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Variability in cold front activities modulating cool-season evaporation from a southern inland water in the USA

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Abstract

Understanding seasonal variations in the evaporation of inland waters (e.g., lakes and reservoirs) is important for water resource management as well as the prediction of the hydrological cycles in response to climate change. We analyzed eddy covariance-based evaporation measurements from the Ross Barnett Reservoir (32°26'N, 90°02'W; which is always ice-free) in central Mississippi during the cool months (i.e., September–March) of 2007 and 2008, and found that the variability in cold front activities (i.e., passages of cold fronts and cold/dry air masses behind cold fronts) played an important role in modulating the exchange of sensible (H) and latent (λE) heat fluxes. Our analysis showed that 2007's warmer cool season had smaller mean H and λE than 2008's cooler cool season. This implies that the warmer cool season did not accelerate evaporation and heat exchange between the water surface and the atmosphere. Instead, more frequent cold fronts and longer periods of cold/dry air masses behind the cold fronts in 2008 resulted in overall larger H and λE as compared with 2007, this primarily taking the form of sporadic short-term rapid 'pulses' of H and λE losses from the water's surface. These results suggest that future climate-induced changes in frequency of cold fronts and the meteorological properties of the air masses behind cold fronts (e.g., wind speeds, temperature, and humidity), rather than other factors of climate change, would produce significant variations in the water surface's energy fluxes and subsequent evaporation rates.

Keywords: lake/reservoir evaporation, eddy covariance fluxes, cold fronts, the surface energy budget

1. Introduction

Evaporation from inland fresh water surfaces (lakes, reservoirs, swamps, wetlands, etc) is an important water loss from local catchments (Rouse *et al* 2005). Though land surface evaporation is expected to increase as a result of rising air temperature (Huntington 2006), quantifying changes in

evaporation from open freshwater areas in response to climate warming remains both a practical and theoretical challenge. During the wintertime, evaporative water losses from ice-free lakes are considerably large (Blanken *et al* 2000), thus it is important to investigate environmental controls on such losses for this season (Liu *et al* 2009).

It is well documented that water surface evaporation rates (or the equivalent latent heat flux, λE) are determined by vapor

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pressure differences between the water–air interface and the air above as well as by the turbulent mixing intensity (Hostetler and Bartlein 1990, Bonan 1995). Saturation usually occurs at the water–air interface and it is described by saturation vapor pressure, which is a function of the water surface temperature (Hostetler and Bartlein 1990). Similarly, sensible heat flux (H) is dependent upon the temperature difference between the water surface and the air above as well as on the turbulent mixing intensity (Hostetler and Bartlein 1990, Bonan 1995). Since water has a high specific heat capacity, water surface temperatures in large bodies/volumes of water show small diurnal variations in surface temperature, and even small sub-seasonal variations, as compared with the surrounding land surface temperatures. This leads to small changes in saturation vapor pressure in the water–air interface. As a consequence, the turbulent exchange of heat and water vapor between the water surface and the atmosphere is largely controlled by changes in over-water meteorological properties (Oswald and Rouse 2004). For instance, cold and dry air masses promote H and λE exchange between the water surface and the atmosphere through increasing the vertical temperature and humidity gradients in the atmospheric surface layer (ASL), while warm and humid air masses suppress and even dampen H and λE exchange through decreasing or inverting the vertical temperature and humidity gradients in the ASL (Rouse *et al* 2005, Liu *et al* 2009). It is noted that net radiation plays a minor role in influencing H and λE on a daily and even sub-seasonal basis (Rouse *et al* 2003, Lenters *et al* 2005, Liu *et al* 2009).

Previous studies have indicated that incursions of synoptic weather systems (e.g., cold fronts) bring in air masses with different meteorological properties and cause dramatic changes in wind speed and direction, temperature, humidity, and atmospheric pressure over a region (Lenters *et al* 2005, Liu *et al* 2009). Consequently, temporal variations in over-water meteorological properties as a result of the passage of synoptic weather systems over a region produce large variations in atmospheric forcings for water–atmosphere interactions and influence the water surface energy exchange (Blanken *et al* 2000, 2003, 2008, Schertzer *et al* 2003, Rouse *et al* 2003, Lenters *et al* 2005, Liu *et al* 2009). High-wind events with cold, dry air masses largely enhance turbulent mixing, resulting in episodic evaporation pulses or λE pulses (Blanken *et al* 2000, Liu *et al* 2009), as well as H pulses (Liu *et al* 2009). Following Blanken *et al* (2003), we define a ‘pulse’ as occurring when the 24 h mean λE or H is at least 1.5 times the value of the 10 day running mean. Entrainments of warm, dry air in the atmosphere down to the lake ASL also lead to very quick evaporation bursts, even in the stable ASL (Blanken *et al* 2003). These H and λE pulses contribute about 45–65% of the total annual evaporation and about 30% of the total H , although they comprise only 24–37% of the observation period (Blanken *et al* 2000, Lenters *et al* 2005, Liu *et al* 2009). These H and λE pulses usually occur almost simultaneously (Liu *et al* 2009). When dominated by southerly winds, Mississippi also experiences high-wind events with warm and humid air masses from the Gulf of Mexico, leading to minimal evaporation and a downward sensible heat transfer

(Liu *et al* 2009). These previous studies provide evidence that synoptic weather activities have great impacts on H and λE over inland waters. It remains unclear, however, how changes in synoptic weather activities in response to climate change affect both temporal variability and the magnitude of evaporation from inland waters.

Here, we provide direct over-water eddy covariance (EC) measurements of the surface energy budget over a large reservoir in central Mississippi during the cool seasons (i.e., from September to March) of 2007 and 2008 (the water is always ice-free during the winter). September is included in our analysis because several strong cold fronts passed over this region during this month in both years. This study adds to the few that have already used the EC technique for longer-term measurements made directly over large inland waters (Blanken *et al* 2000, Vesala *et al* 2006, Liu *et al* 2009, Nordbo *et al* 2011). In the cool seasons, this region is subject to the frequent incursions of cold fronts (Liu *et al* 2009). After the passage of these cold fronts, the cold and dry air masses with high-wind speeds immediately behind the cold fronts influence this region for up to several days (hereafter referred to as high-wind events). Our objective is to quantify the relative contributions of changes in these high-wind events (which are quantified in terms of their frequency, duration, and intensity) and changes in weather conditions (which exclude days with high-wind events) to magnitudes of the surface energy balance components for each cool season. Differences in the meteorological variables and surface energy budgets (e.g., H and λE) between the cool seasons of 2007 and 2008 are then compared to examine how changes in the frequency, duration, intensity, and meteorological properties of high-wind events behind cold fronts in these two seasons affect evaporation and the surface energy exchange over a southern reservoir in Mississippi, USA.

2. Site description, instruments, and methods

We used EC-based data measured during the cool seasons of 2007 and 2008 (September–March) from a tower installed near the center of the Ross Barnett Reservoir (hereafter referred to as ‘the Reservoir’; 32°26′N, 90°02′W; 117.5 m a.s.l.), with a mean depth of 5 m and surface area of approximately 134 km² in central Mississippi. The EC system was mounted at a height of approximately 4 m above the water’s surface on a 5 m-tall aluminum tower (Climatech Corp.) that was positioned on a stationary wooden square platform anchored to the bed of the Reservoir. The distance from the tower to the shore ranged from 2 to about 14 km and the water depth around the tower was approximately 5 m.

The EC system consisted of a three-dimensional sonic anemometer (CSAT3, Campbell Scientific Inc.) and an open-path CO₂/H₂O infrared gas analyzer (IRGA; LI 7500, LI-COR, Inc.). Sensor signals from the EC system were sampled at 10 Hz and recorded with a datalogger (model CR5000, Campbell Scientific Inc.). The detailed post-field data processing and quality control procedures used here to H and λE were documented in Liu *et al* (2009). Considering all

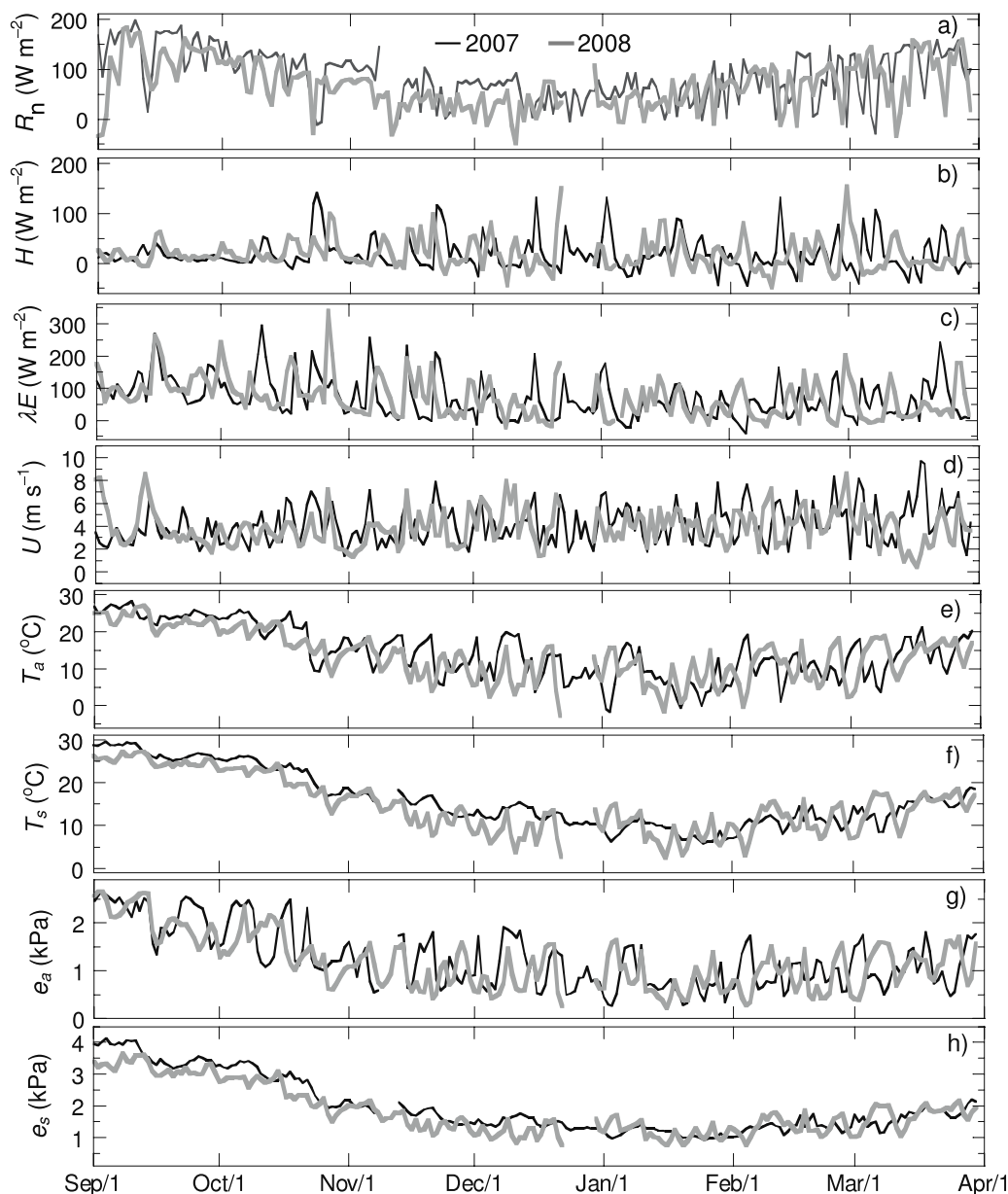


Figure 1. Daily (24 h) mean time-series data of (a) net radiation (R_n), (b) sensible heat flux (H), (c) latent heat flux (λE), (d) wind speeds (U), (e) air temperature (T_a), (f) water surface temperature (T_s), (g) vapor pressure in the overlying air (e_a), and (h) saturation vapor pressure in the water–air interface (e_s) in the 2007 and 2008 cool seasons. Some data gaps were caused by instrument failure and repair.

missing and rejected data, the availability of the 30 min mean flux observations was 92% (93%), 97% (95%), 95% (93%), 93% (88%), 96% (94%), 89% (91%), and 94% (95%) for September, October, November, December, January, February, and March, respectively, in the 2007 (2008) cool seasons (see figure 1 for the average daily fluxes).

We also measured a variety of micrometeorological variables as 30 min averages of 1 s measurements. Net radiation (R_n) was measured with a net radiometer (model Q-7.1, Radiation and Energy Balance Systems (REBS), Inc.). Incoming solar radiation was measured with a silicon pyranometer (model LI-200, LI-COR, Inc.). Air temperature and relative humidity profiles were measured at four heights (i.e., about 1.90, 3.00, 4.00, and 5.46 m above

the water’s surface) using four temperature/humidity probes (model HMP45C, Vaisala, Inc.). Wind speed and direction were measured at the top of the tower (about 5 m above the water’s surface; model 03001, RM Young, Inc.) and additional three 3-cup anemometers (model 03101, RM Young, Inc.) were mounted at the same heights as the HMP45C probes. In this study, we used the air temperature (T_a), atmospheric vapor pressure (e_a), and wind speed (U) measured at 5.46 m above the water’s surface. Water surface temperatures (T_s) were measured with an infrared thermometer (model IRR-P, Apogee, Inc.) mounted on a 1.5 m-long horizontal boom about 1 m above the water’s surface. This instrument is reported to have an accuracy of $\pm 0.2^\circ\text{C}$ for temperature ranges from -10 to $+65^\circ\text{C}$. Vapor pressure at the water’s surface

Table 1. Monthly averaged components of the surface energy fluxes and meteorological variables in the 2007 and 2008 cool seasons. (Note: abbreviations are as follows: R_n : net radiation ($W m^{-2}$), H_m : the mean sensible heat flux ($W m^{-2}$), λE_m : the mean latent heat flux ($W m^{-2}$), U : wind speed ($m s^{-1}$), T_a : air temperature ($^{\circ}C$), T_s : water surface temperature ($^{\circ}C$), e_a : vapor pressure in the atmosphere (kPa), e_s : saturation vapor pressure at the water–air interface (kPa), S_o : incoming solar radiation ($W m^{-2}$).

	R_n ($W m^{-2}$)		H_m ($W m^{-2}$)		λE_m ($W m^{-2}$)		U ($m s^{-1}$)		T_a ($^{\circ}C$)		T_s ($^{\circ}C$)		e_a (kPa)		e_s (kPa)		S_o ($W m^{-2}$)	
	2007	2008	2007	2008	2007	2008	2007	2008	2007	2008	2007	2008	2007	2008	2007	2008	2007	2008
Sept.	148.8	113.4	14.2	16.6	113.2	115.8	3.4	4.1	24.9	23.3	27.1	25.1	2.21	2.11	3.62	3.22	198.4	181.3
Oct.	104.6	81.8	24.1	27.4	102.3	104.2	4.0	3.4	19.9	17.9	22.6	20.4	1.65	1.46	2.79	2.45	176.1	172.1
Nov.	69.8	39.9	21.3	20.8	67.1	65.7	3.8	3.4	13.7	12.1	15.3	13.5	1.09	1.01	1.75	1.59	135.0	119.8
Dec.	51.9	26.9	18.2	20.3	44.2	51.2	4.1	4.6	11.4	8.2	12.3	8.8	1.07	0.68	1.44	1.16	92.3	87.1
Jan.	51.8	37.7	23.1	19.7	52.3	51.5	4.4	4.1	6.9	7.8	8.6	8.6	0.75	0.79	1.11	1.15	103.6	107.1
Feb.	77.6	73.4	14.1	8.2	49.1	46.3	4.6	4.3	10.4	10.7	10.8	11	0.94	0.86	1.31	1.36	129.9	149.4
March	111.0	95.7	12.2	18.7	57.9	53.8	4.9	3.8	14.1	13.4	14.5	14.1	1.12	1.12	1.68	1.66	197.4	156.6
Mean	88.0	67.0	18.2	18.8	69.4	69.8	4.2	4.0	14.5	13.3	15.9	14.5	1.26	1.15	1.96	1.80	147.5	139.1

(e_s) was calculated as the saturation vapor pressure at the infrared-determined water surface temperature (T_s) (Blanken *et al* 2000). Eight water temperature probes, attached to a buoy under the tower platform, were placed at 0.10, 0.25, 0.5, 1.0, 1.5, 2.5, 3.5, and 4.5 m depths below the water’s surface (model 107-L, Campbell Scientific Inc.). Rainfall totals were measured at half-hour intervals with an automated tipping-bucket rain gauge (model TE525, Texas Instruments, Inc.). Sensor signals from all slow-response sensors were also recorded by the CR5000 datalogger. All instruments were powered by two 65 W solar panels (model SP65, Campbell Scientific Inc.) and six deep-cycle marine batteries.

3. Results

3.1. General characteristics of climatic variables and surface energy fluxes

The overall mean T_a for the 2007 cool season was $14.5^{\circ}C$, $1.2^{\circ}C$ higher than that in 2008 (table 1). On average, the T_s in the 2007 cool season ($15.9^{\circ}C$) was also higher than that in 2008 ($14.5^{\circ}C$), partly due to the higher R_n in 2007 as compared to 2008 (table 1). In general, it was warmer in the 2007 cool season than in the 2008 cool season. The higher T_s in the 2007 cool season led to a higher e_s than in 2008 (1.96 versus 1.80 kPa; table 1). Though the total rainfall in the 2007 cool season was 467 mm, 248 mm less than in the 2008 cool season, the e_a was higher in the 2007 cool season (1.26 kPa) than in the 2008 cool season (1.15 kPa). Nevertheless, $T_s - T_a$ in the 2007 cool season was larger than that in 2008 and the two seasons had very close $e_s - e_a$ (table 1).

Though the 2007 cool season was warmer than the 2008 cool season, the warmer conditions did not accelerate evaporation from the Reservoir as compared with 2008. Instead, H_m (hereafter the subscript ‘m’ denotes the mean value for the whole season) and λE_m were more similar than expected in 2007 ($H_m = 18.2 W m^{-2}$ and $\lambda E_m = 69.4 W m^{-2}$) compared to those in 2008 ($H_m = 18.8 W m^{-2}$ and $\lambda E_m = 69.8 W m^{-2}$) (table 1). Our data in table 1 also indicate that higher monthly mean air temperatures did not necessarily lead to a larger monthly mean λE in these two cool seasons. What, then, were the mechanisms

responsible for this correspondence between variations in the atmospheric forcings and fluxes in the two cool seasons? We hypothesize that temporal variability in high-wind events with different intensities, durations, and frequencies, in addition to the meteorological properties associated with the passage of cold fronts in the two cool seasons which exerted different atmospheric forcings, modulated heat and water vapor transfer in ways that led to the slightly larger H_m and λE_m in the 2008 cool season compared to those in the 2007 cool season.

3.2. Influence of variability of cold front activities on the surface energy fluxes

When the 24 h mean time series of the fluxes were examined, H and λE pulses were found throughout the cool seasons (figure 1). These pulses were identified when the one-day mean H (or λE) was at least 1.5 times the magnitude of the 10 day running mean. It was found that these H and λE pulses during these ‘pulse days’ were accompanied by dramatically increased wind speeds, changes in wind direction, and significant drops in temperature and humidity (i.e., high-wind events which had the same duration as defined by the H and λE pulses). Following the method described in Liu *et al* (2009), we collected and analyzed daily synoptic charts that clearly show the status of cold front passages over this region. By comparing synoptic charts from consecutive days, we were able to determine the date on which a cold front was passing over the region (chart source: <http://www.hpc.ncep.noaa.gov/dwm/dwm.shtml>). The timing of the cold front’s passage could then be estimated by examining the time-series data of 30 min mean meteorological variables and H and λE fluxes. We present one typical case in figure 2, which shows the influence of a high-wind event behind a cold front on meteorological variables and H and λE fluxes. Based on synoptic charts we collected (not shown here), a cold front passed over the site on 27 October 2008. The time-series data indicated that the cold front arrived at the site at approximately 0000 LT (local time) on that day, leading to a high-wind event with a rapid increase in wind speed, a dramatic decrease in temperature, and a large drop in water vapor pressure (figure 2). This high-wind event lasted approximately 49 h. The wind speeds increased from about 3.0 to $10.0 m s^{-1}$, with a change in wind

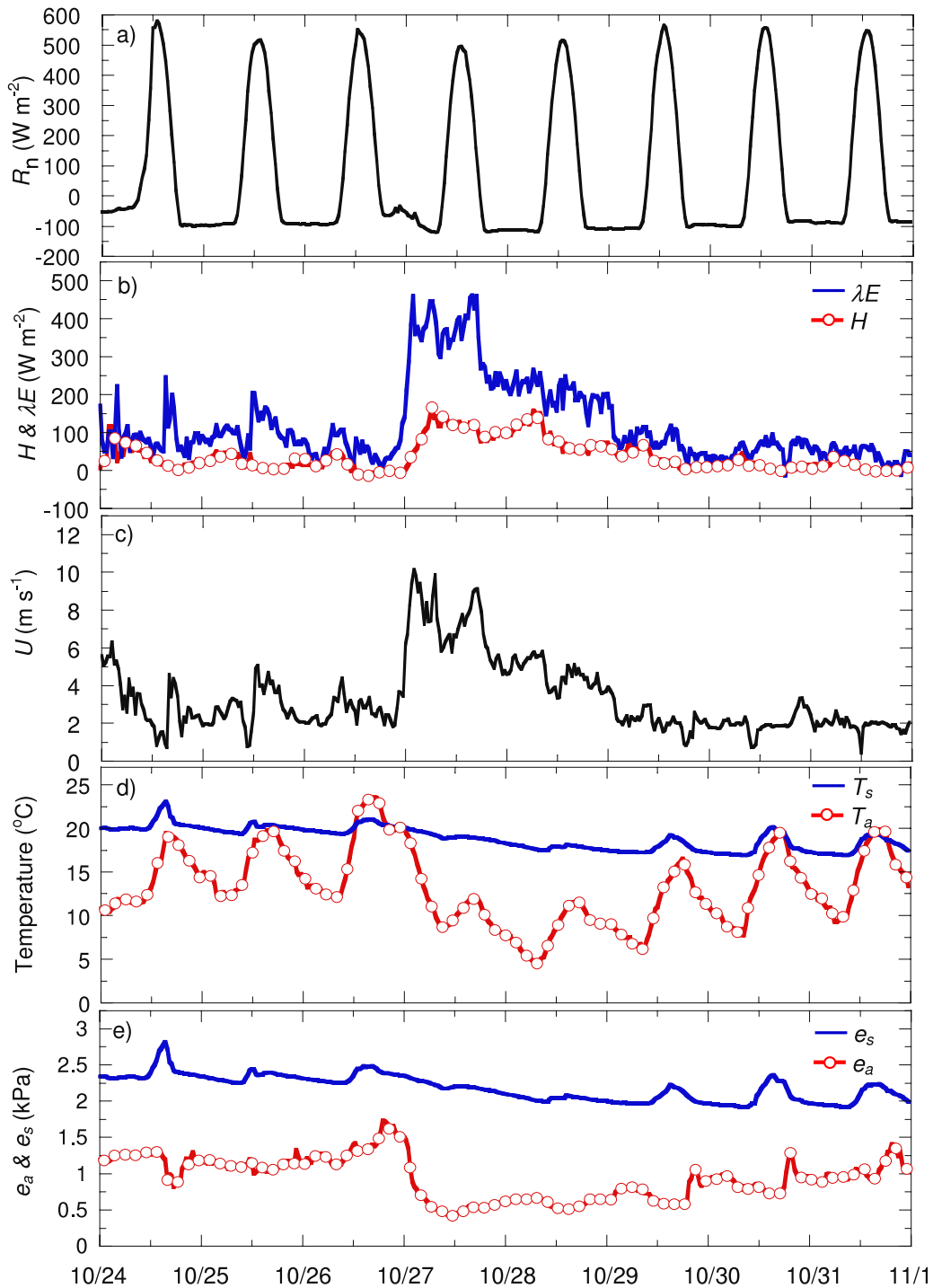


Figure 2. Time-series data of 30 min mean (a) net radiation (R_n), (b) sensible (H) and latent (λE) heat flux, (c) wind speed (U), (d) air temperature (T_a) and water surface temperature (T_s), and vapor pressure for the water–air interface (e_s) and the overlying air (e_a) from 24 to 31 October 2008. The data represent a typical H and λE pulse as a result of a high-wind event behind a cold front that passed over the site at about 0000 LT on 27 October 2008.

direction from southwesterly to northeasterly. Meanwhile, the air temperature dropped from about 19.4°C at 0000 LT on 27 October to 4.5°C at 0730 LT on 28 October, and the water vapor pressure decreased from 1.5 to 0.5 kPa. It is noted that the water surface temperature and surface vapor pressure did not show dramatic changes during this period. As a consequence, high-wind speeds enhanced mechanical

turbulent mixing, leading to a dramatic increase in friction velocity (u_*) up to 0.7 m s⁻¹. Cold air masses promoted thermally generated turbulence through increasing $T_s - T_a$, and the passage of dry air masses led to increased $e_s - e_a$. The combination of enhanced turbulent mixing and increased temperature and humidity gradients in the ASL produced the H and λE pulses (figure 2). We made the same analysis

Table 2. Half-hourly mean meteorological variables during the periods of non-pulse (NP) days and the periods of pulse days (P) in the 2007 and 2008 cool seasons.

Period	R_n ($W m^{-2}$)	U ($m s^{-1}$)	T_a ($^{\circ}C$)	T_s ($^{\circ}C$)	e_a (kPa)	e_s (kPa)	$T_s - T_a$ ($^{\circ}C$)	$e_s - e_a$ (kPa)
2007-NP	92.0	3.8	16.0	16.3	1.43	2.01	0.3	0.58
2007-P	81.2	5.1	10.7	15.4	0.84	1.94	4.7	1.10
2008-NP	75.8	3.5	15.4	16.0	1.35	1.95	0.6	0.60
2008-P	51.6	4.8	9.8	12.3	0.85	1.55	2.5	0.70

Table 3. Half-hourly mean fluxes of sensible and latent heat during the periods of non-pulse days and the periods of pulse days in the 2007 and 2008 cool seasons. (Note: abbreviations are as follows: H_{NP} : sensible heat fluxes in the non-pulse days ($W m^{-2}$), H_P : sensible heat flux on the pulse days ($W m^{-2}$), λE_{NP} : latent heat flux on the non-pulse days ($W m^{-2}$), λE_P : latent heat flux on the pulse days ($W m^{-2}$), ‘Days’ denotes respective non-pulse days and pulse days for each cool season (each season has 212 days). ‘Flux \times days’ denotes the integrated fluxes of sensible heat (or latent heat) during the periods of non-pulse days and the periods of pulse days if multiplied by 84400 ($60 s \times 60 min \times 24 h$).

	H_{NP}		H_P		λE_{NP}		λE_P	
	2007	2008	2007	2008	2007	2008	2007	2008
Flux	3.9	3.1	55.6	54.4	41.5	40.0	141.8	137.8
Days	153	146	59	65	153	146	59	65
Flux \times days	576.7	452.6	3280.4	3536.0	6349.5	5840.0	8363.3	8957.0

for all synoptic charts and time-series data of 30 min mean meteorological variables and fluxes for each cool season. We identified a total number of 22 cold fronts with high-wind events covering a period of 59 days in the 2007 cool season, compared to a total of 30 cold fronts with high-wind events covering a period of 64 days in 2008.

To quantify the influence of high-wind events that are associated with cold fronts on surface energy fluxes, we separated the half-hourly flux data and the corresponding meteorological data into two groups: one group with H and λE pulses driven by high-wind events on pulse days and another group without H and λE pulses driven by non-pulse weather conditions on ‘non-pulse days’ (tables 2 and 3). We calculated that H on the non-pulse days (i.e., H_{NP} in table 3) was 3.9 and 3.1 $W m^{-2}$ in the 2007 and 2008 cool seasons, respectively, while H on the pulse days (i.e., H_P in table 3) was 55.6 and 54.4 $W m^{-2}$ in the 2007 and 2008 cool seasons, respectively. λE on the non-pulse days (i.e., λE_{NP} in table 3) was 41.5 and 40.0 $W m^{-2}$ in the 2007 and 2008 cool seasons, respectively, while λE on the pulse days (i.e., λE_P in table 3) was 141.8 and 137.8 $W m^{-2}$, respectively. Apparently, H_{NP} and λE_{NP} in the 2007 cool season were greater than those in 2008, indicating that the warmer non-pulse weather conditions in the 2007 cool season did lead to enhanced exchange of heat and water vapor (i.e., H_{NP} and λE_{NP}) as compared with those in the 2008 cool season (table 3). These results highlighted the notion that warm non-pulse weather conditions, as compared with cool non-pulse weather conditions, would accelerate evaporation. It is also noted that H_P and λE_P on the pulse days of the 2007 cool season were larger than those in the 2008 cool season (table 3). Since H and λE pulses were the direct consequence of high-wind events that promoted heat and water vapor exchange by enhancing both mechanically- and thermally generated turbulence, the magnitudes of H_P and λE_P during periods of high-wind events behind cold fronts were thus indicators of the intensity of the high-wind events in terms of their impacts on turbulent exchange. Therefore, the larger

H_P and λE_P in the 2007 cool season (larger by 25.8% and 3.8%, respectively) as compared with those in 2008 implies (as confirmed by table 2) that high-wind events in the 2007 cool season were on average stronger than those in the 2008 cool season in terms of wind speed, temperature difference, and vapor pressure difference. Our calculations indicated that the mean wind speed, temperature difference, and vapor pressure difference were 5.1 $m s^{-1}$, 4.7 $^{\circ}C$, and 1.1 kPa, respectively, during the entire pulse period of the 2007 cool season and 4.8 $m s^{-1}$, 2.5 $^{\circ}C$, and 0.7 kPa, respectively, during the entire pulse period of the 2008 cool season (table 2). What, then, was the cause of the higher H_m and λE_m in the 2008 cool season, since H_P and λE_P , as well as H_{NP} and λE_{NP} , were smaller in the 2008 cool season than in the 2007 cool season?

Aside from the intensity of high-wind events having an impact on the magnitudes of H_P and λE_P as discussed above, the duration and frequency of high-wind events which are associated with the passage of cold fronts should also be accounted for in quantifying their contributions to H and λE . The total number of pulse days in the 2008 cool season (65 days) was more than that in the 2007 cool season (59 days) as a result of the increased number of cold front incursions in the 2008 cool season (30) as compared with the 2007 cool season (22) (table 3). Therefore, H_m (and λE_m) for the entire season, as shown in table 1, can be estimated using table 3 as the weighted mean of H_{NP} and H_P (and λE_{NP} and λE_P) over the non-pulse days and the pulse days, respectively. Specifically, H_m and λE_m for the 2007 cool season were 18.2 $W m^{-2}$ (i.e., $(3.9 W m^{-2} \times 153 \text{ days} + 55.6 W m^{-2} \times 59 \text{ days})/212 \text{ days}$) and 69.4 $W m^{-2}$ (i.e., $(41.5 W m^{-2} \times 153 \text{ days} + 141.8 W m^{-2} \times 59 \text{ days})/212 \text{ days}$), respectively; while H_m and λE_m for the 2008 cool season were 18.8 $W m^{-2}$ (i.e., $(3.1 W m^{-2} \times 146 \text{ days} + 54.4 W m^{-2} \times 65 \text{ days})/212 \text{ days}$) and 69.8 $W m^{-2}$ (i.e., $(40.0 W m^{-2} \times 146 \text{ days} + 137.8 W m^{-2} \times 65 \text{ days})/212 \text{ days}$), respectively. The striking consequence is that the lengthening of high-wind events (i.e., the increased number of pulse days)

in the 2008 cool season led to the larger average H_m and λE_m in that season.

It should be mentioned that we also applied the above analysis to monthly data in order to quantify the relative contributions of high-wind events and non-pulse weather conditions to the surface energy fluxes for each month. Our results indicated that the high-wind events modulated the monthly mean surface energy fluxes, even for the months with lower monthly mean air temperatures (January and February) in the 2007 cool season (table 1). It is noted that the year-to-year differences in the surface energy budget observed here are small, but consistent. These small differences may approach the accuracy ranges of the currently available micrometeorological and eddy covariance instruments (Mauder *et al* 2007). However, the use of the same eddy covariance systems and micrometeorological instruments in this study for two years were able to offset significantly the systematic errors that the eddy covariance system might have. Therefore, our results are considered to be robust for such comparison analyses.

4. Discussion and conclusions

In cool seasons, cold fronts are common synoptic weather systems that occur every 5–10 days at our study location. High-wind events behind cold fronts persist for a few days, causing dramatic changes in meteorological conditions (e.g., wind speed, temperature, and humidity) and thus in the atmospheric forcings for land/water–air interactions over the regions they pass. In our study, high-wind events promoted the turbulent exchange of sensible and latent heat both mechanically and thermally, leading to frequent occurrences of H and λE pulses that were, respectively, 15.9 and 3.4 times those on non-pulse days. On non-pulse days, the warmer non-pulse weather conditions (excluding the days with high-wind events) in the 2007 cool season caused larger H_{NP} and λE_{NP} than in the 2008 cool season. On pulse days, the H_P and λE_P in the 2007 cool season were also larger than those in the 2008 cool season. However, our results indicated that the lengthening of high-wind events (more pulse days) in the 2008 cool season as a result of the increased frequency and duration of high-wind events was the main cause of the overall greater average H_m and λE_m in the 2008 cool season compared to the 2007 cool season.

These results suggest that water surface energy fluxes in the cool season are strongly dependent upon cold front activities by altering the magnitude, frequency, and duration of H and λE pulses. Cyclone activities are believed to have changed in past decades in response to climate warming (McCabe *et al* 2001, Teng *et al* 2008). Future changes in cold front activities and the resulting high-wind events (e.g., intensity, frequency, and duration) due to climate warming would likely alter the water surface energy fluxes in the cool seasons. Our results also indicate that a warming climate does not necessarily lead to increased evaporation rates and increased sensible heat exchange between the water surface and the atmosphere in the cool season. The net effects of atmospheric forcings on water surface energy fluxes

are actually dependent upon the relative contributions of atmospheric forcings that are associated with high-wind events as well as those associated with non-pulse weather conditions that exclude high-wind events. H and λE pulses have been seen as an important process that has a significant impact on evaporation over northern high-latitude lakes in winter during ice-free periods (Blanken *et al* 2000). It is interesting to note that the southern reservoir in this study responds to the atmospheric forcings from both high-wind events associated with cold front activities and non-pulse weather conditions in the same ways as these deep, large northern lakes. Given a wide latitudinal gradient, it is stressed that these same exchange mechanisms reflect the generic responses of large bodies of water to atmospheric forcings. It should be mentioned, however, that the processes and results associated with water–air interactions which have been reviewed and provided here were obtained based on data measured over large water surfaces. Since small bodies of water have a lower heat capacity and respond to changes in atmospheric forcings more quickly than large, deep bodies of water, thus leading to different exchange processes over water, caution should be taken when applying these results to inland bodies of water with small volumes.

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