Journal of Great Lakes Research 37 (2011) 707–716

Contents lists available at SciVerse ScienceDirect

Journal of Great Lakes Research

journal homepage: www.elsevier.com/locate/jglr

# Evaporation from Lake Superior: 1. Physical controls and processes

Peter D. Blanken [a,*], Christopher Spence [b,1], Newell Hedstrom [b,2], John D. Lenters [c,3]

[a] Department of Geography and Program in Environmental Studies, University of Colorado, Boulder, CO 80309-0260, USA
[b] Environment Canada, Saskatoon, SK, Canada S7N 3H5
[c] School of Natural Resources, University of Nebraska-Lincoln, Lincoln, NE 68583-0987, USA

ARTICLE INFO

*Article history:*
Received 22 April 2011
Accepted 19 August 2011
Available online 2 October 2011

Communicated by Leon Boegman

*Index words:*
Energy balance
Evaporation
Great Lakes
Ice cover
Lake Superior
Water levels

ABSTRACT

The surface energy balance of Lake Superior was measured using the eddy covariance method at a remote, offshore site at 0.5-h intervals from June 2008 through November 2010. Pronounced seasonal patterns in the surface energy balance were observed, with a five-month delay between maximum summer net radiation and maximum winter latent and sensible heat fluxes. Late season (winter) evaporation and sensible heat losses from the lake typically occurred in two- to three-day-long events, and were associated with significant release of stored heat from the lake. The majority of the evaporative heat loss (70–88%) and sensible heat loss (97–99%) occurred between October and March, with 464 mm (2008–2009) and 645 mm (2009–2010) of evaporative water loss occurring over the water year starting October 1. Evaporation was proportional to the horizontal wind speed, inversely proportional to the ambient vapor pressure, and was well described by the ratio of wind speed to vapor pressure. This ratio remained relatively constant between the two water years, so the differences in evaporative water loss between years were largely associated with differences in lake surface conditions (e.g. water temperature, ice cover, and ice duration). Since late-season water temperature decline is driven by evaporative and sensible heat loss, the potential for a negative feedback mechanism between evaporation and ice cover is discussed.



## Introduction

Lake Superior has a surface area of 82,100 km$^2$, making it the largest freshwater lake in the world, by surface area (Kalff, 2002). Together with the other Great Lakes, the socioeconomic importance of this resource is immense for both the United States and Canada, directly influencing sectors including power production, navigation, industry, commercial operations, agriculture, and recreation (Hartmann, 1990). The economic impact of the Great Lakes is large; for example, spending on recreational boating in 2003 alone was $9.4 billion, and created 60,000 jobs (US Army Corps of Engineers, 2008). The importance of the Great Lakes to the regional meteorology and climate is significant for both countries, since changes in meteorological conditions and water quality often directly impact the economy of the region. Examples include changes in ice cover (Assel et al., 2003) which, in-turn, are linked to lake-enhanced snow events (Cordeira and Laird, 2008; Ellis and Johnson, 2004) affecting transportation; and decreases in water level through recent increases in evaporation (Hanrahan et al., 2010; Sellinger et al., 2008) affecting navigation and shore-line erosion.

Despite the significance of the Great Lakes, the boundary conditions at the air–lake interface that drive the surface meteorology and limnology through the surface energy balance are seldom directly measured, especially for large lakes such as Lake Superior. This is not because the surface energy balance is not recognized as being important, but rather is due to the limited availability of required measurements on the lake itself, and most hydroclimate models still require forcing data from direct measurements (e.g. Huang et al., 2011). As we show here, the magnitude and dynamics of the heat exchange between the air–lake interface are greatest during the winter when buoy-based meteorological measurements are not available. Therefore, much of the past research on the surface energy balance of Lake Superior and other large lakes has been based on empirical models (e.g. Derecki, 1981), buoy-based measurements during the open-water season (e.g. Laird and Kristovich, 2002), or more complex hydrodynamic models (e.g. Beletsky et al., 1999), aided by remotely-sensed data when available (e.g. Lofgren and Zhu, 2000; Nghiem and Leshkevich, 2007).

What is lacking from most previous studies is a direct measurement of the evaporative and sensible heat losses (i.e., more than just bulk aerodynamic relationships), especially during the winter months when these losses are at maximum. The value of such measurements is that they provide an understanding of the temporal patterns of the surface energy balance, and the physical controls of the processes

* Corresponding author. Tel.: +1 303 492 8310.
*E-mail addresses:* blanken@colorado.edu (P.D. Blanken), Chris.Spence@ec.gc.ca (C. Spence), Newell.Hedstrom@ec.gc.ca (N. Hedstrom), jlenters2@unl.edu (J.D. Lenters).
[1] Tel.: +1 306 975 6907.
[2] Tel.: +1 306 975 6049.
[3] Tel.: +1 402 304 0166.

0380-1330/$ – see front matter © 2011 International Association for Great Lakes Research. Published by Elsevier B.V. All rights reserved.
doi:10.1016/j.jglr.2011.08.009

involved. This understanding can then be used to verify, calibrate, and improve the predictive models. It is the purpose of this paper, therefore, to describe the annual patterns of direct measurements of the surface energy balance of Lake Superior, and to understand the first-order controls on the evaporative water loss. Although we describe direct measurements based on the eddy covariance method from a remote site 39-km offshore, our measurements represent an upwind distance of roughly 6 km. The controls on evaporation that we discuss may vary significantly on the spatial scale of this immense lake, therefore any efforts to model the evaporative process over a broad area requires knowledge of the spatial patterns in the atmospheric driving variables and lake temperature. The companion paper by Spence et al. (in press) deals with this aspect of the study.

There are several studies reporting that Lake Superior is responding to climate change. Schertzer and Rao (2009) provide a summary of the past and present state of the Lake; their review can be used as a baseline for climate change modeling or historical trend-analysis studies. Examples of recent studies describing the contemporary changing conditions include Bennington et al. (2010), who used a Reynolds-averaged hydrodynamic model to simulate and analyze circulation in Lake Superior from 1979 to 2006. They simulated increasing lake surface and above-lake air temperatures at a rate of 0.34 and 0.8 °C per decade, respectively, increasing wind speed (0.18 m $s^{-1}$ per decade), and a long-term decline in January–April ice cover (−866 $km^2$ per year). Austin and Colman (2007), on the other hand, found even larger increases in buoy-measured lake temperature over a similar time period, and at a rate that was higher than that of the above-lake air temperatures, rather than lower. Trend-analysis studies show that snowfall has increased dramatically in areas that are subject to lake-effect snowfall (Ellis and Johnson, 2004), although it has been noted that such trends should be treated with caution (Kunkel et al., 2007). Water levels have been decreasing at a rate of approximately 1 cm per year from 1970 to 2005 (Moitee and McBean, 2009), while at the same time experiencing shifts in the seasonal cycle (Argyilan and Forman, 2003; Lenters, 2001, 2004; Quinn, 2002). Given these and other changes, the purpose of this paper is to describe the annual surface energy balance of Lake Superior based on direct measurements, and to discuss how the physical processes regulating the surface energy balance may be affected by climate change.

## Materials and methods

Making year-long, over-water measurements on Lake Superior, the deepest (mean depth 148 m; maximum depth 406 m) and largest (surface area 82,100 $km^2$; volume 12,100 $km^3$) of the Great Lakes, is exceptionally difficult for logistical, cost, and safety reasons. Stannard Rock Light (47.183506, −87.22511), however, a historic lighthouse located on a shoal 39 km from the nearest shore (the Keweenaw Peninsula) provided an ideal location for this project. The now-automated lighthouse, completed in 1882, offers a stable, year-round platform for the meteorological and eddy covariance instruments, as well as good conditions for surface flux measurements, given its location (far from shore), its height (32.4 m above the mean water level), and the absence of any underlying land surface (i.e. island) to influence measurements. Although shallow water (~5 m deep) surround the immediate area, water depths within the source area for the flux measurements (~6 km upwind during the active evaporation season; see below) range from 150 to 300 m. The eddy covariance and supporting meteorological measurements reported here were made between June 12, 2008 and November 4, 2010.

From a mast fastened to the top of the lighthouse structure 32.4 m above the water surface, the turbulent fluxes of sensible ($H$) and latent ($\lambda E$) heat were calculated from 10-Hz measurements of the vertical wind speed ($w$) and the water vapor density ($q$). Wind speed was measured using a 3-D sonic anemometer (model CSAT-3, Campbell Scientific, Logan, UT), while water vapor density was measured using an open-path gas analyzer (model LI-7500, LI-COR Biosciences, Lincoln, NE) located 15 cm away and at the same height as the sonic anemometer. The statistics (means and covariances) of the high-frequency sampled data were collected at 30-min intervals using a datalogger (model CR3000, Campbell Scientific, Logan, UT), with DC power provided by eight 12-V deep-cycle marine batteries charged by six 80-W solar panels. The datalogger and batteries were located in a dry location inside the lighthouse. Post-processing of these data, including quality control, is described below.

On the same mast, slow 5-s samples of ancillary meteorological variables were also measured, with 30-min statistics collected on the same datalogger as the high-frequency data. Air temperature ($T_a$; °C) and humidity (vapor pressure, $e_a$; kPa) were measured with a shielded probe (model HMP45C, Vaisala, Helsinki), and barometric pressure ($P$; kPa) was measured with a pressure transducer located inside the electronics box of the LI-7500. The horizontal wind speed ($U$; m $s^{-1}$) and direction were measured with a wind vane (model 05103, RM Young, Traverse City, MI), in addition to horizontal wind speed and direction calculated from the sonic anemometer data. Water surface temperature ($T_o$) was measured with an infrared thermometer (model IRR-P, Apogee, Logan, UT). At times (29% of all the 0.5-h measurements), the measured $T_o$ was unreliable likely due to condensation or frost formation inside the instrument cavity. Water surface temperature was estimated during such times by solving the stability-corrected flux-gradient equation (Businger et al., 1971) using the measured sensible heat flux and two levels; the water surface and instrument heights. The incident short-wave ($S\downarrow$) and long-wave radiation ($L\downarrow$) were measured with radiometers (models SP Lite and CGR-4, respectively, Kipp and Zonen, Delft).

Lake-wide daily ice cover was estimated from the NOAA-GLERL Great Lakes Ice Atlas, which provides the percentage of a nominal 2.55 km × 2.55 km grid covered by ice based on a blend of observations from several data sources including ships, aircraft, and satellites (Assel, 2005). The daily ice cover percentages were also estimated within the ~6-km turbulent flux footprint of the eddy covariance instruments (described in the Theory/calculation section), using georeferenced ice charts provided by the Canadian Ice Service.

## Theory/calculation

The turbulent fluxes of latent and sensible heat (W $m^{-2}$) were calculated from the covariance of the high-frequency measurements of the vertical wind speed (m $s^{-1}$) with water vapor density (g $m^{-3}$) or virtual air temperature ($T$; °C), respectively:

$$\lambda E = \overline{\lambda w'q'} \tag{1}$$

$$H = \overline{\rho c w'T'} \tag{2}$$

where $\lambda$ is the air temperature-dependent latent heat of vaporization (J $g^1$), $\rho$ is the dry air density (g $m^{-3}$), $c$ is the pressure-dependent specific heat of dry air (J $g^{-1}$ °$C^{-1}$), and primes denote deviations from the 30-min non-overlapping, non-detrended means (overbars). Positive values indicate turbulent fluxes directed away from the surface; negative towards. The 30-min averaging period was chosen since we found that period gave good energy balance closure at an offshore site (~12 km) over the similarly large Great Slave Lake (Blanken et al., 2000), where intermittent high-frequency turbulent events played a significant role in evaporation (Blanken et al., 2003). Since our site was also far from shore where there was no sharp contrast in thermal and roughness characteristics, we did not expect contributions to the turbulent fluxes created by low-frequency turbulence that would require a longer averaging period (e.g. 1 h; Nordbo et al., 2011). The covariance between $w$ and $q$ or $T$ were first mathematically rotated so that the 30-min mean $w$ and $v$ (crosswind component) were zero using the procedure described in Baldocchi et al. (1988). Then $\lambda E$ was