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Bowen Ratio Estimates over Lake Erie

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ABSTRACT. *Estimates of the ratio of sensible heat flux to latent heat flux (the Bowen ratio) are derived for Lake Erie based on the method introduced by Roll (1965), using hourly automated observations of lake and air temperatures for Buoy 45005 for the period 1992–1997 (May through November). Roll's method computes* β *as the product of the drag coefficient (which varies by atmospheric stability class) and the ratio of the difference between lake temperature (*T_{sea}*) and air temperature (*T_{air}*), to the sea-air vapor pressure difference. The latter can be estimated empirically as a function of Tsea and dew point temperature (*T_d*). In the diurnal cycle,* β *usually peaks in the afternoon hours and tends to increase as the warm season proceeds. Specifically, during unstable atmospheric conditions,* β *varies from approximately .15 to .30, and during times of high static stability,* β *tends to have near-zero to slightly negative values. Results allow for an improved understanding of turbulent energy fluxes from a large water body, which may affect other processes, such as ice cover, evaporation rates, and contaminant advection.*

INDEX WORDS: *Bowen ratio, energy fluxes, Lake Erie.*

INTRODUCTION

Because of the economic, recreational, and aesthetic value of the Great Lakes, a firm scientific understanding of the physical, chemical, and biological processes therein is warranted. Many of these processes affect and are affected by atmospheric conditions. Unfortunately, however, direct measurement of all atmospheric variables of importance is done in only sparse locations or not at all. Therefore, official data that are collected must sometimes be used to estimate unmeasured atmospheric variables.

Examples of unmeasured data that have importance in physical, chemical, and biological processes in the Great Lakes are the terms in the surface energy balance. Since the magnitudes of the various components of the surface energy balance depend on many factors, including the type of surface and its characteristics, geographical location, time of year, time of day, and weather (Arya 1988), it is desirable to have field measurements of the energy balance components. Unfortunately, such measurements are lacking because of logistical

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difficulties and the expense involved with direct measurement using field equipment. Therefore, attempts have been made to devise equations to approximate the surface energy balance without requiring the equipment and data that are required by conventional field techniques (Roll 1965, Hsu 1988, Bello and Smith 1990, Rohli and Binkley 1999, Rohli and Hsu 1999).

Roll (1965) introduced a physically-based technique for computing one important feature of the surface energy balance over a water body: the Bowen Ratio (β), a dimensionless quantity which measures the ratio of the turbulent flux densities of sensible heating (Q_h) to latent heating (Q_e). Such a quantity is important to understand because knowledge of this variable quantifies the relative magnitudes of surface energy expenditure to heat the air and to evaporate water, and has been used widely in evaporation estimation (Revfeim and Jordan 1976). Roll's technique is advantageous because it only requires input based on readily-available meteorological data.

An approximation of β is also useful because even though it is usually possible to measure Q_h, even over a lake, Q_e is considerably more difficult to measure accurately since humidity is a required parameter. An estimate of β will allow for the absolute measurement of Q_e and Q_h by using a net pyranometer to measure net incoming radiation (Q^*) and heat flux plates to evaluate the substrate heat flux, Q_g (usually set to zero under the assumption of no net warming or cooling on long time scales).

Unfortunately, little research had been done to identify the validity of Roll's technique. One such effort was recently conducted on a pilot study basis for Louisiana's Lake Pontchartrain (Rohli and Hsu 1999). Results indicated that the technique appears to produce reasonable results.

The purpose of this research is to implement Roll's (1965) technique to estimate β over Lake Erie. This analysis will determine whether the lake has dimensions and thermal properties that cause it to behave like an ocean body (with β = .12 for the northern hemisphere average (Lewis 1995)) or a land-influenced body of water (with higher β). The computed values of β in unstable conditions are used to generate a simple power curve relationship between β and the air-sea temperature difference over the lake, and this regression relationship is compared to that computed by Hsu (1999) and Sadhuram *et al*. (2001) for verification of the computed values. This work supplements and expands on preliminary work of Rohli and Binkley (1999) in the climatological analysis of β over Lake Erie.

BACKGROUND

The Study Area

Lake Erie is selected for analysis for several reasons. First, since output from general circulation models (GCMs) is contradictory with regard to the impact of climate change on water supplies in the Great Lakes region (Lofgren *et al.* 2002), improved understanding of the relationship between changes in energy input and evaporation in Lake Erie seems warranted. Additionally, the impacts of climate in this region are felt by a dense nearby population for whom the lake is of tremendous economic, ecological, and recreational value. For example, changes in the local energy budget could affect ice cover and evaporation rates, and possibly to volatile contaminant movements in and out of the lake surface. Finally, Lake Erie is selected so that β values can be compared to those for Lake Pontchartrain, a subtropical lake in Louisiana for which β has already been estimated, and to other water bodies.

Roll's Method

The technique is described fully in Rohli and Hsu (1999), and therefore it will be reviewed only briefly here. The derivation begins with the well-known "bulk" equations for Q_h and Q_e (Hsu 1988, p. 112)

$$Q_h = \rho\, C_p\, C_h\, u_{10}\, (\theta_{sea} - \theta_{air}) \tag{1}$$

and

$$Q_e = \rho\, L_v\, C_e\, u_{10}\, (q_{sea} - q_{air}) \tag{2}$$

where ρ represents atmospheric density, C_p is the specific heat at constant pressure (1004 J kg^{-1} K^{-1}), u_{10} is the horizontal wind speed at the 10 m level (in m s^{-1}), L_v represents the latent heat of vaporization (2.5008×10^6 J kg^{-1}), q represents specific humidity, and θ is potential temperature. C_h and C_e represent dimensionless turbulent transfer coefficients for sensible and latent heat respectively, which are computed as the quotient of the eddy diffusivity for heat (and water vapor, respectively) and the product of the difference between the wind speed at two heights and those heights. Thus, Equations (1) and (2) above implicitly suggest that in order to maintain a constant flux density (W m^{-2}) of sensible or latent heat (as is required in the surface boundary layer), increases in transport with greater height must be balanced by corresponding decreases in temperature and moisture, respectively.

If air temperature (T_{air}) is taken instead of q, which does not imply any appreciable error at low levels (Hsu 1998, Hsu 1999), and the assumption is made that for unstable and near-neutral conditions over water bodies, $C_h \approx C_e$ (*i.e.*, 1.13×10^{-3} *vs.* 1.15×10^{-3}, see Hsu 1988, p. 113), the ratio of Q_h to Q_e simplifies to

$$\beta = (C_p\ (T_{sea} - T_{air})\ )\ /\ (L_v\ (q_{sea} - q_{air})\ ) \qquad (3)$$

Moreover, since

$$q = (0.622\ e)\ /\ p,$$

where e represents the partial atmospheric pressure exerted by water vapor and p represents surface atmospheric pressure (approximately 1.013×10^6 hPa), for unstable and near-neutral conditions we can substitute

$$\beta = 0.66\ (\ (T_{sea} - T_{air})\ )\ /\ (e_{sea} - e_{air}) \qquad (4)$$

where $e_{sea} = 6.1078 \times 10^{\ ((7.5\ Tsea)\ /\ (237.3\ +\ Tsea)\ )}$, $e_{air} = 6.1078 \times 10^{((7.5\ T_d\ )\ /\ (237.3\ +\ T_d)\ )}$, and T_d represents the dew point temperature (Hsu, 1988, pp. 20–21).

For cases in which $T_{sea} < T_{air}$, the near-surface atmosphere is considered to be stable. In such cases $C_h \approx .66 \times 10^{-3}$ and $C_e \approx 1.15 \times 10^{-3}$ (Hsu, 1988, p. 113), and the following approximation is used:

$$\beta = .38\ (T_{sea} - T_{air})\ /\ (e_{sea} - e_{air}) \qquad (5)$$

Note that in such cases, β is usually negative, since the flux of energy is generally from air to sea (the reverse of the "typical" case). Analyses of β are done by season and by time of day to determine whether there is any fluctuation in the diurnal or annual cycle.

This method has the advantage over profile measurement methods of not requiring field work or expensive equipment; the required atmospheric variables are often available from routine National Weather Service observations. Furthermore, even though u_{10} data are available for some sites, they are susceptible to considerable measurement error. Thus, it is convenient that measurement of u_{10} is not required for the computation of the ratio of Q_h/Q_e (β). Additionally, it is advantageous that wind measurements are not needed, because it is more appropriate that they be made at the same height as the temperature and dew point measurements (5 m above the surface).

While Similarity Theory (*e.g.*, Stull 1988) could be invoked to estimate the eddy diffusivity terms (and therefore C_h and C_e), such an application would require knowledge of the gradient Richardson number (difficult to measure accurately). Even then, however, the estimation of the non-dimensional lapse rate (Φ_H), non-dimensional humidity gradient (Φ_M), and the stability correction factors resulting from Φ_H and Φ_M would be done using empirical approximations. Moreover, since neutral conditions are best met under cloudy skies and strong winds in the lowest few meters of the atmosphere, Lake Erie affords an example of a location in which the Bulk Equations and the approximations in Equations (4) and (5) may provide a sufficiently reasonable estimation of the Bowen Ratio. Thus, a complete set of measurements using Similarity Theory would prove unnecessary.

DATA AND METHODS

Hourly automated observations of lake and air temperatures for Buoy 45005 for the period 1992-1997 (May through November) are obtained (United States Department of Commerce 1998). Figure 1 shows the location of the observation stations. Air temperature is measured at 5.0 meters above site elevation, and water temperature is measured 0.6 meters below site elevation. This deeper temperature is not exactly equivalent to the surface

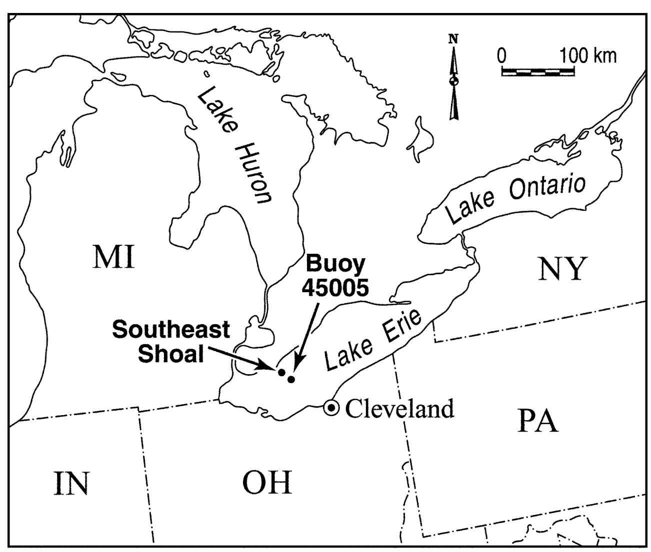

FIG. 1. The study area.

temperature. Also, its relationship to the surface temperature would vary somewhat with season because the water temperature profile would be a function of wind speed and radiation/heating conditions. Nevertheless, this value provides a reasonable approximation to the surface temperature (i.e., within ~1 CE under most cases (Wesely 1979)). Moreover, temperature from depths of within a meter have been used previously to represent surface temperature in the Great Lakes (Lesht and Brandner 1992).

This site is in the west central part of Lake Erie and is in water 13.7 meters deep. In addition, T_d data for the same temporal period are acquired for the automated station at Southeast Shoal, Ontario—a marine station that is 5.5 km from Buoy 45005 in west-central Lake Erie (Bryan Smith, Environment Canada, personal communication, 7 June 1998). It should be noted that regardless of wind direction, the lake fetch at Buoy 45005 and Southeast Shoal should be large enough to allow for true "marine" readings. At its greatest extent, a northerly wind would have an overland fetch of 47.6 km (from the southern edge of Lake St. Clair to the southern tip of Pelee Point, Ontario) before flowing over 8.3 km of water from the southern tip of Pelee Point to Southeast Shoal. Winds from other directions would have an even greater lake fetch. Furthermore, these data represent the most reliable meteorological data available for the lake.

Because of discrepancies between the two data sources and a peculiarity in the empirical equation, some data quality control activities are necessary. In rare hourly observations in which T_d at Southeast Shoal exceeds T_{air} at Buoy 45005, the observation is eliminated. In addition, because the difference between T_{sea} and T_d is in the denominator of the empirical equation, convergence of these two quantities produces infinitely large β values. Therefore, any hourly readings during which $|T_{sea} - T_d| < 0.5°C$ are deleted. In all, 22,747 of the original 24,460 observations (~93%) are retained for calculation of β by month and hour of the day.

The observations are then divided into the following stability categories: "unstable" ($T_{sea} > T_{air} + 0.5°C$), "near-neutral" ($|T_{sea} - T_{air}| \leq 0.5°C$), and "stable" ($T_{air} > T_{sea} + 0.5°C$). As noted earlier, the drag coefficient C_h used is 1.15 for near-neutral and unstable conditions and 0.66 for cases of thermal stability (Hsu 1988). In all, 11,425 cases are identified as "unstable," 4,840 hours are classified as "near-neutral," and 6,482 observations fall into the "stable" category. Values of β are calculated for each stability category by month and hour of the day.

Variability of β is then analyzed by synoptic weather type. A daily (1200 UTC) weather classification system for Cleveland, Ohio (Britton 1999) is used for this task. In all, 656 observations are included for the period 27 May 1992 to 21 November 1995. Britton's (1999) technique is based on the surface cyclone model which has been used to categorize weather types in the British Isles (Lamb 1972), Louisiana (Muller 1977), and Pennsylvania (Comrie 1992). Previous analysis suggests that the classification does provide accurate stratification of weather types (Britton 1999), and although the number of events per month in this analysis is small, some differences in meteorological characteristics are apparent. It should be noted that Cleveland is deemed to be sufficiently near the study area that the generic weather classification pattern for Cleveland is still valid over the western part of Lake Erie.

In the Britton (1999) system, *Continental High (CH)* occurs when a mid-latitude surface anticyclone (high pressure system) is located in interior North America east of the Rocky Mountains (Fig. 2a). Clockwise advection results in relatively cold air and north to northwesterly flow into Cleveland. *Plains Return (PR)* occurs when the anticyclonic center drifts farther to the east, and is southwest of the Great Lakes (Fig. 2b), producing westerly to southwesterly air into Cleveland. *Appalachian Return (AR)* involves the occurrence of an anticyclonic system directly north or northeast of Cleveland (Fig. 2c) and an easterly to northeasterly air flow. *Gulf Return (GR)* flow results when high pressure is located over the southeastern United States or just off the Eastern seaboard (Fig. 2d). A relatively warm, moist southeasterly to southerly flow of air into the Lake Erie area with air originating from the Gulf of Mexico or Atlantic Ocean would result. *Frontal Gulf Return (FGR)* is characterized by tropical or subtropical return flow that is affected by convergence associated with a surface cold front that is approaching Cleveland (Fig. 2e). However, Cleveland remains in the "warm sector" in this situation, until frontal passage occurs. The defining criterion in this category is whether the presence of the front is actually influencing the weather of Cleveland to a significant enough extent so as to dominate the weather conditions at the time. *Frontal Overrunning (FOR)* occurs immediately after the passage of a frontal system through Cleveland, but while the weather properties indicate that Cleveland is still being influenced by the

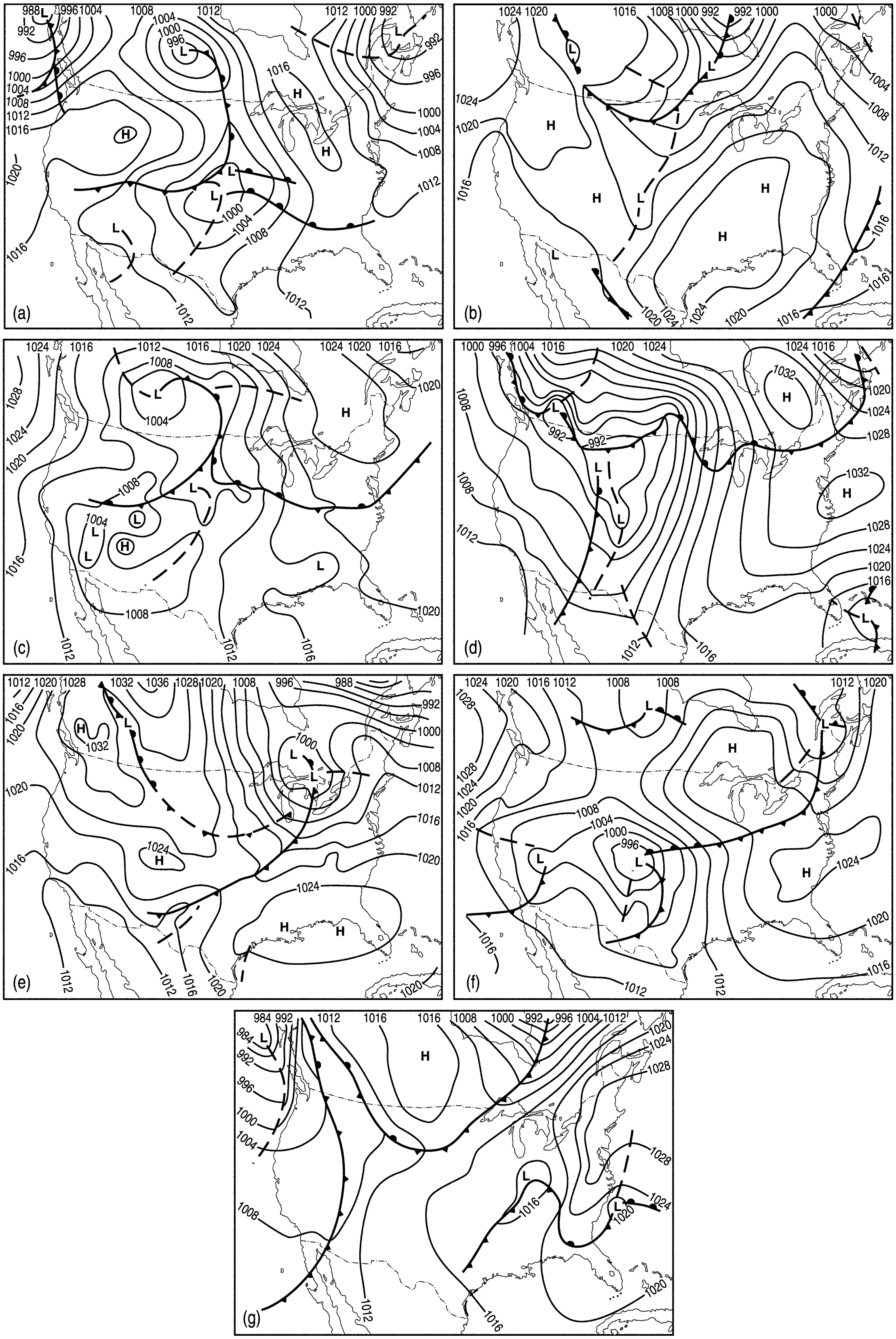

FIG. 2. Examples of the weather types for Cleveland, Ohio (Britton 1999). a) Continental High; b) Plains Return; c) Appalachian Return; d) Gulf Return; e) Frontal Gulf return; f) Frontal Overrunning; g) Occluded Low. More details on the properties of these weather types are provided in Britton (1999).

TABLE 1. Mean surface weather conditions at Buoy 45005, by Month. Dewpoint depression is from Southeast Shoal.

		Air Temp. (ºC)	Water Temp. (ºC)	Dewpoint Depression (ºC)	Sea Level Pressure (mb)	Wind Speed (ms^{-1})	Wave Height (m)	Visibility (km)
April	Mean	5.62	4.35	1.99	1,012.90	4.48	0.29	23.96
	Std Dev	1.77	1.14	2.12	5.81	1.93	0.36	17.60
	n	33	33	22	33	33	22	17
May	Mean	11.76	10.55	2.37	1,014.15	4.73	0.32	20.64
	Std Dev	2.57	2.22	2.01	6.84	2.36	0.35	16.66
	n	95	94	85	95	94	85	67
June	Mean	18.07	17.33	2.40	1,013.62	4.26	0.28	21.56
	Std Dev	3.11	2.87	1.65	4.44	2.43	0.26	17.44
	n	107	106	75	107	107	75	57
July	Mean	21.82	22.23	2.80	1,013.88	5.29	0.40	27.53
	Std Dev	2.08	1.49	1.63	4.32	2.28	0.33	16.23
	n	150	149	113	150	150	113	84
August	Mean	21.55	22.55	2.71	1,016.58	4.85	0.37	21.57
	Std Dev	2.23	1.54	1.68	3.91	2.27	0.31	15.26
	n	145	145	110	145	145	110	80
September	Mean	18.3	20.49	3.44	1,015.96	6.22	0.58	25.47
	Std Dev	2.94	1.48	1.44	6.34	2.64	0.36	15.67
	n	136	136	104	136	136	104	75
October	Mean	13.02	14.91	3.62	1,017.02	6.58	0.61	24.05
	Std Dev	2.91	2.12	1.52	7.23	2.4	0.39	15.10
	n	88	88	58	88	88	58	56
November	Mean	6.14	8.94	4.12	1,018.99	7.43	0.88	29.59
	Std Dev	4.45	2	1.31	9.31	2.89	0.49	14.70
	n	72	72	33	72	72	33	33
Overall	Mean	16.60	17.37	2.91	1,015.43	5.46	0.44	24.07
	Std Dev	6.02	5.77	1.74	6.14	2.6	0.38	16.15
	n	826	823	600	826	825	600	469

frontal zone (Fig. 2f). FOR weather can also occur as warm air overruns colder air at a warm front positioned to the south of Cleveland. Finally, *Occluded Low (OL)* is characterized by an area of low pressure in its dissipating stages, usually tracking northeast along the Eastern seaboard (Fig. 2g). Since the cyclone would usually be centered north or northeast of Cleveland, this weather type would probably generate a north to northwesterly wind and would produce atmospheric instability in the Cleveland area.

The period of overlap between weather-type classification and available meteorological data consists of 42 PR, 108 CH, 120 FOR, 39 AR, 163 GR, 179 FGR, and 5 OL days. It should be noted that these frequencies should not be used as indicators of the overall frequency of the weather patterns, since overlapping temperature and dew point data are unavailable for the cold season and various other days. Britton (1999) should be consulted for more complete descriptions of the circulation types and for frequencies of each of the weather types.

RESULTS AND DISCUSSION

Mean values of the input parameters are shown for May, July, and September in Table 1 and Figures 3a–3c, respectively. In May, the temperatures

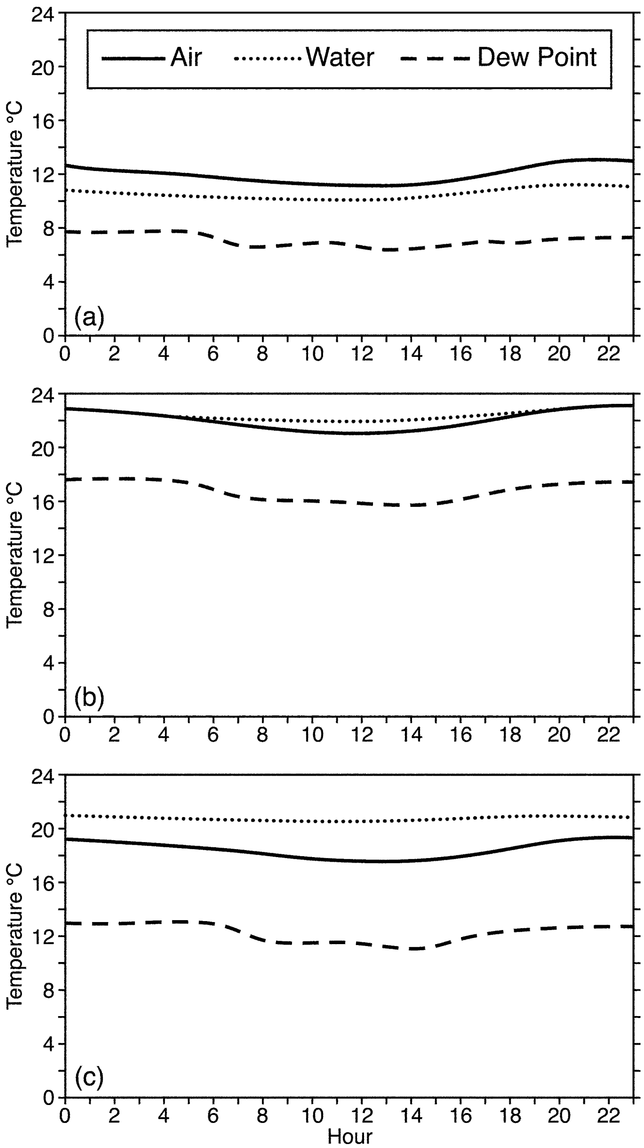

FIG. 3. Mean diurnal (in hours UTC) air, water, and dew point temperatures by month for western Lake Erie. a) May; b) July; and c) September.

shown are not unexpected, with the air warmer than the water throughout the daily cycle (Fig. 3a). By mid-summer (Fig. 3b), Lake Erie's water temperature approaches or even exceeds air temperature, especially beginning in the mid-afternoon hours. By late summer (Fig. 3c), the lake is considerably warmer than the air above it on a climatic average. As at most mid-latitude locations, the diurnal dew point curve is relatively flat, and shows significantly higher values in mid-summer than in May and September. Likewise, the differences in weather conditions by synoptic weather type are shown in Tables 2 and 3.

The monthly means and standard deviations of β by atmospheric stability class show some interesting features. During times of unstable atmospheric conditions (Fig. 4a), β seems to be somewhat higher than for values observed over tropical oceans (~0.10 (Oke 1987)) and Lake Pontchartrain (~0.10 – 0.12 (Rohli and Hsu 1999)), and this result is supported by Roll's (1965) observation of generally higher β at higher latitudes. A possible explanation for this phenomenon is that since the air immediately over the water is near saturation, Equation 3 can be approximated as

$$\beta = (C_p\ (T_{sea} - T_{air})\ )\ /\ (L_v\ (q_{sea} - q_{air})_{saturated}) \qquad (6)$$

and

$$\beta \approx (C_p/L_v)\ (dT/de_s) \qquad (7)$$

where e_s represents saturation vapor pressure. Since de_s/dT increases with higher temperatures, it is logical that β may be generally lower for subtropical areas than for midlatitude lakes such as Erie.

Consistent with this temperature relationship, for unstable conditions values generally decrease through early summer, and then exhibit a sharp increase later in the season (Fig. 4a). More specifically, the mean β during unstable conditions decreases slightly from 0.26 in April to 0.13 in July and August, and increases dramatically up to 0.52 in November. These results suggest that when the water is warmer than the air above it (as in unstable situations, and which are typical of late summer and fall), relatively little evaporation is occurring compared with tropical and subtropical marine environments. It is also likely that β over Erie is influenced by the adjacent landmass and polar air which flows out over the lakes early in the frost-free season (creating high environmental lapse rates and unstable conditions), thereby contributing to lower β values (Roll 1965) early in the season.

The sharp increase in autumn is not surprising because as T_{air} (and therefore e_s) drops, Q_e becomes limited by atmospheric saturation and is relatively less important than Q_h. Perhaps this result could also be related to the abundance of moisture and cloud cover under saturated conditions that acts to reduce the surface-to-atmosphere moisture gradient. Therefore, β increases. Such a scenario would be exacerbated during unstable conditions when Q_h would be strong in the upward direction. It is inter-

TABLE 2. Mean May weather conditions, by weather type (no OL days occurred in May), 1992–1995.

Weather Type		Air Temp. (°C)	Water Temp. (°C)	Dewpoint Depression (°C)	Sea Level Pressure (mb)	Wind Speed (ms^{-1})	Wave Height (m)	Visibility (km)
AR	Mean	13.4	13.7	3.1	1,023.2	7	0.3	1.8
	Std Dev	—	—	—	—	—	—	-
	n	1	1	1	1	1	1	1
CH	Mean	10.15	10.73	3.39	1,017.01	5.13	0.53	20.67
	Std Dev	2.46	1.84	1.74	3.71	2.93	0.43	20.62
	n	19	18	18	19	19	18	18
FGR	Mean	13.74	11.27	2.39	1,009.84	5.18	0.31	22.70
	Std Dev	2.05	2.13	2.06	6.14	2.03	0.29	14.15
	n	18	18	18	18	18	18	18
FOR	Mean	12.43	11.99	2.65	1,014.91	4.38	0.30	18.29
	Std Dev	2.42	1.36	1.14	6.64	1.78	0.19	18.98
	n	8	8	8	8	8	8	8
GR	Mean	12.83	11.92	1.31	1,019.55	3.24	0.13	18.42
	Std Dev	2.15	1.51	1.79	4.79	1.37	0.19	12.40
	n	16	16	16	16	16	16	16
PR	Mean	10.83	9.62	0.67	1,011.63	3.30	0.03	26.53
	Std Dev	0.76	1.03	1.63	5.58	0.10	0.08	20.43
	n	6	6	6	6	6	6	6

TABLE 3. Mean July Weather Conditions, by Weather Type (no OL days occurred in July), 1992–1995.

Weather Type		Air Temp. (°C)	Water Temp. (°C)	Dewpoint Depression (°C)	Sea Level Pressure (mb)	Wind Speed (ms^{-1})	Wave Height (m)	Visibility (km)
AR	Mean	21.23	22.37	2.53	1,019.4	3.97	0.33	11.75
	Std Dev	3.35	1.17	2.25	3.29	1.93	0.29	1.95
	n	3	3	3	3	3	3	3
CH	Mean	21.89	23.31	3.28	1,015.06	5.48	0.46	23.86
	Std Dev	1.84	1.19	1.07	4.3	2.54	0.29	21.54
	n	14	14	14	14	14	14	14
FGR	Mean	22.83	22.71	2.73	1,012.21	5.80	0.43	28.77
	Std Dev	1.91	1.44	1.83	2.69	1.72	0.39	14.18
	n	28	28	28	28	28	28	28
FOR	Mean	22.18	22.45	2.67	1,013.22	5.05	0.37	29.90
	Std Dev	1.78	1.62	1.91	2.83	2.31	0.34	15.94
	n	18	18	18	18	18	18	18
GR	Mean	22.29	22.43	2.77	1,016.26	4.89	0.29	24.70
	Std Dev	1.56	1.51	1.24	2.36	1.44	0.18	13.15
	n	15	15	15	15	15	15	15
PR	Mean	21.05	22.18	3.5	1,013.42	5.97	0.63	38.19
	Std Dev	2.41	2.12	0.57	2.51	3.16	0.34	18.74
	n	6	6	6	6	6	6	6

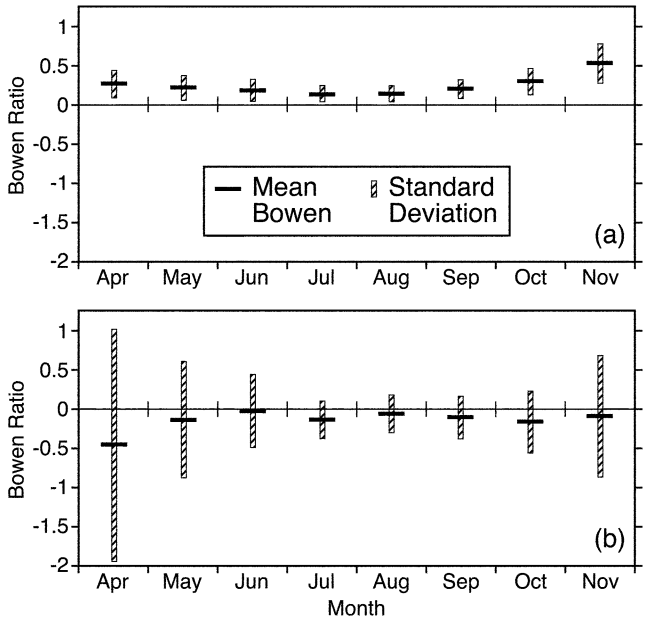

FIG. 4. Mean and standard deviation of β ***by atmospheric stability category: a) unstable atmosphere, b) stable atmosphere.***

esting that β still increases into late summer and early autumn despite the fact that these tend to be the times of maximum evaporation from Lake Erie (Croley 1995). This suggests that as the lakes warm, unstable conditions allow for relatively greater increases in Q_h than Q_e.

The computed values of β are used to generate a simple power curve for prediction of β based on air-sea temperature differences during hourly observations having unstable conditions. The resulting relationship of

$$\beta = 0.097(T_{sea} - T_{air})^{0.74} \quad (8)$$

is reasonably near that calculated based on calculations during unstable conditions made by Hsu (1999) in the East China Sea, near San Diego, and in the equatorial Atlantic Ocean, who found

$$\beta = 0.146\ (T_{sea} - T_{air})^{0.49} \quad (9)$$

and by Sadurham *et al.* (2001) for the northern Bay of Bengal, who calculated

$$\beta = 0.094\ (T_{sea} - T_{air})^{0.80} \quad (10)$$

During times of thermal stability, β becomes slightly negative (Fig. 4b), indicating that Q_h is directed from the atmosphere to the surface. Such a phenomenon is typical of the "oasis effect" (Oke 1987), whereby the atmosphere actually supplies sensible heat to the surface and a temperature inversion and subsidence are common. Variation about the mean values is very high during the spring and autumn months. In spring, it appears that in some circumstances during stable conditions, the atmosphere can heat the shallow water body effectively, resulting in a strongly negative Q_h and therefore a strongly negative β. For example, April β values in stable conditions average –0.45. However, during other times of stable spring conditions, the gradient of Q_e may also be slightly negative (*i.e.,* from the atmosphere to the near-surface lake). This would result in a relatively high positive β. An ice-covered lake in spring is the most obvious situation that would produce this scenario. The net effect, however, is still a negative mean through the season, varying from –0.45 in April to –0.03 in June. During near-neutral conditions (not shown), β remains near zero as would be expected when lake and air temperatures are nearly equal.

Diurnally, it is not surprising that β usually peaks in the afternoon hours and tends to increase as the warm season proceeds (Fig. 5a–b). It should be noted that the data in Figure 5a–b were compiled using the appropriate drag coefficient for the stability class of that observation, but the resulting β values shown include all stability classes. In addition, the number of observations becomes substantially lower than for the analyses in Figures 4a and 4b because of the segregation of the data by month and hour.

Minimum mean values of β occur in April at 0700 local apparent time (LAT) (-0.64, n = 34) and 2000 LAT (-0.73, n =30), and maxima occur in November at 1600 LAT (0.49, n = 74) (Fig. 5a–b). In general, mid-summer ratios fluctuate relatively little diurnally. For example, August values vary from 0.04 at 2300 LAT (n =164) to 0.10 at 1300 LAT, while ratios in April vary from the minima listed above up to 0.10 at 1400 LAT (n =30).

Table 4 shows the mean and standard deviation of β by weather type. Results of analysis of variance procedures suggest that the null hypothesis of no significant association between weather type and β is rejected. However, subsequent Scheffé's Simultaneous Confidence Interval Tests (Scheffé 1959) reveal that OL is the only weather type that differs significantly from any other weather types, but the very low number of observations is a likely cause of this result. The conclusion to be reached from

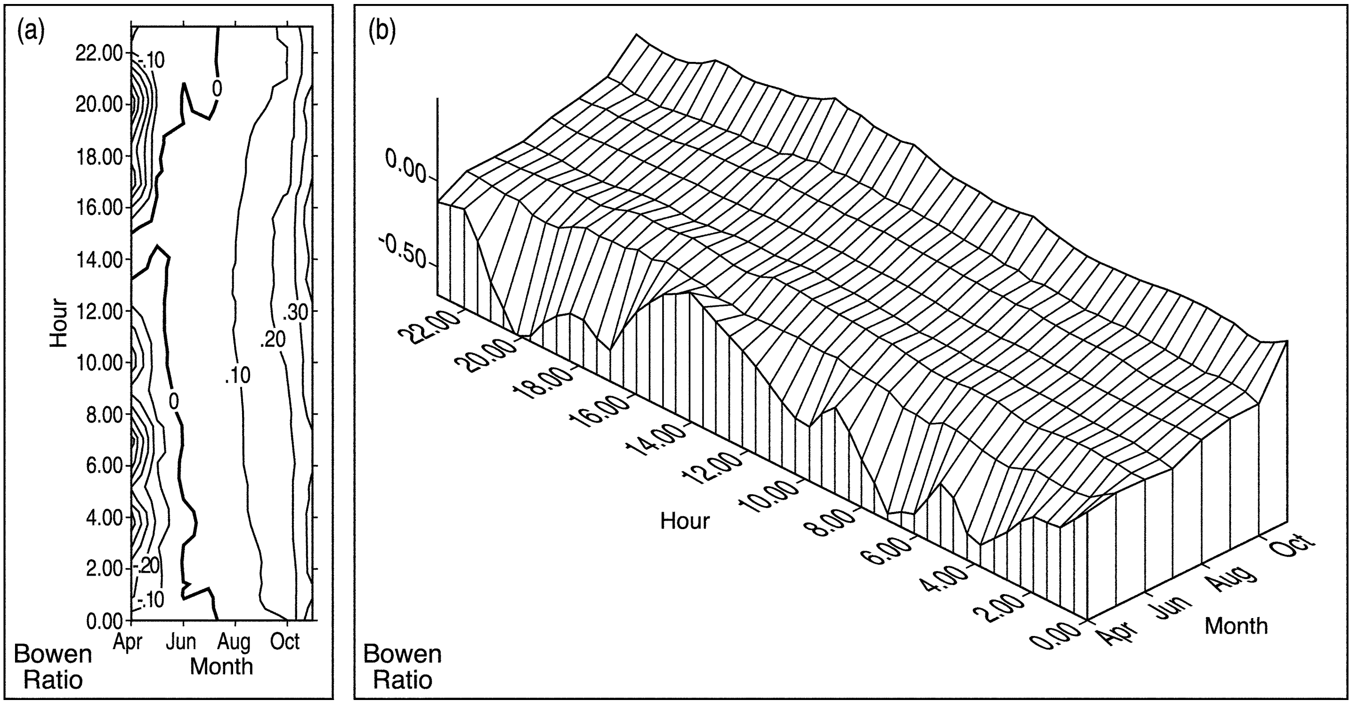

FIG. 5a–b. Mean diurnal cycle of β*, by month.*

TABLE 4. Mean and standard deviation of β*, by weather type*

Weather Type	N	Mean	Standard deviation
Continental High	108	.161	.179
Plains Return	42	.107	.350
Appalachian Return	39	.172	.101
Frontal Overrunning	120	.087	.283
Gulf Return	163	.109	.229
Frontal Gulf Return	179	.111	.338
Occluded Low	5	.402	.124

this analysis is that the within-group variability in β exceeds the between-group variability. Thus, it is unlikely that the synoptic weather situation alone is a valuable indicator of the amount of evaporation.

SUMMARY AND CONCLUSIONS

Results of this research suggest that the diurnal and seasonal cycles of β appear to be reasonable compared with field-measured values, even under different categories of atmospheric stability. More specifically, during unstable atmospheric conditions, β varies from approximately .15 to .30, and during times of thermal stability, β tends to have near-zero to slightly negative values.

Although some problems may exist with the use of these equations, the results seem to be generally useful for environmental assessments of an important water body, for which little atmospheric energy balance information was previously available. Such information is useful to those requiring estimates of the energy balance in that Q_h or Q_e can be calculated using field methods, while the other turbulent flux can be derived based on the known value and the derived β (Rohli and Hsu 1999). Future field research is needed to calibrate this equation with measured observations (Bello and Smith 1990), so that subsequent estimation can be done using the calibrated equation.

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