Eddy Covariance Measurements of Water Vapor and Energy Flux over a Lake in the Badain Jaran Desert, China (Postprint)

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Abstract

Exploring the surface energy exchange between atmosphere and water bodies is essential to gain a quantitative understanding of regional climate change, especially for the lakes in the desert. In this study, measurements of energy flux and water vapor were performed over a lake in the Badain Jaran Desert, China from March 2012 to March 2013. The studied lake had about a 2-month frozen period (December and January) and a 10-month open-water period (February-November). Latent heat flux (LE) and sensible heat flux (Hs) acquired using the eddy covariance technique were argued by measurements of longwave and shortwave radiation. Both fluxes of longwave and shortwave radiation showed seasonal dynamics and daily fluctuations during the study period. The reflected solar radiation was much higher in winter than in other seasons. LE exhibited diurnal and seasonal variations. On a daily scale, LE was low in the morning and peaked in the afternoon. From spring (April) to winter (January), the diurnal amplitude of LE decreased slowly. LE was the dominant heat flux throughout the year and consumed most of the energy from the lake. Generally speaking, LE was mostly affected by changes in the ambient wind speed, while Hs was primarily affected by the product of water-air temperature difference and wind speed. The diurnal LE and Hs were negatively correlated in the open-water period. The variations in Hs and LE over the lake were differed from those on the nearby land surface. The mean evaporation rate on the lake was about 4.0 mm/d over the entire year, and the cumulative annual evaporation rate was 1445 mm/a. The cumulative annual evaporation was 10 times larger than the cumulative annual precipitation. Furthermore, the average evaporation rates over the frozen period and open-water period were approximately 0.6 and 5.0

mm/d, respectively. These results can be used to analyze the water balance and quantify the source of lake water in the Badain Jaran Desert.

Full Text

Preamble

Eddy Covariance Measurements of Water Vapor and Energy Flux Over a Lake in the Badain Jaran Desert, China

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Abstract

Exploring surface energy exchange between the atmosphere and water bodies is essential for gaining a quantitative understanding of regional climate change, particularly for lakes in desert environments. In this study, measurements of energy flux and water vapor were conducted over a lake in the Badain Jaran Desert, China from March 2012 to March 2013. The studied lake experienced approximately a 2-month frozen period (December and January) and a 10-month open-water period (February–November). Latent heat flux (LE) and sensible heat flux (Hs) acquired using the eddy covariance technique were evaluated alongside measurements of longwave and shortwave radiation. Both radiation fluxes exhibited seasonal dynamics and daily fluctuations throughout the study period. Reflected solar radiation was substantially higher in winter than in other seasons. LE displayed distinct diurnal and seasonal variations; on a daily scale, LE was low in the morning and peaked in the afternoon. From spring (April) to winter (January), the diurnal amplitude of LE decreased gradually. LE represented the dominant heat flux throughout the year and consumed most of the energy from the lake. Generally, LE was most strongly influenced by changes in ambient wind speed, while Hs was primarily affected by the product of water-air temperature difference and wind speed. The diurnal LE and Hs were negatively correlated during the open-water period. The variations in Hs and LE over the lake differed markedly from those observed on the nearby land surface. The mean evaporation rate over the lake was approximately 4.0 mm/d for the entire year, with a cumulative annual evaporation of 1445 mm/a. The cumulative annual evaporation was 10 times larger than the cumulative annual precipitation. Furthermore, the average evaporation rates during the frozen period and open-water period were approximately 0.6 and 5.0 mm/d, respectively. These results can be used to analyze the water balance and quantify the source of lake water in the Badain Jaran Desert.

Keywords: eddy covariance; energy flux; radiation; evaporation; precipitation; lake; Badain Jaran Desert

1 Introduction

Lakes cover extensive areas of the Earth' s surface and make vital contributions to regional energy balance and the water cycle. Through biogeochemical and biophysical processes, lakes—having different heat capacity and albedo than surrounding land surfaces—can influence climate at local, regional, and even global scales (Bonan, 1995; Small et al., 1999; Magnuson et al., 2000; Walter et al., 2006; Long et al., 2007; Liu et al., 2012; Lee et al., 2014). Environmental changes, open-water surface hydrological processes, and the water-atmosphere system all significantly influence surface energy exchange, making it important to quantitatively understand how these factors affect climate and weather (Hostetler and Bartlein, 1990; Bates et al., 1993, 1995; Tsuang et al., 2001; Li et al., 2016). Previous lake studies have found that lakes can act as substantial energy reservoirs that introduce massive seasonal thermal lags in the surrounding landscape due to their large heat capacities (Blanken et al., 2000; Schertzer et al., 2003; Downing et al., 2006; Nordbo et al., 2011). Intraseasonal surface evaporation and energy radiation of lakes are significant and closely correlated with strong winds and other weather events (Rouse et al., 2003, 2005; Lenters et al., 2005; Liu et al., 2009; Kettle et al., 2012). However, few studies have focused on surface evaporation and energy radiation of lakes in regions with strong winds, particularly in desert regions.

Eddy covariance (EC) techniques have been widely and successfully used to measure carbon, water, and energy fluxes between the atmosphere and water bodies, forests, and other terrestrial ecosystems (Barford et al., 2001; Xu et al., 2013; Baldocchi, 2014; McGloin et al., 2014; Jung et al., 2017), as well as lakes (Blanken et al., 2000; Liu et al., 2009; Xiao et al., 2013; Biermann et al., 2014). However, few EC measurements have been made over inland lakes, and little or no work has been conducted on lakes in deserts due to the harsh environment. Meanwhile, it should be noted that energy flux depends on many factors, such as lake size, depth, and surrounding terrain. Consequently, EC measurements need to be conducted across regions with different climate and weather conditions (Lenters et al., 2005; Liu et al., 2009). Furthermore, the EC observation network needs to be supplemented by studies of lakes in deserts to provide reliable and comprehensive knowledge of energy and water vapor fluxes over different terrains.

The Badain Jaran Desert is one of the driest regions in China, yet more than 110 perennial lakes are distributed among megadunes in its hinterland (Wang et al., 2016). These lakes vary in shape, size, and salinity. Previous studies have investigated the geochemistry of groundwater and lakes (Yang and Williams, 2003; Chen et al., 2004; Gates et al., 2008a, b), lake-level fluctuations recorded by palaeoshorelines (Yang et al., 2010), aeolian geomorphology and evolution of megadunes (Yang and Williams, 2003; Zhu et al., 2012; Rioual et al., 2013), and the key scientific question of lake water sources (Dong et al., 2016).

From a water balance perspective, water vapor exchange is an important component of the water budget; from an energy balance perspective, lake-atmosphere exchange represents a significant part of energy transfer, particularly lake evaporation. Accurately quantifying lake evaporation presents a particular research challenge, especially in water-scarce areas (Xu et al., 2013; McGloin et al., 2014). Studying the water cycle and energy flux in the Badain Jaran Desert is of great significance for understanding lake-atmosphere interactions and lake water sources, as well as for water resource management.

The aims of this study are to quantify the energy flux over a lake in the Badain Jaran Desert to provide detailed knowledge about thermal conditions, estimate lake evaporation using EC measurements, and determine the relationships between sensible heat flux (Hs) and latent heat flux (LE) with several environmental factors. This study may fill a gap in current knowledge of lake-atmosphere interactions over lakes in arid regions, particularly deserts. The results could provide a scientific reference for future studies on lake water sources.

2.1 Study Area

The Badain Jaran Desert (39°04 15 -42°12 23 N, 99°23 18 -104°34 02 E) lies in the western part of Inner Mongolia Autonomous Region, China. It is surrounded by the ancient Juyan Lake and Guaizi Lake to the north, the Beida Mountain and Heli Mountain to the south, the ancient Gurinai Lake and the Zhengyi Gorge of the Heihe River to the west, and the Yabrai Mountain to the southeast (Ma et al., 2014). The Badain Jaran Desert covers an area of approximately 52×10^{3} km^2 (Zhu et al., 2010). The prevailing wind direction in the interior and surrounding regions is northwesterly in spring, easterly in summer and autumn, and westerly in winter (Chen, 2011). Wind speeds in the northern part of the desert are much greater than those in the southern part (Yang et al., 2003; Dong et al., 2004). The annual mean wind speed varies from 2.8 to 4.6 m/s, with the highest wind speeds observed in April and May each year. The annual mean diurnal temperature range is 34.4°C (Ma et al., 2014), and the mean temperatures in winter and summer are -9.1°C and 25.3°C, respectively. The Badain Jaran Desert is therefore classified as a "cold desert" (Warner, 2004). Mean annual precipitation is 90.1–115.4 mm in the southern desert and 35.2–42.9 mm in the northern desert, characterizing the climate as continental arid. Shallow groundwater and springs are present around lakes in the desert, with total dissolved solids (TDS) less than 1 g/L. The lakes are primarily located in the southeastern desert, on leeward slopes of dunes in interdune depressions.

Yindeertu Lake, located in the Badain Jaran Desert, is a large lake with a small island far from the shore (39°50 N, 102°27 E; 1169 m a.s.l.; Fig. 1 [Figure

1: see original paper]). It has a surface area of approximately 1.03 km^2. The measured maximum depth is approximately 9.4 m, and the mean depth is about 5.6 m. No runoff has been observed over the lake. The TDS content exhibits significant seasonal variations, ranging from 140 to 220 g/L. During the study period (March 2012 to March 2013), freeze-up and ice break-up occurred in late December and early March, respectively. The dominant plant species around the lake are *Artemisia desertorum* and *Glycyrrhiza yunnanensis*. The lake is surrounded by dunes ranging from 200 to 300 m in height. The dunes and lakes in the Badain Jaran Desert are relatively stable and do not move within a single year.

2.2 Experimental Measurements

The eddy covariance (EC) measurement system (E1) was located in the center of Yindeertu Lake, 250 m from the nearest shore (Fig. 1c). This distance is sufficient to satisfy flux measurement requirements for all wind directions. The radiometers were mounted on the water surface beside the EC system. Data were collected from March 2012 to March 2013.

In principle, the EC measurement method provides direct estimates of momentum flux, sensible heat flux (Hs), and latent heat flux (LE). In this study, these turbulent fluxes were measured by an EC system comprising a three-dimensional sonic anemometer (R3-50, GILL, UK) and an open-path CO_2/H_2O infrared gas analyzer (LI-7500A, Li-COR, USA). These instruments were set at a height of 2.3 m and separated by a horizontal distance of 20 cm. Wind velocities in three dimensions and atmospheric temperature fluctuations were measured, together with variations in water vapor density. Sensor signals were acquired with a logger (LI-7550, Li-COR, USA) at 10 Hz. Relative humidity and air temperature were obtained using Humicap sensors (HMP155, Vaisala, Finland) installed at 3.0 m height within airtight double-steel radiation shields. Water surface temperature was determined using an infrared radiometer (Apogee, SI-111, USA) installed at 1.5 m above the water surface beside the EC system. Net radiation was obtained through a four-component net radiometer (NR01, Hukseflux, Netherlands) at 1.5 m above the water surface, as with the infrared radiometer. A tipping-bucket rain gauge (HOBO RG3-M, Onset, USA) installed at 1.5 m height recorded total precipitation at 30-min intervals automatically. Slow-response sensor signals were obtained with a CR-3000 data logger (Campbell Scientific Inc., USA), averaged over 30-min intervals. All sensors were fast-response and small enough to cause no flow disturbances, meeting the requirements for the EC technique (Lorrai et al., 2010; Lemaire et al., 2017).

Another EC measurement system (E2) was located on nearby flat sandy land on the lower part of a megadune beside Sumubarunjilin, 6.68 km from the lake-based EC system (E1). This EC system included the same types of instruments as the lake-based system (Hu et al., 2015).

2.3 Data Processing

Monthly averages of air temperature, water surface temperature, albedo over the lake, wind speed, relative humidity, water vapor density, and total precipitation were obtained through the EC system.

The values of Hs (W/m^2) and LE (W/m^2) were determined using the EC method as follows (Eqs. 1 and 2, respectively):

$$Hs = \rho_\alpha c_\rho \overline{w'T'} \tag{1}$$

$$LE = \rho_\alpha L_v \overline{w'q'} \tag{2}$$

where ρ_α (kg/m^3) is air density; c_ρ ($J/(kg \cdot K)$) is the specific heat of air at constant pressure, i.e., 1004 $J/(kg \cdot K)$; L_v (J/kg) is the latent heat of water vapor, which can be obtained from an equation of air temperature (Aubinet et al., 2000); and the parameters w' (m/s), T' (K), and q' (kg/kg) are the deviations from the temporal averages of vertical wind speed, air temperature, and specific humidity, respectively. Furthermore, the overbars in Equations 1 and 2 indicate the time average (over 30 min in this study).

The component fluxes of the radiation balance were expressed as Equation 3:

$$R_n = DR - UR + DLR - ULR \tag{3}$$

where R_n, DR, and UR are the net all-wave radiation, incident solar radiation, and reflected solar radiation, respectively; ULR is the outgoing longwave radiation from the lake surface; and DLR is the incoming longwave sky radiation. The units of all these parameters are W/m^2.

Monthly variations of R_n, DR, UR, DLR, ULR, $ULR - DLR$, the daily total evaporation rate (ET), and the seasonal and daily variations of LE and Hs were explored in four months representing different seasons: January (winter), April (spring), July (summer), and October (autumn). Correlation analyses were performed to identify the driving forces behind heat fluxes. Finally, comparisons of heat fluxes between the lake and land surface, and between evaporation and precipitation, were obtained.

2.4 Data Quality Control and Flux Footprint Analysis of EC

After periodic calibration and instrument maintenance, raw data were obtained from the EC system at 10 Hz. These data were processed using EddyPro Express post-processing software (Version 6.2, Li-COR, USA) following recommended procedures. All data underwent rigorous quality control through a uniform procedure.

The key step of the EC technique involves both calculation and correction. After extensive debate and examination of field observation results, we established standards for data collection, calculation, and correction. Specifically, we removed raw data outliers and applied corrections for angle-of-attack errors caused by imperfections in the sonic anemometer (Gash and Dolman, 2003; Van der Molen et al., 2004) using Nakai correction functions (Nakai et al., 2006). A double rotation was applied to the outliers (Aubinet et al., 2000). Compensation was used to account for the time lag between the gas analyzer and sonic anemometer measurements. The average vertical deflection in wind direction was -1.1°. Frequency response correction (Moncrieff et al., 2005), sonic temperature correction (Van Dijk et al., 2004), and density fluctuation correction (WPL-correction) (Webb et al., 1980) were applied (Hu et al., 2015). We also conducted a turbulence stationarity test (Foken and Wichura, 1996) and an overall turbulence characteristic test, evaluated data quality, and divided data according to quality level. Furthermore, we identified data requiring interpolation or that were missing. Interpolated data accounted for 29% of the total samples.

Several gap-filling methods were considered (Aubinet et al., 2000; Falge et al., 2001; Liu et al., 2013). Data recorded during dew events and rainy days were not utilized to avoid errors resulting from liquid water on the sensor window of the CO_2/H_2O analyzer. An important thermodynamic factor—the energy of evaporation—mainly came from DR, as did Hs and LE. The DR data were relatively reliable and continuous. Therefore, gaps in daytime data were filled using the relationship between DR and measured Hs, LE, and ET. In this study, data gaps were filled by cumulative calculation. If data were missing for less than 3 days, they were interpolated. In contrast, data missing for more than 3 days were not interpolated unless accumulated data were required. Nighttime data gaps were filled with data from adjacent dates by linear interpolation. Half-hourly Hs and LE were finally obtained using the EC method.

It is important to quantify the sampling area of turbulent flux measurements (Hs and LE), known as the flux footprint, defined as the upwind area contributing to the flux. We assessed the flux footprint using the Lagrangian particle dispersion model proposed by Kljun et al. (2004). During daytime, there was no apparent prevailing wind direction on the lake, although northerly frequencies were relatively large. However, there was an obvious change in prevailing wind direction between day and night: southerly during daytime and northwesterly during nighttime. The source areas making the most important contribution over the whole day, during nighttime, and during daytime were located at distances of 40.6, 38.2, and 42.9 m from the EC system, respectively. Approximately 70% of the source areas were located at distances of 137.3, 138.5, and 136.2 m from the EC system over the whole day, during nighttime, and during daytime, respectively. Because the distance from the nearest lake shore exceeded 250 m, these results indicated that observation conditions substantially met the requirements for successful flux observation, as data were obtained from the lake.

3.1 General Characteristics of Over-Lake Meteorology

Monthly average air temperatures during the observation period showed that from March 2012 to March 2013, the highest air temperature occurred in July 2012, with a monthly average of 27.7°C, while the lowest occurred in January 2013, with a monthly average of -6.7°C. Air temperature over the lake was above 0°C from March to December and below 0°C from December to February of the following year (Fig. 2a [Figure 2: see original paper]). The maximum (35.5°C) and minimum (-19.3°C) half-hourly temperature averages were found in August and December, respectively (Fig. 2a). Based on observed surface temperature and albedo over the lake from March 2012 to March 2013, we concluded that the lake had approximately a 2-month frozen period (December and January). These two months served as a basis for comparison. When the lake was frozen, variations in lake surface temperature were quite similar to those in air temperature. The highest and lowest lake surface temperatures were 30.1°C and -13.4°C, respectively (Fig. 2b). The mean wind speed was 1.7 m/s at 2.5 m height, with a maximum half-hourly average wind speed of 10.1 m/s. Wind speeds in summer were greater than those in winter (Fig. 2c). Relative humidity fluctuated during the study period, averaging 35.9% (Fig. 2d). No humidity data were collected in August due to sensor malfunction. Total precipitation was 145 mm during the study period, with most occurring from March to September (Fig. 2e). An extreme precipitation event occurred on 20 July 2012 (47 mm), which was extremely heavy for a desert area.

3.2 Variations in Radiation

Variations in R_n, DR, UR, DLR, ULR, and $ULR - DLR$ from March 2012 to March 2013 are shown in Figure 3 [Figure 3: see original paper]. Generally, R_n, DR, DLR, and ULR from the lake surface exhibited significant seasonal variations (Figs. 3a, b, d, and e). However, UR displayed a completely different pattern (Fig. 3c), being significantly higher in January and February than in other months. This may be because UR was closely related to underlying surface conditions. The lake was covered with ice from December 2012 to February

2013, and the albedo of the ice surface was much higher than that of open water. Therefore, UR values were high during this period. No snow was present on the lake during the frozen period, so the albedo was entirely from ice without snow contribution.

Seasonal and daily variations in the five radiation types (including R_n, DR, UR, DLR, and ULR) are presented in Figure 4 [Figure 4: see original paper]. The diurnal amplitude of longwave radiation (i.e., DLR and ULR) was smaller than that of shortwave radiation (i.e., DR and UR). Net longwave radiation on the lake surface (i.e., $ULR - DLR$) was always positive during the study period (Fig. 3f), indicating that the lake continuously lost energy through longwave radiation. Net longwave radiation exhibited two peaks in June and November.

Both shortwave and longwave radiation fluxes showed seasonal dynamics and daily fluctuations, but differed in several specific aspects. DR exhibited smooth, symmetric changes during the day and peaked at solar noon (Fig. 4b), suggesting that clouds are relatively rare in the Badain Jaran Desert. In comparison, the peak value of UR was larger in January than in other months, and UR exhibited more complex, non-symmetric changes in all months except January (Fig. 4c). The values of ULR and DLR were positive during both daytime and nighttime, with daytime values greater than nighttime values (Figs. 4d and e). The fluctuations of ULR and DLR were smaller than those of other radiation types (including R_n, DR, and UR). The variations of R_n were similar to those of DR, with positive peaks occurring at noon (Figs. 4a and b). Furthermore, R_n values were negative at night, with the largest amplitude occurring in midsummer.

In general, all radiation values were larger in summer (July) than in winter (January), except for UR values, which were greater in winter than in summer (Fig. 4). The maximum R_n and DR values usually occurred during 12:00–14:00 (LST), the maximum UR value appeared at 09:30 in October, and the maximum DLR and ULR values were observed between 15:00 and 18:00.

3.3 Variations in Heat Fluxes

During the study period, LE exhibited remarkable seasonal changes, whereas Hs did not show obvious seasonal variations (Fig. 5 [Figure 5: see original paper]). LE increased from January to July and decreased from July to January of the following year. The minimum and maximum LE values were -25.3 and 799.0 W/m^2, respectively, with an average of 122.7 W/m^2. Positive LE values indicate heat loss from the lake via evaporation, while negative LE values suggest condensation. Over the course of the year, LE was predominantly positive, with negative values occurring only occasionally in winter. The average Hs was 5.7 W/m^2, much smaller than the average LE. Hs varied slightly throughout the year, with larger values observed in winter.

Compared to Hs, LE exhibited clear seasonal and diurnal variations (Fig. 6 [Figure 6: see original paper]). On a daily scale, LE was low in the morning and peaked in the afternoon, whereas the maximum Hs occurred in the morning and the minimum occurred in the afternoon or evening. These results suggest that available energy was primarily consumed by latent heating in the afternoon and by sensible heating in the morning. From spring (April) to winter (January), the diurnal amplitude of LE decreased gradually, while changes in the diurnal amplitude of Hs varied only slightly. R_n exhibited obvious differences between daytime and nighttime periods, with positive values during daytime and negative values during nighttime. The loss of R_n occurred mainly via LE throughout the year (except for winter). R_n might be influenced by wind in spring and autumn and by solar radiation in summer.

As shown in Figure 7a [Figure 7: see original paper], the daily total evaporation rate (ET) peaked in summer. From 1 April 2012 to 31 March 2013, the maximum ET was 11.74 mm/d in June, and the mean ET was approximately 4.0 mm/d. Furthermore, the average ET was approximately 0.6 mm/d during the frozen period (December and January) and 5.0 mm/d during the open-water period (February–November). Approximately 40% and 60% of evaporation occurred during nighttime and daytime, respectively. Moreover, ET generally decreased throughout late summer and autumn, maintained relatively low values from December to January, and increased from March to June. The average ET was lower in July than in June and August. It should be noted that the maximum ET did not occur in July, despite the highest temperatures occurring in this month. The cumulative evaporation was 1445 mm from 1 April 2012 to 31 March 2013 (Fig. 7b [Figure 7: see original paper]), which was 10 times larger than the accumulated precipitation over the same period.

4.1 Radiation and Heat Fluxes

Annual variations in DR were largely affected by solar elevation and weather conditions, including reflection, scattering, and absorption due to ambient atmospheric conditions. Monthly variations in DR resulted from changes in the sun' s azimuthal angle, while daily variations were due to cloud effects. UR was determined mainly by underlying surface conditions, particularly albedo. The studied lake was frozen in winter (January), and albedo was higher in winter than in other seasons; therefore, UR was relatively large in winter (Fig. 4c). The diurnal curve of albedo was not symmetric, possibly due to highly variable sand hills surrounding the lake that tended to block incident UR. The monthly average albedo was 0.13 from March to November; however, it was 0.21 from December to February, which is greater than the value (i.e., 0.07) reported for a water reservoir in a semi-arid region (Gallego-Elvira et al., 2010) but smaller than values in alpine meadows or savannas (Beringer and Tapper, 2002; Al-Riahi et al., 2003; Berbert and Costa, 2003; Zhang et al., 2010). However, the albedos obtained in this study are similar to average values for the Earth and

Northern Hemisphere (Gupta et al., 1999), suggesting that energy loss from the lake ecosystem via radiation may be lower than from other ecosystems due to relatively low albedo.

The exchange of longwave radiation between the atmosphere and lake surface is controlled primarily by surface features, clouds, atmospheric temperature, and atmospheric water vapor pressure. Over the studied lake, water vapor pressure varied from 0.1 to 2.0 kPa (Fig. 8 [Figure 8: see original paper]). However, previous studies reported that water vapor pressure in most ecosystems typically varies from 2.0 to 5.0 kPa (Kellner, 2001; Hunt et al., 2002; Wever et al., 2002). The water vapor pressure in this desert lake ecosystem was slightly higher than in deserts (ranging from 0.01 to 1.5 kPa) but much smaller than in most other ecosystems. Due to low ambient water vapor pressure and relatively few clouds, DLR over the studied lake was obviously lower than the value given by NASA for the Earth as a whole (Gupta et al., 1999).

The diurnal amplitude of LE was smaller in winter and larger in spring, indicating that ice cover effectively restricted heat exchange between air and water. Simultaneously, ice cover significantly increased surface albedo and reduced absorbed incident shortwave radiation.

4.2 Driving Forces Behind Heat Fluxes

Turbulent exchanges of water vapor and heat flux over lakes are affected by the surrounding environment. Heat flux over a lake surface is determined mainly by ambient temperature, turbulent mixing intensity, and water-air vapor pressure deficit (Hostetler and Bartlein, 1990; Bonan, 1995; Liu et al., 2009; Nordbo et al., 2011). Air at the interface between the lake and overlying atmosphere is always saturated, and the saturation pressure is a function of lake surface temperature at the interface (Hostetler and Bartlein, 1990). Meanwhile, several other parameterizations have been proposed, especially for lake evaporation (Rosenberry et al., 2007).

After calculating the relationship between Hs and LE while accounting for several environmental factors (Fig. 9 [Figure 9: see original paper]), we reached conclusions similar to those of previous studies. Specifically, there were obvious correspondences between Hs and water-air temperature difference (Nordbo et al., 2011; Liu et al., 2015), between Hs and wind speed × water-air temperature difference (Blanken et al., 2000, 2003; Liu et al., 2009; Nordbo et al., 2011), between LE and wind speed (Liu et al., 2009), and between LE and wind speed × water-air vapor pressure deficit (Blanken et al., 2003; Liu et al., 2009; Nordbo et al., 2011). According to Blanken et al. (2000), LE could be better described by wind speed × water-air vapor pressure deficit than by water-air vapor pressure deficit alone at Great Slave Lake. Considering wind' s influence on latent heat and evaporation, we analyzed correlations of LE with wind speed and wind speed × water-air vapor pressure deficit. We then performed correlation analyses of Hs with water-air temperature difference, Hs with wind speed × water-air temperature difference, LE with wind speed, and LE with wind speed × water-air vapor pressure deficit for further explanation (Fig. 9). Our results suggest that water-air temperature difference explained 45% of the variation in Hs, while the product of water-air temperature difference and wind speed explained 55% of Hs variability. In contrast, about 49% of LE variation could be explained by changes in wind speed, and 48% was explained by the product of water-air vapor pressure deficit and wind speed. Our results were similar to findings of Nordbo et al. (2011) but opposite to conclusions of Liu et al. (2009). The reason for this difference is that the lake in Liu et al. (2009) is located in a humid coastal area, whereas the lake in this study is situated in an arid desert. Complex terrain conditions can produce advection (Kaimal and Finnigan, 1994). According to our observations, the atmosphere over the lake exhibited a pronounced inversion layer with lake-breeze circulations, implying that horizontal transport of latent heat cannot be ignored in arid regions. Meanwhile, salinity could affect heat exchange between lake and atmosphere, while groundwater recharge (temperature of 18°C during the study period) supplies a large amount of heat. Therefore, LE and Hs exhibited weak relationships with environmental factors ($R^2 < 0.50$).

Several previous studies have shown that heat fluxes are sensitive to R_n, vapor pressure deficit of the atmosphere-lake system, and aerodynamic conductance (Williams et al., 1998; Malhi et al., 2002; Vourlitis et al., 2002; Kumagai et al., 2005; Loescher et al., 2005; Nordbo et al., 2011; Pan et al., 2017), whereas other studies have indicated that measured LE values are not sensitive to these environmental factors (Blanken et al., 2000; Loescher et al., 2005; Liu et al., 2009). These differences are due to the presence of vegetated surfaces where stomatal conductance control is high.

4.3 Comparison of Heat Fluxes Between Lake and Land Surface

The variations in Hs and LE observed by the EC system on the nearby land surface (Hu et al., 2015) differed from those observed on the lake surface. On the land surface, Hs is strong and LE is weak (Hu et al., 2015), whereas on the lake surface, Hs is weak and LE is strong (Fig. 5). There is a negative correlation between changes in Hs and LE. In the desert hinterland, LE could become negative at times even over the lake, indicating downward latent heat transport. This phenomenon has also been observed in other deserts (Wang and Mitsuta, 1990, 1992; Harazono et al., 1992). The reason may be that air flows across the dry land by advection, and the water vapor flux is directed downward on the land, implying downward water vapor delivery. On the land surface, there were obvious changes in Hs during the year (Hu et al., 2015). Specifically, from March to November, Hs was mainly positive, meaning heat was transferred to the atmosphere. However, from November to March, Hs

changed dramatically, with numerous negative values occurring, indicating heat transfer from the atmosphere to the land surface. On the lake surface, LE was occasionally negative, implying downward water vapor delivery.

4.4 Evaporation

The ET values observed on Yindeertu Lake in the Badain Jaran Desert exceeded those observed on high-latitude lakes (Schertzer et al., 2000; Rouse et al., 2003; Oswald and Rouse, 2004) but were similar to those measured in semi-arid reservoirs (Gallego-Elvira et al., 2010). The average ET was 0.6 mm/d during the frozen period for Yindeertu Lake, which is smaller than previously simulated ET (1.3–1.4 mm/d) for Lahontan Lake (Hostetler, 1991). This difference might be due to similar latitudes but different longitudes between the two lakes. Furthermore, the atmosphere over the lake in our study was unstable at night, resulting in strong lake-breeze circulation that increased evaporation (accounting for 40% of total evaporation).

Because net solar radiation and water-air vapor pressure deficit are large in the desert and coupled with dry air, evaporation is strong in desert areas (Zhu et al., 2012). However, ET from the lake surface measured in this study is much smaller than estimates by Chen et al. (2004) and Gates et al. (2008a, b). Note that net evaporation is 10 times larger than precipitation in the study area, indicating that the lake is supplied mainly by groundwater during the study period.

5 Conclusions

This study measured radiation components, evaporation rates, and heat fluxes using an EC system from March 2012 to March 2013 over a lake in the Badain Jaran Desert, China. Both shortwave and longwave radiation exhibited seasonal dynamics and large fluctuations. LE exhibited both seasonal variations and diurnal dynamics. LE was mainly affected by changes in wind speed, while Hs was primarily affected by the product of water-air temperature difference and wind speed. The diurnal LE and Hs were negatively correlated during the open-water period, and the effective energy of the lake was mainly directed into LE. Generally, LE was the dominant heat flux throughout the year. The average turbulent heat flux over the lake surface was greater than the net solar radiation received by the lake, indicating that other energy supplies to the lake cannot be ignored. LE showed negative values occasionally both on the land surface and lake surface, implying downward water vapor delivery. The cumulative evaporation was 1445 mm/a, and the mean evaporation rate was about 4.0 mm/d. The total evaporation was 10 times greater than the total precipitation, indicating that the lake was supplied mainly by groundwater during the study period. Continuous and long-term EC measurements will be needed to enable interannual analysis in the future.

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Note: Figure translations are in progress. See original paper for figures.