2011). Venäläinen et al. (1999) argued that the differences in magnitudes of $H$ may be explained by the effect of advection: as the overwater fetch increases, the air temperature at the reference level adjusts more toward the temperature of the water surface, leading to smaller temperature differences and, thus, smaller $H$. Similarly, the larger overwater fetch would have increased the air moisture at the reference level, which adjusts toward that of the water–air interface, leading to smaller moisture differences. Therefore, what would be the causes that were responsible for the increased $\lambda E$ in this study versus the results from other research? The increased $\lambda E$ with the increased size of water bodies may be due to the increased overwater fetch, which causes the increased wind speeds and mechanical turbulent mixing (as indicated by our relatively large $u_*$ as compared with previous studies) (Spence et al. 2003; Nordbo et al. 2011).

Fig. 6. Average monthly diurnal cycles of water surface temperatures at different depths.
3) DAYTIME VERSUS NIGHTTIME ENERGY FLUXES

In this study, nighttime is defined as the period when the half-hourly incoming solar radiation \( (S) \) was less than 0.5 W m\(^{-2}\). The results suggested that the diurnal variations of \( H \) and \( \lambda E \) were profoundly out of phase with \( R_n \) (Fig. 5). Therefore, analyzing the surface energy fluxes and control factors separately for daytime and nighttime should enable us to better understand the different mechanisms that influence the daytime and nighttime surface energy budget. In addition, nighttime evaporation is one of the major concerns in quantifying the local and regional water budget in catchments (Liu et al. 2009).

On an annual basis (Table 2), the \( u_\ast \) data indicated that mechanical mixing was slightly stronger at night (\( u_\ast = 0.20 \text{ m s}^{-1} \)) than during the daytime (\( u_\ast = 0.18 \text{ m s}^{-1} \)); such a result corresponds to different wind speed magnitudes for the nighttime (\( U = 3.99 \text{ m s}^{-1} \)) and daytime (\( U = 3.77 \text{ m s}^{-1} \)). As indicated by the Monin–Obukhov stability parameter (\( \zeta \)), the nighttime ASL (\( \zeta = -0.23 \)) over the water was more unstable than the daytime ASL (\( \zeta = -0.17 \)), though vertical temperature differences between the water surface and the overlying ASL were close for both the nighttime (\( \Delta T = 1.5^\circ \text{C} \)) and the daytime (\( \Delta T = 1.4^\circ \text{C} \)). In summary, nights showed slightly stronger mechanical turbulent mixing and more unstably stratified ASL. As a consequence, nighttime \( H \) (21.2 W m\(^{-2}\)) was larger than daytime \( H \) (12.8 W m\(^{-2}\)). However, nighttime \( \lambda E \) (83.9 W m\(^{-2}\)) was still lower than daytime \( \lambda E \) (89.8 W m\(^{-2}\)), probably because of the smaller vertical vapor pressure differences between the water–air interface and the overlying ASL at night (\( \Delta e = 0.75 \text{ kPa} \)) as compared to those in the daytime (1.08 kPa). Our results indicated that nighttime evaporation was a large contributor to the annual water loss from the reservoir.

Monthly variations in daytime and nighttime energy fluxes and meteorological variables were also clearly evident (Fig. 10; Table 2). For all months, the average \( u_\ast \) was slightly larger (i.e., therefore indicating stronger mechanical turbulent mixing) at night than during the daytime, with some months being very close. In general, the monthly average nighttime \( H \) was consistently larger than the monthly average daytime \( H \), with lower values from February to July and slightly higher values for the rest of the year (Fig. 10; Table 2). The ASL was strongly unstable at night compared to that during the daytime for all months. The larger \( H \) at night is best explained by the stronger mechanical mixing (\( u_\ast \)) and larger \( T_s - T_a \) (i.e., a more unstable ASL) at night than during the daytime. In contrast, the monthly average \( \lambda E \) during the daytime was higher than that at night for all months except January. Both daytime and nighttime \( \lambda E \) showed an increase from January to July and a decrease after July. The increased \( e_s - e_a \) was likely the primary factor leading to the increased \( \lambda E \) during the daytime versus the nighttime. However, the causes for the higher nighttime \( \lambda E \) in January remain unclear.

3. d. Environmental controls on seasonal variations in energy partitioning

1) PHYSICAL PROCESSES IN CONTROLLING SEASONAL VARIATIONS OF ENERGY BALANCE

The monthly average \( R_n \) increased as the seasons progressed toward summer, reaching its maximum of 196.2 W m\(^{-2}\) in July and decreasing thereafter to reach its minimum of 25.9 W m\(^{-2}\) in December (Fig. 11a; Table 1). Although \( R_n \), which is always positive in terms of its monthly mean, provided the primary energy source for heating water on a monthly basis, the body of water did not consistently act as a heat sink throughout the year. Instead, the monthly average heat storage change \( S \) increased during the first half of the year and decreased thereafter, suggesting that the water released...
heat to the overlying atmosphere in January, transitioned to gain heat from February through August, and released heat from September to December (Fig. 11b; Table 1). These distinctive seasonal variations in Rn and S had a significant impact on the partitioning of the available energy into $H$ and $\lambda E$. In general, the monthly average of $H$ was small, ranging from 9.7 W m$^{-2}$ in May to 27.0 W m$^{-2}$ in October (Table 1). The $H$ was relatively small from February to July and became larger in all of the other months (Fig. 11). The monthly average $\lambda E$ increased from about 50 W m$^{-2}$ in January to 133.5 W m$^{-2}$ in July, and then decreased to about 55 W m$^{-2}$ by December (Table 1; Fig. 11). It is clear that $\lambda E$ followed monthly variation patterns of Rn quite well, but $H$ showed a completely different behavior from Rn.

The monthly variations in the surface energy partitioning over the water are provided in Table 3. The value of $H/Rn$ was very low, with the annual mean of 0.16, indicating that only a small amount of Rn that was absorbed by the water was used for the sensible heating of the atmosphere. The monthly average of $H/Rn$ was lowest from May to July (0.06–0.08), became larger in February, March, April, August, and September (0.11–0.18),

![Fig. 8](image-url)
and was largest during the other months (0.33–0.80). In contrast, a large portion of the Rn absorbed by the water was utilized for evaporation, with the annual mean $\lambda E/Rn$ of 0.81. Values of $\lambda E/Rn$ showed an overall increase from a mean value of 0.61 for February–April, to 0.63 for May–July, 1.03 for August–October, and 1.89 for November–January. Note that the monthly mean $\lambda E/Rn$ exceeded 1 for September, October (1.60), and became greater than 2 for September (2.18), October (2.27), and November–January (2.93) (Table 3). From February to August, our results indicated that a large portion of the Rn absorbed by the water was transferred to the ASL through turbulent exchanges of $H$ and $\lambda E$ [i.e., $(H + \lambda E)/Rn = 0.64–0.91$]; the remainder of the Rn was stored in the water (i.e., $S/Rn = 0.06–0.27$). From September to December and January, however, the sum of $H$ and $\lambda E$ was roughly 1 to 3 times the level of Rn. The excessive portion of energy that was transferred into the ASL through $H$ and $\lambda E$ was actually fueled by the release of the stored energy from the water, as indicated by the negative values of $S$ in Table 1 and $S/Rn$ in Table 3 for this period.

It is noted that variations in $H$ and $\lambda E$ are well explained by the changes in the temperature and the humidity difference between the surface water and the overlying air, respectively. The monthly averages of both $T_a$ and $T_s$ increased and reached their maximum in July (27.4°C for $T_a$ and 29.3°C for $T_s$), dropping off afterward as the seasons progressed toward winter (Table 1). On a monthly basis, $T_s$ was always larger than $T_a$ throughout the year, creating a positive vertical temperature difference (Fig. 11d). Correspondingly, a thermally unstable, convective ASL developed (i.e., the monthly average $\xi$ was always negative, as shown in Fig. 11c); thus, $H$ was positive in all months. The quantity $\xi$ was lowest in February and March, increased afterward, reached its maximum in October, and then slightly decreased in winter, corresponding to the monthly variation patterns in $T_s - T_a$ (Figs. 11c,d).

The water–air interface was typically at saturation, and the saturation vapor pressure at this interface was a function of the surface temperature. Given that the monthly average of $T_s$ was larger than that of $T_a$ in the ASL, the saturation vapor pressure at the water–air interface, which was quantified by $e_s$, was always greater than the saturation vapor pressure in the ASL at $T_a$. Because it was subject to the influence of air masses from surrounding landscapes, however, the overwater air was not even close to

![Fig. 9. Average monthly diurnal cycles of ASL stability ($\xi$).](image)

**TABLE 2. Monthly averaged daytime and nighttime values of meteorological variables and components of the surface energy fluxes in 2008:** $H$: sensible heat flux (W m$^{-2}$); $\lambda E$: latent heat flux (W m$^{-2}$); $u_w$: friction velocity (m s$^{-1}$); $\xi$: ASL stability parameter ($\xi = z/L$, where $z$ is the height of the eddy covariance system and $L$ is the Monin–Obukhov length); $\Delta T$: temperature difference between the water surface and the overlying air, respectively. The monthly averages of both $T_a$ and $T_s$ increased and reached their maximum in July (27.4°C for $T_a$ and 29.3°C for $T_s$), dropping off afterward as the seasons progressed toward winter (Table 1). On a monthly basis, $T_s$ was always larger than $T_a$ throughout the year, creating a positive vertical temperature difference (Fig. 11d). Correspondingly, a thermally unstable, convective ASL developed (i.e., the monthly average $\xi$ was always negative, as shown in Fig. 11c); thus, $H$ was positive in all months. The quantity $\xi$ was lowest in February and March, increased afterward, reached its maximum in October, and then slightly decreased in winter, corresponding to the monthly variation patterns in $T_s - T_a$ (Figs. 11c,d).

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saturation (i.e., its monthly average RH ranging from 63.9% to 73.6%) (Table 1). The temporal variations in both $e_s$ and $e_a$ led to $e_s - e_a$ that increased as the seasons progressed; it reached its maximum in July and decreased thereafter (Table 1; Fig. 11e). Given the sufficient mechanical mixing (i.e., as reflected by $u^*$ in Fig. 11f) and the unstably stratified ASL (i.e., as reflected by $\zeta$), the monthly variations of $\lambda E$ followed those of $e_s - e_a$ quite well.

2) ROLES OF SYNOPTIC AIR MASSES IN AFFECTING THE SURFACE ENERGY BUDGET

Synoptic weather systems, such as anticyclones and cyclones, alternately affect this region and bring in air masses of different meteorological properties. These different air masses provide distinctive atmospheric forcings and temporal variations on synoptic scales, thus influencing the water surface exchange processes. We calculated the bin average magnitudes of fluxes and other meteorological variables against wind directions in 2008 (i.e., producing one point every $5^\circ$, from $0^\circ$ to $360^\circ$) (Fig. 12). The graph shows clearly that air masses from different directions had varying impacts on overwater meteorological properties and, in turn, affected the turbulent exchanges of $H$ and $\lambda E$. Except for a small direction sector (i.e., roughly $30^\circ$–$150^\circ$), wind speeds were consistent for all directions in each season, though the higher overall wind speeds occurred during the winter, spring, and fall (Fig. 12e). For all seasons, northerly winds with cool/dry air masses enhanced the vertical temperature ($T_s - T_a$) and humidity ($e_s - e_a$).
differences, and southerly winds with warm/humid air masses reduced the differences (Figs. 12c,d). As a consequence, $H$ and $\lambda E$ responded to changes in over-water atmospheric forcings in relation to southerly and northerly winds (Figs. 12a,b). For all four seasons, $H$ and $\lambda E$ were generally smaller and under the influence of southerly winds, with smaller $T_s - T_a$ and $e_s - e_a$ indicated in comparison to those that were under the influence of northerly winds.

However, southerly and northerly winds had different influences on the magnitudes of $e_s - e_a$ and $T_s - T_a$ for different seasons. Apparently, southerly winds reduced and even inverted the vertical temperature difference (i.e., negative $T_s - T_a$) in winter, and somewhat in spring and fall, when warm, humid air masses from the Gulf of Mexico passed over the cold water surface. During these periods, $H$ became negative and $\lambda E$ was close to zero. Northerly winds enhanced $T_s - T_a$ in the spring, fall, and winter, leading to an increase in $H$. Meanwhile, northerly winds increased $e_s - e_a$ and, thus, promoted evaporation in spring, summer, and fall (but not in winter). However, the large $\lambda E$ in winter was more likely due to the enhanced mechanical mixing (i.e., higher wind speeds) than the increased $e_s - e_a$.

Previous studies have indicated significant changes in Northern Hemisphere surface cyclone frequency and intensity in the past decades and projected changes as a result of climate warming (Lambert 1995; McCabe et al. 2001). An important question for future research is whether the intensity, frequency, and duration of these two types of air masses will change in response to climate warming, which would affect the surface energy budget and evaporation rates over water during cool seasons. Our results also suggest that possible shifts in northerly and southerly winds associated with changes in cyclonic and anticyclonic activities would have significant impacts on seasonal and annual variations in the water surface energy budget, evaporation rates, and hydrologic processes of this region.

### Comparisons between midlatitudes and high latitudes

Long-term eddy covariance measurements of the surface energy budget over inland waters are limited. Almost all direct measurements with durations that exceed 5 months were made over high-latitude water lakes, including those over Great Slave Lake, Canada (e.g., Blanken et al. 2000; Rouse et al. 2005); Great Bear Lake, Canada (Rouse et al. 2008); Lake Valkea-Kotinen, Finland (e.g., Vesala et al. 2006; Nordbo et al. 2011); and Lake Superior (Blanken et al. 2011; Spence et al. 2011) (as summarized partly in Nordbo et al. 2011; see Fig. 1 for their locations). It should be noted that, because of the incomparable sizes and depths of these water bodies, comparisons between our study over a midlatitude reservoir and those over high-latitude lakes may not be comprehensive in terms of seasonal variations in the water surface energy budget. However, such comparisons may allow us to analyze similarities and differences in the relative predominance of different physical processes across latitudes.

High-latitude lakes experience two distinct periods over the course of a year: the ice-covered period and the ice-free period. During the ice-covered period, which lasts from approximately December/January to May/June (e.g., Blanken et al. 2000; Rouse et al. 2005, 2008; Vesala et al. 2006; Nordbo et al. 2011), the water–atmosphere energy exchange ceases and $H$ and $\lambda E$ are negligible. Instead, the reservoir in this study is ice free and the energy exchange occurs throughout the year. During the ice-free period for the high-latitude lakes and the corresponding period for the reservoir, $\text{Rn}$

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### Table 3. Monthly mean values of the surface energy partitioning in 2008: $S_o$: incoming solar radiation (W m$^{-2}$); $\text{Rn}$: net radiation (W m$^{-2}$), $H$: sensible heat flux (W m$^{-2}$); $\lambda E$: latent heat flux (W m$^{-2}$); $S$: the heat storage change in the water; and $B$: Bowen ratio, the ratio of sensible heat flux to latent heat flux.

<table>
<thead>
<tr>
<th>Month</th>
<th>$\text{Rn}/S_o$</th>
<th>$H/S_o$</th>
<th>$\lambda E/S_o$</th>
<th>$S/S_o$</th>
<th>$H/\text{Rn}$</th>
<th>$\lambda E/\text{Rn}$</th>
<th>$S/\text{Rn}$</th>
<th>$(H + \lambda E)/\text{Rn}$</th>
<th>$B$</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.30</td>
<td>0.24</td>
<td>0.54</td>
<td>-0.27</td>
<td>0.80</td>
<td>1.80</td>
<td>-0.91</td>
<td>2.60</td>
<td>0.44</td>
</tr>
<tr>
<td>February</td>
<td>0.60</td>
<td>0.11</td>
<td>0.39</td>
<td>0.07</td>
<td>0.18</td>
<td>0.64</td>
<td>0.12</td>
<td>0.82</td>
<td>0.28</td>
</tr>
<tr>
<td>March</td>
<td>0.57</td>
<td>0.06</td>
<td>0.30</td>
<td>0.12</td>
<td>0.11</td>
<td>0.53</td>
<td>0.21</td>
<td>0.64</td>
<td>0.21</td>
</tr>
<tr>
<td>April</td>
<td>0.71</td>
<td>0.09</td>
<td>0.48</td>
<td>0.08</td>
<td>0.13</td>
<td>0.66</td>
<td>0.11</td>
<td>0.79</td>
<td>0.19</td>
</tr>
<tr>
<td>May</td>
<td>0.69</td>
<td>0.04</td>
<td>0.38</td>
<td>0.18</td>
<td>0.06</td>
<td>0.54</td>
<td>0.27</td>
<td>0.60</td>
<td>0.11</td>
</tr>
<tr>
<td>June</td>
<td>0.72</td>
<td>0.06</td>
<td>0.48</td>
<td>0.10</td>
<td>0.08</td>
<td>0.66</td>
<td>0.14</td>
<td>0.74</td>
<td>0.11</td>
</tr>
<tr>
<td>July</td>
<td>0.75</td>
<td>0.05</td>
<td>0.51</td>
<td>0.09</td>
<td>0.06</td>
<td>0.68</td>
<td>0.12</td>
<td>0.74</td>
<td>0.09</td>
</tr>
<tr>
<td>August</td>
<td>0.69</td>
<td>0.09</td>
<td>0.55</td>
<td>0.04</td>
<td>0.13</td>
<td>0.79</td>
<td>0.06</td>
<td>0.91</td>
<td>0.17</td>
</tr>
<tr>
<td>September</td>
<td>0.63</td>
<td>0.09</td>
<td>0.64</td>
<td>-0.05</td>
<td>0.15</td>
<td>1.02</td>
<td>-0.07</td>
<td>1.17</td>
<td>0.14</td>
</tr>
<tr>
<td>October</td>
<td>0.48</td>
<td>0.16</td>
<td>0.61</td>
<td>-0.13</td>
<td>0.33</td>
<td>1.27</td>
<td>-0.26</td>
<td>1.60</td>
<td>0.26</td>
</tr>
<tr>
<td>November</td>
<td>0.33</td>
<td>0.18</td>
<td>0.55</td>
<td>-0.18</td>
<td>0.53</td>
<td>1.65</td>
<td>-0.53</td>
<td>2.18</td>
<td>0.32</td>
</tr>
<tr>
<td>December</td>
<td>0.31</td>
<td>0.25</td>
<td>0.65</td>
<td>-0.27</td>
<td>0.80</td>
<td>2.12</td>
<td>-0.88</td>
<td>2.93</td>
<td>0.38</td>
</tr>
<tr>
<td>Mean</td>
<td>0.61</td>
<td>0.10</td>
<td>0.49</td>
<td>0.02</td>
<td>0.16</td>
<td>0.81</td>
<td>0.04</td>
<td>0.96</td>
<td>0.20</td>
</tr>
</tbody>
</table>

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shows very similar variation patterns that increase to their maximums in June and decrease afterward (e.g., Rouse et al. 2005; this study). However, $H$ and $\lambda E$ display different temporal variation patterns. Over the high-latitude lakes, $H$ and $\lambda E$ are dampened during the first part of the ice-free period until late summer (e.g., the end of September for the Great Slave Lake), when the atmosphere warms faster than the water and creates a stable ASL. During this period, a substantial amount of $Rn$ is stored in the water, leading to an increase in water temperature (i.e., a period typically referred to as the heating stage) (e.g., Rouse et al. 2003; Oswald and Rouse 2004). Our observations also indicate that there is a heating stage for the midlatitude reservoir. However, $H$ and $\lambda E$ over the reservoir are typically transferred from the water surface to the overlying atmosphere (i.e., positive $H$ and $\lambda E$) during this heating stage, although occasional negative $H$ and $\lambda E$ occur during short periods of time when the overlying atmosphere is wetter and warmer than the water–air interface. Since the shallow bodies of water in this reservoir have a lower heat capacity and respond to changes in atmospheric forcings more quickly than the deep bodies of water in the Great Slave Lake, thus leading to different exchange processes over water, we believe that these different $H$ and $\lambda E$ variation patterns that are described above are mainly attributed to different water depths. In fact, this substantially different behavior in the seasonal variations of $H$ and $\lambda E$ also occur between shallow and medium/deep lakes over high latitudes, as documented in Rouse et al. (2005). With a small heat capacity, shallow lakes warm more rapidly early in the season than deep lakes, promoting steep temperature differences over the lake. This condition favors substantial flux exchanges that are consistent with our observations over this southern reservoir. Based on the data present in Vesala et al. (2006) and Nordbo et al. (2011) over a high-latitude shallow lake, $H$ and $E$ in the early season are also comparable to our observations, reflecting similarities in fluxes exchange in this season. As seasons progress into winter, $Rn$ decreases and the overwater air masses are drier and cooler than the water surface. Driven by the energy released by the water, $H$ and $\lambda E$ are substantial and even larger than $Rn$ for both the high-latitude lakes and the reservoir (Blanken et al. 2000; this study). As a result, the water temperature gradually decreases during this period, which is typically referred to as the cooling stage.

Throughout the year, $H$ and $\lambda E$ over the reservoir and the high-latitude lakes are strongly modulated by alternative incursions of large-scale air masses brought in by different synoptic weather systems (e.g., cyclones or anticyclones) (Blanken et al. 2000; Lenters et al. 2005; Liu et al. 2009, 2011), leading to a substantial contribution to the water surface energy exchange. It is also interesting to note that both the midlatitude water bodies (e.g., Assouline et al. 2008; this study) and the high-latitude lakes (e.g., Vesala et al. 2006; Nordbo et al. 2011) exhibit very similar diurnal variation patterns in $H$ and $\lambda E$. Typically, $H$ displays sine waves that reach their maximum in the morning and minimum in the late afternoon, and $\lambda E$ shows bell shapes that indicate minimums in the morning and maximums in the late afternoon (Vesala et al. 2006; Nordbo et al. 2011; this study). Our comparisons suggest that the southern reservoir in this study responds to the atmospheric forcings on both diurnal and seasonal scales in a manner that resembles that of northern lakes with comparable sizes and depths. Given a wide latitudinal gradient, these same exchange mechanisms reflect the
generic responses of water bodies to atmospheric forcings (Liu et al. 2011).

4. Conclusions

Over the course of a full year in 2008, we analyzed the components of the surface energy budget for the open-water surface of the Ross Barnett Reservoir in Mississippi. Positive vertical temperature and vapor pressure differences (i.e., positive $T_v - T_a$ and $e_v - e_a$) were observed during each month of the year, though negative differences occurred during certain short periods of time when the overwater air masses were warmer and drier than the water surface. The annual mean Rn, $H$, and $\lambda E$ were 108.4, 17.1, and 87.1 W m$^{-2}$, respectively. Approximately 81% of the Rn absorbed by the water was transferred to the atmosphere through $\lambda E$ and the remainder was transferred through $H$. Our results indicated that $H$ and $\lambda E$ were distinctively out of phase with Rn on diurnal and seasonal scales. Consequently, the turbulent transfer of $H$ and $\lambda E$ to the ASL was still substantial during nights with a negative Rn and in winter when Rn was small. In 2008, $H$ at night (21.2 W m$^{-2}$) was almost 1.6 times larger than it was during the daytime (12.8 W m$^{-2}$). Nighttime evaporation constituted a large evaporative loss of water from the reservoir (i.e., $\lambda E = 82.8$ W m$^{-2}$ at night versus 91.4 W m$^{-2}$ during the daytime). From February to August, about 60%–91% of the Rn was used for the turbulent exchanges of $H$ and $\lambda E$, with the remainder of the Rn being stored in the water. From September to January, $H$ and $\lambda E$ exceeded Rn by almost 3 times, with the energy deficit being supplied by the release of the previously stored heat in the water. Intraseasonal and seasonal variations in the turbulent exchanges of $H$ and $\lambda E$ were strongly affected by alternative incursions of large-scale air masses that were brought in by different synoptic weather systems (e.g., cyclones or anticyclones) throughout the year. Southerly winds with warm and humid air masses generally suppressed turbulent exchanges of $H$ and $\lambda E$. Our analysis suggested that the southern reservoir in this study responded to the atmospheric forcings on both diurnal and seasonal scales in a manner that replicated the findings at northern lakes of comparable sizes and depths.

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