J. Great Lakes Res. 26(3):305–314
Internat. Assoc. Great Lakes Res., 2000

Surface Energy Fluxes on the Great Lakes Based on Satellite-Observed Surface Temperatures 1992 to 1995

Brent M. Lofgren[1,*] and Yongchun Zhu[2]

[1]*NOAA/Great Lakes Environmental Research Laboratory*
2205 Commonwealth Blvd.
Ann Arbor, Michigan 48105

[2]*Cooperative Institute for Limnology and Ecosystems Research*
University of Michigan
Ann Arbor, Michigan 48109

ABSTRACT. *Accurate estimates of surface energy exchange components are critical for understanding many physical processes of large lakes and their atmospheric environment. In this paper, the seasonal cycle of latent, sensible, and total heat flux from the surface of the Great Lakes is estimated. Lake surface temperatures derived from the NOAA/AVHRR satellite, along with meteorological data from surface station observations are incorporated in order to estimate spatial distributions of fluxes. Several well-known features are evident. Among these are the very high outgoing fluxes of latent and sensible heat during the late fall and early winter, which drive strong cooling of the lake surface and consequent convective mixing within the lake column. Another is greater seasonal variation of surface temperature and fluxes in shallower waters than in deeper waters. Due to strong static stability of the overlying atmospheric boundary layer during the spring, both the magnitude and the spatial variations of latent and sensible heat fluxes are small during the spring and, to a lesser degree, during the summer. The annual cycles of latent and sensible heat flux over the Great Lakes are roughly opposite in phase to the same fluxes over land, indicating a large exchange of energy via atmospheric advection between the lake and land surfaces. A major weakness of the method used here is that heat fluxes are calculated on the basis of an ice-free surface, making the derived fluxes for January through March roughly estimated.*

INDEX WORDS: *Surface energy flux, climate, energy budget, Great Lakes.*

INTRODUCTION

Accurate characterization of the exchange of heat between the Great Lakes and the atmosphere is important for analysis of climate change and lake hydrodynamics. Surface heat fluxes are of paramount importance in forcing meteorological phenomena such as lake effect snow, lake breeze, thunderstorms, downwind moderation of air temperatures, and the static stability of the air column in the vicinity of the Great Lakes. Anomalies in lake surface temperature can manifest themselves through changes in these resultant atmospheric phenomena. Because the latent heat of evaporation represents the link between the energy and water budgets of the Great Lakes, lake surface energy fluxes can make a large difference in the amount of water available to the hydrologic system of the Great Lakes.

The surface temperature of the Great Lakes is part of a feedback loop involving their energy budget. At the same time that the lake surface temperatures influence the sensible, latent, and thermal infrared heat fluxes, the lake surface temperature is itself affected by those same fluxes. Subsurface temperatures also come into play through diffusive exchange of heat with the surface. This entire process influences the static stability of the water column, which is of great importance in determining lake hydrodynamics.

The intensive field campaign of the International Field Year for the Great Lakes (IFYGL, Pinsak and Rodgers 1981) provided heat fluxes averaged over Lake Ontario. This campaign used a method in which latent and sensible heat fluxes were calcu-

*Corresponding author: E-mail: lofgren@glerl.noaa.gov

lated as residual quantities in the heat budget, with the net radiation and the Bowen ratio derived from direct observations. Schertzer (1978) made similar calculations for Lake Superior and Bolsenga (1975) made them for Lake Huron. Derecki (1975) and Schertzer (1987) used a mass transfer method (equivalent to a simple bulk aerodynamic formulation) to derive turbulent heat fluxes over Lake Erie and Derecki (1975) also used a water balance method. All of these studies are summarized in Schertzer (1997). Using a model of vertical thermal mixing in the Great Lakes, combined with horizontal distributions of anomalies from the spatial mean temperature based on satellite observations, K. Schneider (personal communication, 1993) estimated surface heat fluxes for one seasonal cycle over all of the Great Lakes. Chu (1998) presents heat fluxes over time periods of several days from a dynamical model of Lake Erie, with the inclusion of a parameterization of cloud radiative effects based on satellite observations. It demonstrates the key role of surface heat fluxes in a dynamical lake model.

The work described here extends the results of these previous investigations, using an approach similar to the mass transfer method, with a major difference being that the formulation here uses diffusion coefficients that depend on the static stability and wind shear in the atmospheric boundary layer atmosphere. Newer technology contributes by providing remotely sensed lake surface temperatures with full coverage over the Great Lakes. This paper summarizes the methodology and derived spatial distributions over each of the Great Lakes for latent, sensible, and net heat fluxes on a monthly climatological basis. Additional details in graphical form can be found in Lofgren and Zhu (1999).

METHOD

Great Lakes heat fluxes are calculated using the same method as in Croley (1989), where the details of the formulation are presented. The basic equations for evaporation and sensible heat flux are similar:

$$H = C_p \rho U_* \theta_* \tag{1}$$

$$LE = -L \rho U_* q_* \tag{2}$$

where H is the sensible heat flux *into* the lake surface, C_p is the heat capacity of air at constant pressure, L is the latent heat per unit of evaporation, E is the evaporation rate, ρ is the air density, U_* is the frictional velocity, θ_* is the frictional temperature, and q_* is the frictional mixing ratio. The product LE is the latent heat flux out of the lake surface. Note that heat flux quantities in the tables and figures to follow will use the convention that positive numbers indicate heat fluxes *into* the lake surface, so they will show $-LE$ as the latent heat flux.

The frictional velocity, temperature, and mixing ratio in the above equations are calculated as follows:

$$U_* = \frac{kU}{[\ln(z/z_0) - S_1]} \tag{3}$$

$$\theta_* = \frac{k(\theta_a - \theta_w)}{[\ln(z/z_0) - S_2]} \tag{4}$$

$$q_* = \frac{k(q_a - q_w)}{[\ln(z/z_0) - S_2]} \tag{5}$$

where U is the wind speed at the reference height z, k is von Kármán's constant, z_0 is the roughness length, S_1 and S_2 are adjustments to the fluxes that depend on the static stability and wind sheer of the atmospheric boundary layer (Croley 1989), θ_a is the potential temperature of the air at the reference height, θ_w is the potential temperature of the water surface, q_a is the water vapor mixing ratio at the reference height, and q_w is the saturation mixing ratio at the temperature of the water surface. The Charnock relation is used to represent water surface roughness (z_0) as a function of U_* (Croley 1989).

Under a strict interpretation, the potential temperature would be given by:

$$\theta = T\,(p_0/p)^{R/C_p} \tag{6}$$

where θ is the potential temperature, T is the *in situ* temperature, p is the air pressure, p_0 is a reference air pressure of 1,000 hPa, R is the ideal gas constant for dry air (287 J/kg/K), and C_p is the thermal capacity of dry air at a constant pressure (1,004 J/kg/K). The reference height is taken as 8 m above the surface; using the hydrostatic approximation and an air density of 1.2 kg/m^3, the surface–reference height differential results in a pressure difference of 0.94 hPa. At an ambient pressure near 1,000 hPa, this discrepancy in pressure would reduce the potential temperature by a factor of 2.7×10^{-4}. For temperatures within the expected range of approximately 240 to 305K, this results in a maximum error in relative potential temperature between the two levels of 0.082K. Given this small possible error, the *in situ* temperatures are used in place of potential temperatures.

The full system of heat flux equations (1) through (6) is solved iteratively, because the factors S_1 and S_2 and the roughness length depend on the frictional velocity and the sensible heat flux. For purposes of calculating the stability dependence as in equation (6) of Croley (1989), the near-surface air temperature, denoted as Υ, is given the constant value of 276.5K; small variability relative to this absolute temperature would result in minimal changes to the surface fluxes.

This paper focuses on sensible (H) and latent heat (LE) fluxes, but also presents total heat flux Q into the lake surface:

$$Q = R_s + R_{dl} - R_{ul} - LE + H \qquad (7a)$$

where R_s is absorbed solar (shortwave) radiation, R_{dl} is downward longwave radiation from the atmosphere and clouds, and R_{ul} is upward longwave radiation from the surface. This paper also refers to net radiative heating R_n:

$$R_n = R_s + R_{dl} - R_{ul} \qquad (7b)$$

The absorbed solar radiation, downward longwave radiation, and upward longwave radiation are estimated as in Croley (1989).

INPUT DATA

The model described above requires both meteorological data and lake surface temperature data as input. The meteorological data are historical data collected at stations surrounding the Great Lakes (Fig. 1) and obtained from a database at the Great Lakes Environmental Research Laboratory (T. Croley and T. Hunter 1998, personal communication). These data include air temperature and dew point at screen height (2 m), cloud cover, and wind speed at 10 m above ground level. These data were transferred to a grid with 10 km spacing using an inverse-distance-weighting technique. Equations (11) through (13) from Croley (1989) are then applied to estimate overwater meteorological conditions. These equations are based on empirical parameters using the method of Phillips and Irbe (1978) and depend on water temperature, land-based air temperature, wind speed, and land-based dew point temperature, but not on wind fetch. The dewpoint temperature is converted into water vapor mixing ratio or water vapor pressure, as appropriate for the equation in which it is being used. These data were from the period 1992 to 1995. One special case is the station at Thunder Bay, Ontario (indicated by a circle next to its location in Fig. 1). It was found that the air temperature data at Thunder Bay were severely inconsistent with those of nearby stations throughout the years 1994 and 1995. Therefore, the data from this station were discarded during those times.

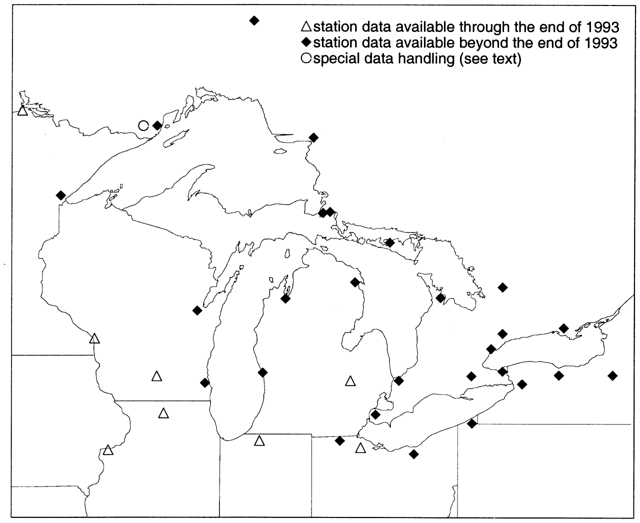

FIG. 1. The distribution of stations from which meteorological data are available. The diamonds indicate that data were available for the entire time span 1992 to 1995. The squares indicate that data were available for those stations only through the end of 1993. The station at Thunder Bay, Ontario has an open circle next to it to indicate special data handling because of apparent data errors (see text).

The lake surface temperatures are from the Great Lakes Coastwatch Program (http://coastwatch.glerl.noaa.gov). They are derived from Advanced Very High Resolution Radiometer (AVHRR) measurements of surface temperatures taken from the National Oceanic and Atmospheric Administration (NOAA) Polar Orbiting Satellites. They are available on a 2.56 km grid, but have been transferred to a 10 km grid using inverse-distance weighting from neighboring points within 10 km. Although this may be an unconventional method of transferring data from one grid to another, it was used here because of pre-existing software. Subsequent comparisons show that the resulting temperature fields match very well with those obtained using two-dimensional linear interpolation and nearest-neighbor schemes. The available data cover 1992 to 1995, although for the first 3 months of each year, data from 1992 to 1994 are discarded because of the lack of satellite data that has been processed to compensate for cloud contamination. The results for the months of January, February, and March are shown in the following section as means over 1995 only, whereas the other months are averaged over the 4 years 1992 to 1995. For this reason, and because ice cover has been ignored in the calculation of heat fluxes, the fluxes presented here for the first 3 months of the year should be regarded with particular caution. This is especially true for Lake Erie, whose shallow depth allows ice to form there more frequently than on the other lakes (Assel *et al.* 1983).

RESULTS

The annual cycle of heat fluxes is shown in Figure 2, which gives monthly spatial means over each lake of each type of heat flux, using data from 1992 to 1995, except for January, February, and March, which pertain only to 1995. Figures 3 to 5 show spatial distributions which, for the sake of space and clarity, are limited to Lake Huron and only alternating months of the year. Readers interested in more complete data, in the form of color maps, are encouraged to see Lofgren and Zhu (1999) and its supplementary figures available through the Internet. The annual and spatial means of each type of heat flux for each lake are given in Table 1. Table 2 shows the monthly means and spatial standard deviation over Lake Huron for each month.

Latent Heat Flux

The dotted curves in Figure 2 and the maps in Figure 3 show a predictable seasonal cycle of latent heat flux. Because the latent heat of evaporation, L, is nearly constant (a very weak function of water temperature), the maps of Figure 3 are nearly proportional to maps of evaporation. There are large negative latent heat fluxes at the beginning and end of the year (the winter), and much smaller values during the summer, even positive at times. Positive latent heat flux (negative evaporation) corresponds to water condensation at the surface, or may be construed as the formation of fog just above the surface

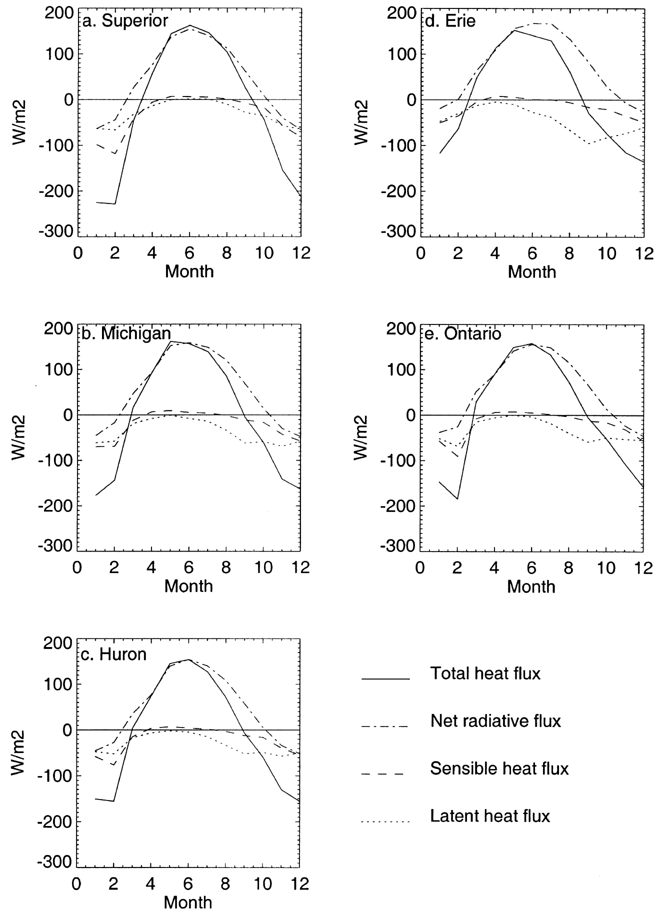

FIG. 2. Annual cycle of heat fluxes (W/m²) averaged over (a) Lake Superior, (b) Lake Michigan, (c) Lake Huron, (d) Lake Erie, and (e) Lake Ontario. Calculated fluxes for April to December are averages for 1992 to 1995, while January through March represent 1995 values.

that warms the water by precipitating into it. Such processes are often suppressed in models using bulk aerodynamic schemes for evaporation and surface energy exchange. Although these processes may not be very accurately portrayed by the methodology used in this study, they are retained here. Negative evaporation tends to have very small amplitude, as it always occurs in a situation in which the atmosphere above the lake is stable, yielding little turbulent mixing in the atmospheric boundary layer.

January has the largest negative latent heat flux over most areas (Fig. 3a and Table 2). This results, in part, from the low absolute humidity of the cold overlying air and the relative warmth of the water, which causes high humidity in the very lowest lev-

TABLE 1. Annual mean heat fluxes (W/m^2) averaged over each of the Great Lakes. A positive flux indicates a heat flux* into *the lake surface.

	Latent heat flux	Sensible heat flux	Net radiative flux	Total heat flux
Superior	–30.35	–33.45	42.34	–21.46
Michigan	–37.34	–20.70	55.66	–2.38
Huron	–31.90	–21.10	47.13	–5.87
Erie	–45.41	–16.60	71.76	9.75
Ontario	–34.57	–20.58	51.91	–3.24

TABLE 2. Monthly mean heat fluxes (W/m^2) averaged over Lake Huron, with spatial standard deviation in parentheses, along with Bowen ratios based on monthly mean values. For January, February, and March, the means and standard deviations are from 1995; for all other months, they are from 1992 through 1995. A positive flux indicates a flux* into *the lake surface.

	Latent heat flux	Sensible heat flux	Net radiative flux	Total heat flux	Bowen ratio
January	–47.23 (9.98)	–57.85 (11.36)	–45.50 (7.06)	–150.58 (27.24)	1.22
February	–52.01 (25.06)	–76.24 (35.69)	–27.25 (12.97)	–155.50 (72.88)	1.47
March	–16.78 (7.97)	–14.79 (9.65)	35.09 (10.73)	3.52 (12.02)	0.88
April	–6.37 (1.84)	3.61 (1.54)	76.75 (7.07)	73.99 (7.39)	–0.57
May	–1.66 (7.28)	7.19 (2.34)	139.71 (4.75)	145.24 (12.58)	–4.33
June	–4.55 (11.39)	4.28 (2.66)	154.50 (6.53)	154.23 (17.86)	–0.94
July	–14.46 (12.92)	1.74 (2.48)	140.39 (5.79)	127.67 (18.21)	–0.12
August	–33.60 (13.98)	–3.03 (2.52)	108.05 (6.23)	71.42 (20.62)	0.09
September	–51.92 (13.81)	–12.35 (3.17)	57.96 (3.72)	–6.31 (18.79)	0.24
October	–48.58 (10.73)	–16.06 (3.73)	5.75 (2.54)	–58.89 (15.45)	0.33
November	–57.25 (7.79)	–39.49 (6.04)	–36.53 (3.46)	–130.27 (14.81)	0.69
December	–50.03 (7.68)	–54.23 (8.59)	–51.51 (5.40)	–155.77 (20.29)	1.08

els above the lake. Strong winter winds and static instability of the atmospheric boundary layer result in larger values of S_1 and S_2 in (3) and (5), further enhancing the latent heat flux. The formulation of S_1 and S_2 is described in Croley (1989). The water temperature cannot dip below 0°C on the macroscale, but the air temperature can, with humidity dropping correspondingly. Because the air remains cold into February, these strong negative latent heat fluxes diminish only slightly, although in the real world, they could be greatly diminished by the formation of ice.

During March, the air warms and moistens considerably while the lake water cools, leading to considerably weaker negative latent heat flux (Fig. 3b and Table 2). This trend continues during April, and by May (Fig. 3c and Table 2), most of the area of the large lakes has near zero latent heat flux. Lake Erie and shallow areas of the other lakes are exceptions to this. Their water has warmed rapidly enough to keep better pace with the air temperature, allowing them to maintain some evaporation.

Through June and July (Fig. 3d and Table 2), the area of active evaporation spreads from the shorelines toward the deeper areas at the centers of the lakes. By August, most of the lakes have evaporation occurring, but the evaporation remains strongest in the shallower areas. Lake Superior, the northernmost and deepest of the Great Lakes, warms more slowly than the other lakes, and thus has the least negative latent heat flux throughout the summer.

The deeper lakes tend to have their maximum surface temperature during September, and meanwhile the air temperatures and humidities are decreasing. These factors are reflected in further increases in the magnitude of negative latent heat flux during September (Fig. 3e and Table 2). Although the water cools throughout the fall, the air cools and dries more rapidly, resulting in a larger

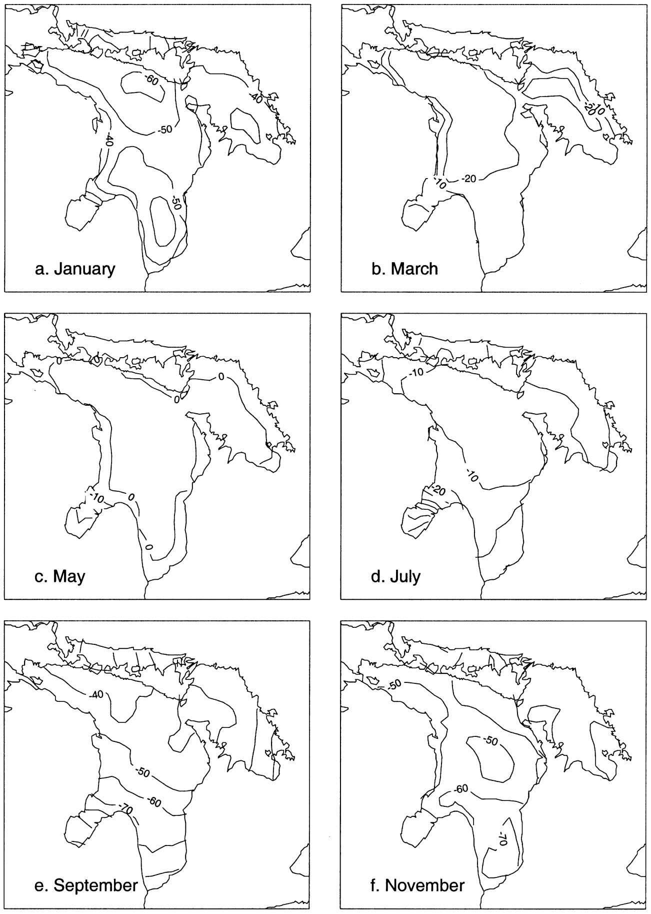

FIG. 3. The spatial distribution of 1992 to 1995 monthly mean latent heat flux (W/m²) over Lake Huron during the month of (a) January, (b) March, (c) May, (d) July, (e) September, and (f) November. The contour interval is 10 W/m².

gradient in humidity between the lake surface and the overlying air. This enhanced gradient directly increases the negative latent heat flux as in (5). The accompanying unstable state of the atmospheric boundary layer further increases the negative latent heat flux by increasing S_1 and S_2 in (3) and (5). This latter effect is evidenced by the continued large negative latent heat flux during October, November (Fig. 3f), and December (Table 2).

Lake Erie and shallow parts of the other lakes are exceptions to this general rule—they cool more quickly and do not maintain high rates of latent heat flux into the late fall and early winter. In general, throughout the year, because of the lower heat capacity of the shallower parts of the Great Lakes, their temperatures are closer to those of the air than in the deeper areas, making the amplitude of the seasonal cycle of latent heat flux smaller. Also, the phase of the annual lake temperature cycle lags the air temperature less in shallower areas, meaning that the annual cycle of latent heat flux lags by less.

Note also that the standard deviation in space of the latent heat flux (Table 2) follows the magnitude of the latent heat flux as a general rule. Because the smaller-magnitude latent heat fluxes occur during periods of atmospheric stability, their sensitivity to lake temperature is reduced, leading to reduced spatial variability.

Sensible Heat Flux

Sensible heat fluxes are shown in the dashed curves in Figure 2, the maps in Figure 4, and in Table 2. The sensible heat flux generally follows the same trends as the latent heat flux, being large and negative in the fall and winter and much

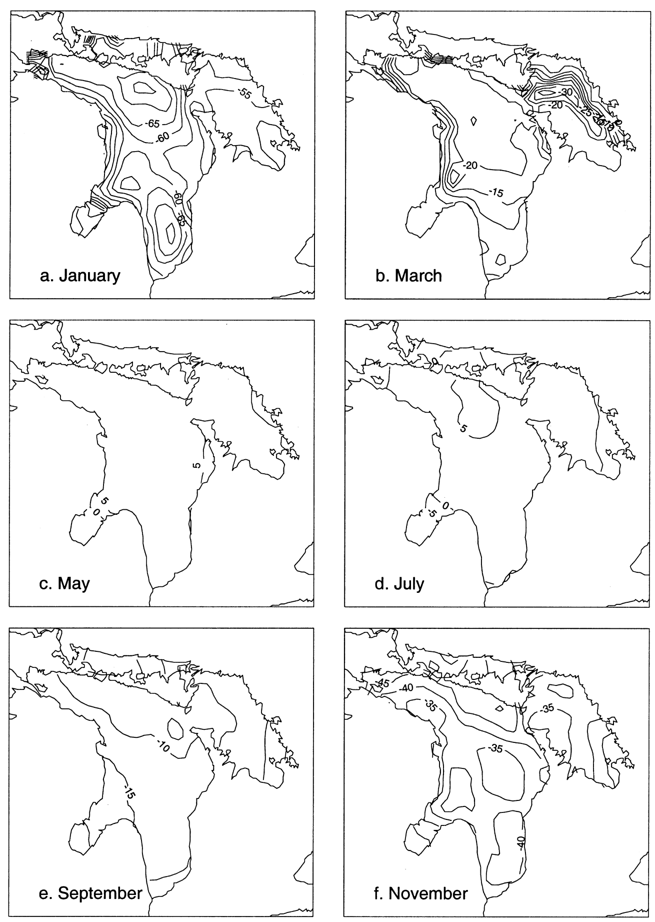

FIG. 4. As in Figure 3, but for sensible heat flux, and with a contour interval of 5 W/m².

smaller during the spring and summer. There are especially large negative winter values in deeper parts of the lakes (Figs. 4a,b for Lake Huron). From April until August, many of the sensible heat flux values are positive, indicating that the water surface is colder than the overlying air. This situation is more readily achieved than one of positive latent heat flux, which requires that the humidity of the air be greater than the saturation humidity at the surface water temperature. Additionally, drier air will require a more strongly stable atmospheric boundary layer for positive latent heat flux to occur, suppressing the turbulence that helps drive evaporation. As with the latent heat flux, the more stable atmosphere during the spring and summer reduces the spatial variability of the sensible heat flux (Table 2 and Fig. 4).

The magnitude of summertime fluxes of both sensible and latent heat are much smaller than those during the winter, despite large temperature differences between the surface and the atmosphere. This results from the strong dependence of these fluxes on atmospheric stability as given through the definitions of Croley (1989) of the factors S_1 and S_2, combined with the dependency of the surface roughness on the frictional velocity in the Charnock relation. Also because of this, when considering monthly mean heat flux values, heat fluxes out of the lake during a cold spell can outweigh small heat fluxes into the lake during more normal spring or summer conditions. Whether the cumulative sensitivity of surface heat fluxes to atmospheric stability and surface roughness length is realistic is unknown without field verification.

Bowen Ratio

The Bowen ratio (sensible heat flux divided by latent heat flux) is an important measure. Given a finite amount of available energy from net radiative flux, the average Bowen ratio determines the amount of energy that goes into sensible heat flux, which can strongly affect the static stability of the lower atmosphere and the thermal forcing of lake breezes. A greater proportion of latent heat flux can increase the relative humidity of the atmospheric boundary layer and lower the lifted condensation level (cloud base). The monthly lake-mean Bowen ratio for Lake Huron can be found in Table 2, and a rough estimate for other lakes can be derived by inspecting the sensible and latent heat curves in Figure 2. While the sensible heat flux is less negative than the latent heat flux throughout most of the year, sensible heat flux has a larger negative value than latent heat flux during January, February, and December (Fig. 2). This would help to destabilize the atmospheric boundary layer and force it to mix more strongly upward (a process that is not explicitly considered in this model) and thus allow for greater surface-to-atmosphere energy and moisture fluxes. At relatively high water temperatures, the strong dependence of the saturation humidity on temperature ensures that the latent heat flux from a water surface will always be stronger than the sensible heat flux. However, the (saturated) surface humidity is a nonlinear function of water surface temperature and has a lesser dependence on temperature at lower temperatures.

An estimate can be arrived at, using simplifying assumptions, of the temperature at which sensible heat flux switches from being less than latent heat flux to greater, i.e., the temperature at which the Bowen ratio equals one. Using (1), (2), (4), and (5), the Bowen ratio is unity when

$$B = C_p(\theta_a - \theta_w)/[L(q_a - q_w)] = 1 \tag{8}$$

The small difference between the surface and the reference height is ignored by substituting the difference in *in situ* temperatures for the difference in potential temperatures. It is further assumed that the air is saturated, making this the case with the lowest possible evaporation and thus the upper limit of temperature at which the Bowen ratio could possibly be unity. In this case,

$$B = C_p\Delta T/L\Delta q_s = 1 \tag{9}$$

where ΔT is the difference in air temperature between the surface and the reference level and Δq_s is the corresponding difference in saturation mixing ratio at a given temperature. Given that the mixing ratio $q \simeq .61e/p$, where e is the water vapor pressure and p is the air pressure (assume 1,000 hPa), and in the limit of small ΔT, $B = 1$ when

$$de/dT = pC_p/.61L = .655 \text{ hPa/K} \tag{10}$$

Using a standard water vapor pressure table, the condition for $B = 1$ is satisfied at approximately 6°C (279K). Thus, a temperature, taken as a mean between the water and atmosphere, below about 6°C is a necessary, but not sufficient, condition for having a Bowen ratio greater than one over a water surface. The condition of Bowen ratio being greater than 1 is prominent on Lake Erie, where the latent heat flux peaks in September, but the sensible heat flux does not reach its peak until February.

Net Radiative Flux and Total Heat Flux

The net radiative flux, defined in (7b) is shown in the dash-dotted curves in Figure 2 and in Table 2. Net radiative flux is dominated by incident solar radiation, modulated by surface albedo and cloud cover effects on solar radiation. It is also dependent on downward longwave radiation as a function of atmospheric temperature, humidity, and cloud cover, and on surface temperature-dependent upward longwave radiation. Its seasonal peak in June and trough in December are in direct agreement with the seasonal cycle of incident solar radiation.

The solid curves in Figure 2, the maps in Figure 5, and Table 2 show the monthly values of total heat flux. This adds net radiative heating to the sensible and latent heat fluxes, as in (7a). According to Figure 2, each lake has negative total heat flux from October through February, with additional negative rates appearing in March on Lakes Superior and Huron, and September on Lakes Michigan and Erie. It is expected that the annual mean total heat flux over an entire lake will be very close to zero. Except for Lake Superior, where it is about –21 W/m^2, its magnitude is smaller than 10 W/m^2 over each lake (Table 1). These are reasonably good results, considering that there is no constraint built into the calculations to guarantee a closure of the heat budget, and that there is a wide variety of possible errors in the input data and flux calculation methods.

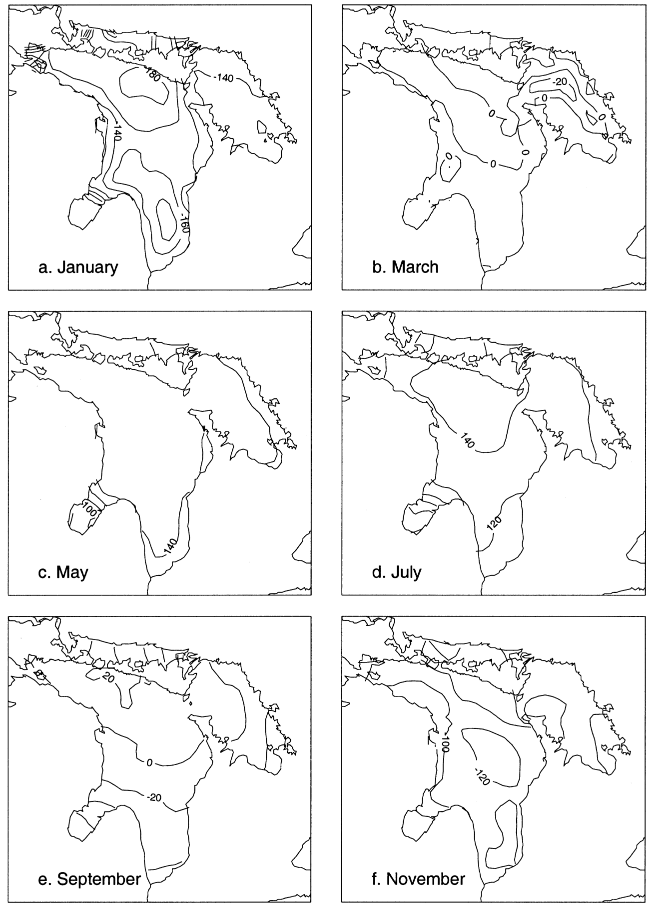

FIG. 5. As in Figure 3, but for total heat flux, and with a contour interval of 20 W/m^2.

Comparison With Other Studies

Comparisons of these results with those of other studies yield mixed results. The latent heat fluxes derived in this study compare well with those estimated for Lake Superior (Schertzer 1978), Lake Huron (Bolsenga 1975), and Lake Ontario (Pinsak and Rodgers 1981), except during the fall, when the estimates tend to be less negative than those in the previous studies. Estimates of sensible heat flux were generally less than previous studies; there was increased heat loss during the late fall and winter and decreased heat gain during the summer through sensible heat flux. These differences in sensible heat flux can be explained by the boundary-layer stability dependence that was introduced into the model formulation used in this study. The change in latent heat flux is less readily explained upon these grounds.

The three studies mentioned (Schertzer 1978, Bolsenga 1975, and Pinsak and Rodgers 1981) all used estimates of the Bowen ratio taken from boundary layer gradients in temperature and water vapor mixing ratio derived from observations at shore stations. These estimates involved assumptions that differed from those used in this study regarding the relationship between temperatures and humidities observed at land stations and those prevailing over the lakes. These varying assumptions resulted in different Bowen ratios, with those given by the present study generally exceeding those in previous studies.

The agreement with previous studies in latent heat flux over Lake Erie is quite poor. The monthly fluxes given by the present study are roughly half those of Derecki (1975) using either the mass transfer or water budget method or Schertzer (1987) using the mass transfer method. The reasons for this cannot be fully explained by the inclusion of boundary-layer stability dependence or the use of differing assumptions regarding the relation between overland and overlake temperature and humidity. Schertzer (1987) achieved closure in the overall energy budget to within approximately 3 W/m^2. Particularly vexing is that the water budget method does not rely on overlake meteorological conditions, but should yield an accurate lake evaporation rate as a residual in the water budget of the lake, with the main uncertainty lying in the overlake precipitation rate. The results obtained in this study are not directly comparable to those of Chu (1998). Chu's technique differs in that the reported heat fluxes were calculated from lake temperatures that were modeled, diurnal variation in energy fluxes was taken into account, and the energy fluxes were reported as hourly values over periods of roughly 15 days, rather than on a time-averaged basis.

Seasonal Phase and Atmospheric Heat Transport Implications

As illustrated in Figure 2, the annual cycle of the total heat flux is nearly in phase with those of the sensible and latent heat flux, and larger in magnitude than their sum. This means that the sensible and latent heat fluxes are working in concert with the net radiative heating to yield the annual cycle of water temperatures. The radiative heating warms the water most strongly during the summer. The sensible and latent heat fluxes each act overall to cool the water, most strongly during the winter. This constructive relationship between the phases of the solar and turbulent fluxes is opposite to the usual situation over land, where the low thermal capacity and thermal conductivity of the soil constrain the turbulent fluxes to approximately balance the net radiation in the diurnal average.

In terms of the heat budget of the atmospheric boundary layer over the Great Lakes, the wintertime inputs of latent and sensible heat from the lakes may be partially offset by radiative loss (although this might also be offset by absorption of longwave radiation upwelling from the relatively warm water surface). However, they may be primarily compensated by net horizontal divergence of advective heat flux from the lakes, in the form of both latent and sensible heat. This means that surrounding continental regions, possibly from a very wide area, contribute to cooling the lakes by acting as a heat sink. Conversely, during the summer, those continental regions act as a heat source for the lakes, although this relationship is weaker.

DISCUSSION AND CONCLUSIONS

This paper has presented seasonal cycles of latent, sensible, and total heat fluxes for the Laurentian Great Lakes, illustrated with spatial distributions for Lake Huron. These fluxes represent the effects of the thermal capacity of these large water bodies, which causes a phase lag between their surface temperature and the temperature of the overlying air. This results in strong latent and sensible heat fluxes out of the lakes during the winter, when the relatively warm water and cold, dry air create a strong vertical gradient in temperature and humidity, combined with strong turbulence due to instability of the atmospheric boundary layer. During the summer, the vertical temperature gradient is reversed, and sometimes the humidity gradient is also, but strong static stability of the atmospheric boundary layer suppresses turbulent fluxes in the opposite direction.

The total heat flux is negative during the winter and positive during the summer. The calculated fluxes contrast sharply with the typical annual cycle over neighboring land surfaces, where total heat flux is near zero year-round, and turbulent fluxes are most negative during the summer. This suggests that a significant exchange of energy takes place through atmospheric advection between the lakes and surrounding land areas.

Whether the results presented here, covering the period 1992 to 1995, are representative of prior or subsequent time periods is uncertain. Secular trends in global air temperatures, which are likely to affect the Great Lakes region, have been modeled (Houghton *et al.* 1996) and observed (Easterling *et al.* 1997). Long-term variability in air temperatures can also occur naturally and water temperatures can undergo natural and anthropogenic changes, any of which can affect surface energy fluxes.

There is no constraint built into the simulations presented here that there be an overall energy balance over the annual cycle, but the results were mostly good in this respect. A major caveat is that ice was not considered as a factor in the heat flux calculations, which would particularly influence the months of January through March.

The lack of agreement with several previous studies, particularly that of Derecki (1975), in which a water budget method was used for estimation of evaporation over Lake Erie, will cause some pause in taking literally the quantitative lake-air fluxes given here. An extensive field campaign may be required in order to rectify this situation and devise a full set of components to a bulk aerodynamic formulation (including equations analogous to all of those in the formulation given by Croley 1989) that are derived in concert with one another rather than independently, and yield agreement with independent derivations of lake surface heat fluxes, such as those measured by eddy-correlation instruments or by water budget closure.

An additional possible source of error in this study was the necessity of using an empirical parameterization to convert meteorological variables measured over land to estimated values over the lakes. In particular, the method used here (based on Phillips and Irbe 1978) had no dependence on wind fetch, a consideration which would likely lead to larger turbulent energy fluxes on the windward side of each lake.

ACKNOWLEDGMENTS

Thanks to G. Leshkevich for providing the CoastWatch satellite data, to T. Hunter for assistance in acquiring and preparing the meteorological data, to D. Schwab for providing the geospatial interpolation code and for technical advice on the preparation of figures, and to T. Croley for advice on using the heat flux simulation system. Thanks to the two reviewers for very helpful comments that have improved the content and presentation of this paper. Yongchun Zhu was supported partially by an NSF Postdoctoral Fellowship and partially by a CILER Postdoctoral Fellowship funded by the Great Lakes Environmental Research Laboratory. This is GLERL Contribution No. 1163.

REFERENCES

Assel, R.A., Quinn, F.H., Bolsenga, S.J., and Leshkevich, G.A. 1983. *NOAA Great Lakes Ice Atlas*. Ann Arbor, MI: Great Lakes Environmental Research Laboratory.

Bolsenga, S.J. 1975. Estimating energy budget components to determine Lake Huron evaporation. *Wat. Resour. Res.* 11:661–666.

Chu, Y.-F. 1998. The incorporation of hourly GOES data in a surface heat flux model and its impacts on operational temperature predictions in bodies of water, Ph.D. thesis, The Ohio State University, Columbus, OH.

Croley, T.E., II 1989. Verifiable evaporation modeling on the Laurentian Great Lakes. *Water Resour. Res.* 25:781–792.

Derecki, J.A. 1975. *Evaporation from Lake Erie.* NOAA Technical Report, Ann Arbor, MI. ERL 342–GLERL 3.

Easterling, D.R., Horton, B., Jones, P.D., Peterson, T.C., Karl, T.R., Parker, D.E., Salinger, M.J., Razuvayev, V., Plummer, N., Jamason, P. and Folland, C.K. 1997. Maximum and minimum temperature trends for the globe. *Science* 277:364–367.

Houghton, J.T., Meira Filho, L.G., Callender, B.A., Harris, N., Kattenberg, A., and Maskell, K. 1996. *Climate Change 1995: The Science of Climate Change*. Cambridge, England: Cambridge University Press.

Lofgren, B.M., and Zhu, Y. 1999. *Seasonal climatology of surface energy fluxes on the Great Lakes*. NOAA Technical Memorandum, Ann Arbor, MI. ERL GLERL-112. (Available via File Transfer Protocol at ftp://ftp.glerl.noaa.gov/publications/tech_reports/glerl-112/tm-112.pdf. Supplementary graphics in JPEG format are available in the directory ftp://ftp.glerl.noaa.gov/publications/tech_reports/glerl-112/jpgs.)

Phillips, W.D., and Irbe, J.G. 1978. *Land-to-lake comparison of wind, temperature, and humidity on Lake Ontario during the International Field Year for the Great Lakes (IFYGL).* Downsview, Ont.: Atmospheric Environment Service, Environment Canada. Rep. CLI-2-77.

Pinsak, A.P., and Rodgers, G.K. 1981. Energy balance. In *IFYGL–The Internation Field Year for the Great Lakes*, eds. E.J. Aubert and T.L. Richards, pp. 169–197. Ann Arbor, MI: Great Lakes Environmental Research Laboratory.

Schertzer, W.M. 1978. Energy budget and monthly evaporation estimates for Lake Superior, 1973. *J. Great Lakes Res.* 4:320–330.

_____. 1987. Heat balance and heat storage estimates for Lake Erie, 1967 to 1982. *J. Great Lakes Res.* 13:454–467.

_____. 1997. Freshwater lakes. In *The Surface Climates of Canada*, eds. W.G. Bailey, T.R. Oke, and W.R. Rouse, pp. 124–148. Montréal, PQ: McGill-Queen's University Press.

_____, and Croley, T.E., II 1999. Climate and lake responses. In *Potential Climate Change Effects on Great Lakes Hydrodynamics and Water Quality,* eds. D.C.L. Lam and W.M. Schertzer, pp. 1–74. Reston, VA: ASCE Press.

Submitted: 20 August 1999
Accepted: 16 May 2000
Editorial handling: David J. Schwab