An Improved Aerodynamic Evaporation Technique for Large Lakes With Application to the International Field Year for the Great Lakes

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An improved bulk transfer technique was developed for large-lake evaporation based upon recent boundary layer research near the air-water interface. A variable bulk transfer coefficient, dependent upon atmospheric stability, is given as a function of the nondimensional wind speed gradient, the potential temperature gradient, and the Monin-Obukhov length. The technique, which requires the same data as the simplified mass transfer equation, can be readily applied to large lakes throughout the world. This technique has been applied to the Lake Ontario data set collected during the International Field Year for the Great Lakes. The inclusion of stability increases calculated evaporation during the unstable high-evaporation months and decreases calculated condensation during the stable late spring months to more realistic levels. Comparisons between the Lake Hefner mass transfer equation and the technique recommended here indicate that the mass transfer equation may overestimate Lake Ontario evaporation by approximately 20%.

INTRODUCTION

Evaporation from their water surfaces constitutes a major water loss from the Laurentian Great Lakes. This loss ranges from about 460 mm/yr on Lake Superior to about 900 mm/yr on Lake Erie. Because of their large surface areas (from 19,000 km^2 for Lake Ontario to 82,100 km^2 for Lake Superior) the lower atmospheric boundary layer behaves similarly to that existing over a large sea or ocean rather than a lake. The result is that evaporation becomes a fairly strong function of atmospheric stability. However, the majority of Great Lakes evaporation studies to date [*Richards and Irbe,* 1969; *Derecki* 1976] have been based upon mass transfer formulations from the classic Lake Hefner study [*U.S. Geological Survey,* 1954]. *Phillips* [1978] was one of the first investigators to include atmospheric stability effects on Great Lakes evaporation. He varied the bulk moisture coefficient as a step function dependent upon air-water temperature differences.

This paper describes an improved bulk transfer technique for large-lake evaporation based upon recent boundary layer research near the air-water interface. A variable bulk transfer coefficient, dependent upon atmospheric stability, is given as a function of the nondimensional wind speed, the potential temperature gradients, and the Monin-Obukhov length. The technique, which requires the same data as the simplified mass transfer equation, can be readily applied to large lakes throughout the world.

This technique has been applied to the Lake Ontario data set collected during the International Field Year for the Great Lakes. This data set differs from that described by *Phillips* [1978] in that it consists of measured rather than simulated overlake data.

THEORETICAL CONSIDERATIONS

The evaporation rates in this study were computed from the bulk aerodynamic evaporation formulation expressed as

$$E = \rho C_E(q_0 - q)U \tag{1}$$

where

- E evaporation rate;
- ρ air density;
- q specific humidity;
- q_0 saturation specific humidity at the water surface temperature;
- U wind velocity;
- C_E bulk evaporation coefficient.

At the current state of the art the bulk evaporation coefficient C_E is usually assumed to equal the sensible heat coefficient C_H, which has been extensively studied. This assumption is also made for this study.

The analysis begins by expressing the nondimensional wind speed and potential temperature gradients in the surface boundary layer, ϕ_1 and ϕ_2, respectively, as

$$\phi_1 = \frac{KZ}{U_*}\frac{\partial U}{\partial Z} \tag{2a}$$

$$\phi_2 = \frac{Z}{\theta_*}\frac{\partial \theta}{\partial Z} \tag{2b}$$

where

- K von Kármán's constant;
- Z vertical space coordinate;
- U mean wind speed;
- U_* friction velocity, equal to $(\tau/\rho)^{1/2}$;
- τ surface shear stress;
- ρ air density;
- θ potential temperature;
- θ_* scaling temperature, equal to $-(1/KU_*)(H/\rho C_p)$;
- H turbulent heat flux, equal to $-\theta_* U_* K\rho C_p$;
- C_p specific heat of air at constant pressure.

The gradients may be integrated, following *Panofsky* [1963], in the form

$$U = (U_*/K)[\ln(Z/Z_0) - \Psi_1] \tag{3}$$

$$\theta - \theta_0 = \beta\theta_*[\ln(Z/Z_0) - \Psi_2] \tag{4}$$

where

$$\Psi_1 = \int_0^{\epsilon} \frac{1 - \phi_1(\epsilon)}{\epsilon}\, d\epsilon \tag{5}$$

$$\Psi_2 = \int_0^{\epsilon} \frac{1 - \phi_2(\epsilon)/\beta}{\epsilon}\, d\epsilon \tag{6}$$

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and

Z_0 roughness length;
β value of ϕ_2 at neutral stability;
$\epsilon = Z/L$;
L Monin-Obukhov length, equal to $-U_*^3 C_p \rho \bar{\theta}/(KgH)$;
$\bar{\theta}$ mean absolute potential air temperature;
θ_0 potential temperature at Z_0.

The coefficient β is included in this derivation to allow evaluation of the equation developed by *Businger et al.* [1971], which contains a β value of 0.74 rather than the usually accepted value of 1.0.

The stability ranges for this study are defined as unstable, $Z/L < 0$; neutral, $Z/L = 0$; stable, $0 < Z/L < 1$; and strongly stable, $Z/L \geq 1$.

Businger et al. [1971] and *Dyer* [1974] recommend the following forms of ϕ_1 and ϕ_2 for unstable conditions:

$$\phi_1 = [1 - \gamma_1(Z/L)]^{-1/4} \tag{7}$$

$$\phi_2 = \beta[1 - \gamma_2(Z/L)]^{-1/2} \tag{8}$$

Carrying out the integration to obtain Ψ_1 and Ψ_2, following *Paulson* [1970], we obtain

$$\Psi_1 = 2 \ln [(1 + X_1)/2] + \ln [(1 + X_1^2)/2] - 2 \tan^{-1} X_1 + (\pi/2) \tag{9}$$

$$\Psi_2 = 2 \ln [(1 + X_2^2)/2] \tag{10}$$

where

$$X_1 = [1 - \gamma_1(Z/L)]^{1/4} \tag{11}$$

$$X_2 = [1 - \gamma_2(Z/L)]^{1/4} \tag{12}$$

The values of $\gamma_1 = \gamma_2 = 16$ and $\beta = 1$, as recommended by *Paulson* [1970] and *Dyer* [1974], were used in these computations. However, the impact of the *Businger et al.* [1971] recommendation that $\beta = 0.74$, $\gamma_1 = 15$, and $\gamma_2 = 9$ was also evaluated and will be discussed later.

For stable and strongly stable conditions, *Webb*'s [1970] recommendation that

$$\phi_1 = \phi_2 = 1 + \gamma_3(Z/L) \qquad Z/L < 1 \tag{13}$$

and

$$\phi_1 = \phi_2 = 1 + \gamma_3 \qquad Z/L \geq 1 \tag{14}$$

was used in this study.

The alternative form for ϕ_2 suggested by *Businger et al.* [1971],

$$\phi_2 = \beta + \gamma_3(Z/L) = \beta[1 + \gamma_4(Z/L)] \tag{15}$$

where

$$\gamma_4 = \gamma_3/\beta \tag{16}$$

was also evaluated.

Carrying out the integration of (13), (14), and (15), we obtain

$$\Psi_1 = \Psi_2 = -\gamma_3(Z/L) \qquad Z/L < 1 \tag{17}$$

$$\Psi_1 = \Psi_2 = -\gamma_3[1 + \ln (Z/L)] \qquad Z/L \geq 1 \tag{18}$$

$$\Psi_2 = \gamma_4(Z/L) \tag{19}$$

For neutral conditions, $\Psi_1 = \Psi_2 = 0$.

The relationship between Z_0 and U_* for a large overwater surface was obtained from *Charnock*'s [1955] relationship for neutral conditions, which states

$$U/U_* = (1/K) \ln (gZ/U_*^2) + C \tag{20}$$

Setting $C = -(1/K) \ln \alpha$, substituting into (20), and rearranging, we obtain

$$U/U_* = (1/K) \ln (gZ/\alpha U_*^2) \tag{21}$$

By comparison with (3) for $\Psi_1 = 0$,

$$Z_0 = \alpha U_*^2/g \tag{22}$$

Equation (21) was solved for α by using *Smith and Banke*'s [1975] findings:

$$C_D = (U_*/U)^2 = 1.62 * 10^{-3} \tag{23}$$

for $U = 15$ m/s at $Z = 10$ m, yielding a value for α of 0.0101.

The sensible heat coefficient is derived from the relationship

$$H = -\rho C_p K U_* \theta_* = -\rho C_p C_H(\theta - \theta_0)U \tag{24}$$

Substituting for θ_* in (4),

$$\theta_* = \frac{(\theta - \theta_0)}{\beta[\ln (Z/Z_0) - \Psi_2]} \tag{25}$$

and solving for C_H, we obtain

$$C_E = C_H = \frac{KU_*}{\beta U[\ln (Z/Z_0) - \Psi_2]} \tag{26}$$

where C_E is assumed to equal C_H.

Hicks [1976] summarizes eight field evaluations of von Kármán's K with the result that the most commonly reported value of K was 0.41, found by four investigators. Three investigations yielded values of 0.42, and one, that conducted by *Businger et al.* [1971], a value of 0.35. *Dyer* [1974] discusses some of the problems in interpreting results reported by *Businger et al.* [1971]. In this study a value of $K = 0.41$ is used in the computations. A subroutine using an iterative algorithm with known values of θ_0, θ, U, and Z was developed to solve for C_H and C_E. The values for the coefficients and constants used in the computations as well as findings by *Businger et al.* [1971] are summarized in Table 1.

TABLE 1. Summarized Coefficients and Constants for Determination of the Sensible Heat Coefficient C_H

Coefficient or Constant	Value Used in Study		*Businger et al.* [1971] Value
	Value	Reference	
γ_1	16	*Paulson* [1970]	15
γ_2	16	*Dyer* [1974]	9
γ_3	5.2	*Webb* [1970]	4.7
γ_4			6.4
K	0.41	*Hicks* [1976]	0.35
β	1.00		0.74
α	0.0101		0.0450

DATA COLLECTION SYSTEM

The data collection system consisted of the 21 meteorological buoys placed in Lake Ontario as part of the International Field Year for the Great Lakes (IFYGL) in 1972. The locations of the buoys are shown in Figure 1. The buoy data used in this study consisted of wind velocity, air and water surface temperatures, and dew point temperature or relative humidity. Typical U.S. and Canadian buoys are shown in Figure 2. Information on the meteorological sensors and cali-

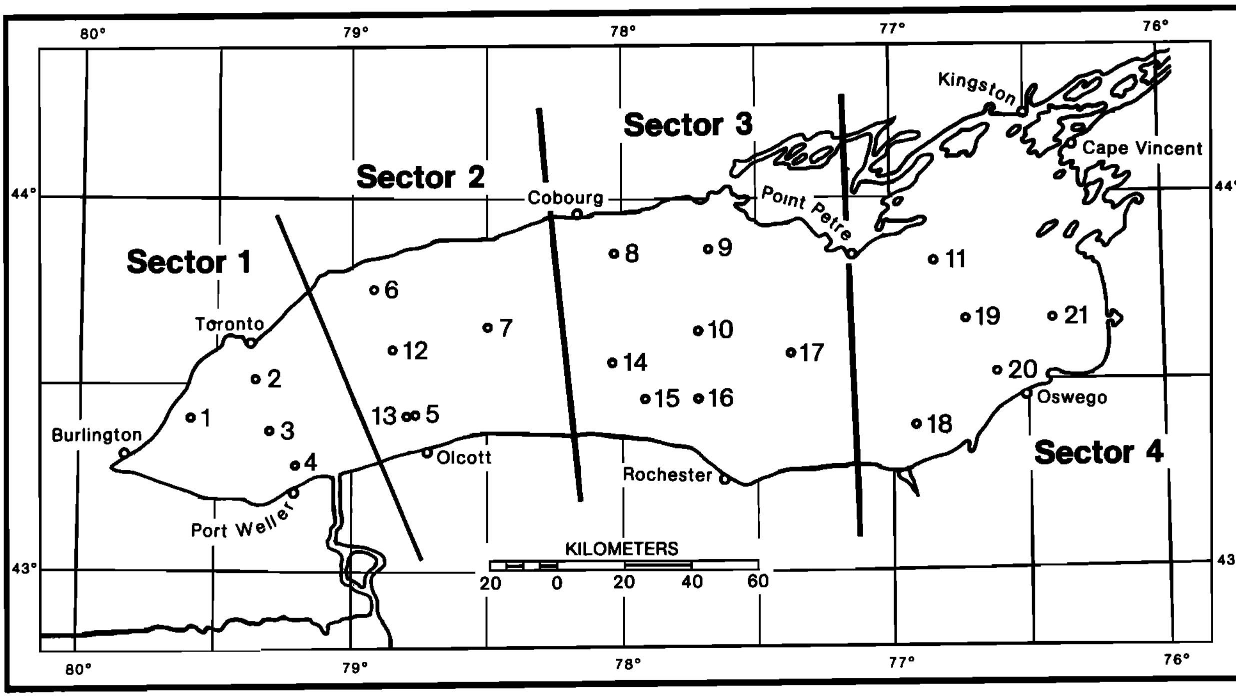

Fig. 1. Buoys and sectors used in IFYGL Lake Ontario analysis.

bration is given by *Foreman* [1976] for the U.S. system and by *Bryon* [1976] for the Canadian system.

The sensor heights were 3 m on the U.S. buoys and 4 m on the Canadian buoys. The wind velocity and the specific humidity and vapor pressure gradients were standardized at the 3-m level by

$$P_3 = P_4 \frac{\ln Z_3/Z_0}{\ln Z_4/Z_0} = 0.97P_4 \tag{27}$$

where P_3 and P_4 are the wind, specific humidity, or vapor pressure gradients referred to at the 3-m and 4-m levels, respectively, where Z_3 is 3 m, Z_4 is 4 m, and Z_0 is 0.01 cm.

The first buoys were placed in the lake in early April 1972, and the last buoys removed in mid-December 1972. This limited the analysis to 32 weeks. Another limitation was missing data from many of the platforms during the data collection period. The percent of available data ranged from a low of 34% in May to a high of 63% in September. The collected data

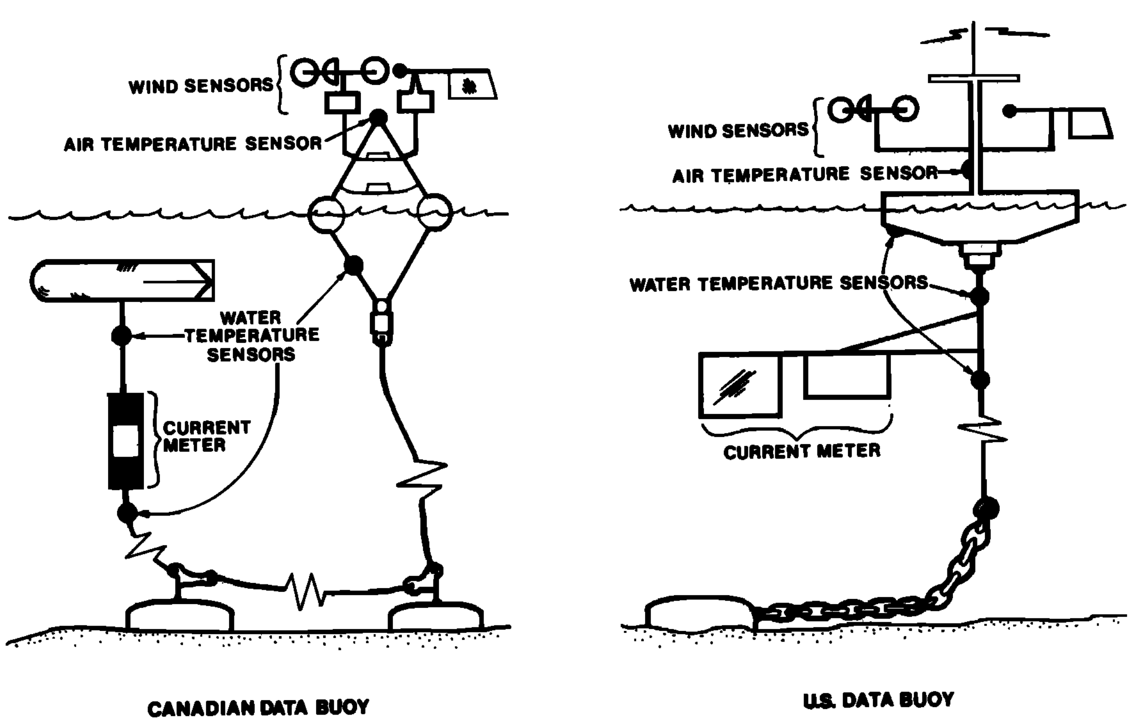

Fig. 2. U.S. data buoy and Canadian data buoy [from *Pickett*, 1975].

TABLE 2. Differences in Daily Evaporation Between Equations (30) and (31)

	June–July		October–November	
	Buoy 5	Buoy 11	Buoy 5	Buoy 11
Total evaporation difference, mm	4.0	−0.7	2.7	−1.5
Mean daily difference, mm	0.07	−0.01	0.05	−0.03
Standard deviation, mm	0.18	0.11	0.33	0.22
Absolute maximum difference, mm	0.7	0.4	1.3	0.6

were also biased toward the eastern and central areas of the lake.

Temporal and Spatial Averaging

The data were collected at 6-min sampling intervals from the U.S. system and at 10-min intervals from the Canadian system. These data were then combined into hourly and daily averages providing the framework for the computations. The desired averaging period for the study was daily, with weekly and monthly evaporation estimates being generated from that. The degradation of the computations using daily averaged rather than hourly averaged meteorological parameters was investigated with data from buoys 5 and 11. Daily evaporation was computed from

$$E = \sum_{i=1}^{24} \rho C_E(q_s - q_a)U \cdot 3600 \tag{28}$$

$$E = \rho \bar{C}_E(\bar{q}_s - \bar{q}_a)\bar{U} \cdot 86{,}400 \tag{29}$$

where q_s is the saturation specific humidity at the water surface temperature, q_a is the specific humidity of the air, and the overbars refer to daily average values.

The results, summarized in Table 2 for both the June–July low-evaporation regime and the October–November high-evaporation regime, show that no degradation occurs from the use of daily averaged meteorological parameters. Daily averaged parameters were then used for the remainder of the study.

Two methods for spatial averaging of daily evaporation were investigated for use in the study. Straight averaging was not considered as it would introduce a bias toward the western end of the lake due to the unequal distribution of the data platforms. The first method consisted of dividing the lake into the four sectors shown in Figure 1. Weighting factors based upon sector areas were then determined for each sector. The daily evaporations computed from the reporting buoys.

The second procedure consisted of using the grid square technique to develop equivalent Thiessen polygon weighting factors for each reporting buoy on a daily basis. Two grid sizes were investigated, a coarse 4 × 4 km grid and a finer 2 × 2 km grid. Both grids gave identical results. Considering the entire lake as one sector reduced the May–November evaporation by 12%. Table 3 gives weekly evaporation comparisons between the grid square procedures and the four-sector analysis. The analysis shows no significant differences between the two procedures. As the four-sector procedure is approximately 7 times less expensive than the grid square technique, it was chosen for use in the remainder of the calculations.

Sensitivity Analysis

The relative sensitivity and error variance was determined by a modification of the sensitivity function reported by *Coleman and DeCoursey* [1976]. The usual form of a sensitivity function, expressed as a ratio of the evaporation, would be

$$S_i = \frac{1}{E}\frac{\partial E}{\partial X_i} \tag{30}$$

where S_i is the sensitivity and X_i are the independent parameters. For the evaluation of complex formulations, (30) is usually placed in finite difference form as

$$S_i = \frac{1}{E}\frac{\Delta E}{\Delta X_i} \tag{31}$$

where $\Delta E/\Delta X$ is the change in evaporation per unit change in X_i. The main deficiency in (30) and (31) is that they do not consider the magnitude or variability of the variables X_i. This difficulty is overcome by defining the relative sensitivity, or relative importance, Ψ_R, as

$$\Psi_{Ri} = \frac{\Delta E}{\Delta X_i}\frac{(X_{i\max} - X_{i\min})}{E} \tag{32}$$

where $(X_{i\max} - X_{i\min})$ defines the range of the variable X_i. Equation (31) differs from that of *Coleman and DeCoursey* [1976] in that instead of the parameter range $(X_{\max} - X_{\min})$ they substitute $(X - X_{\min})$.

Coleman and DeCoursey's [1976] error variance function

$$E[V(X)] = \sum_{i=1}^{n}\left(\frac{\Delta E}{\Delta X_i}\right)^2 \operatorname{Var} X_i \tag{33}$$

TABLE 3. Weekly Evaporation Comparison Between Grid Square and Four-Sector Procedures

Week Ending Date, 1972	Weekly Evaporation, mm: 4-km Grid, A	2-km Grid, B	Four-Sector, C	Evaporation Difference, mm: $A - B$	$A - C$
April 16	2	2	0	0	2
April 23	2	2	2	0	0
April 30	3	3	3	0	0
May 7	−1	−1	−1	0	0
May 14	2	2	2	0	0
May 21	−1	−1	−1	0	0
May 28	0	0	0	0	0
June 4	−1	−1	−1	0	0
June 11	2	2	1	0	1
June 18	−1	−1	−1	0	0
June 25	−2	−2	−2	0	0
July 2	−1	−1	0	0	−1
July 9	3	3	3	0	0
July 16	0	0	0	0	0
July 23	2	2	2	0	0
July 30	13	13	13	0	0
Aug. 6	13	13	13	0	0
Aug. 13	17	17	17	0	0
Aug. 20	10	10	10	0	0
Aug. 27	5	5	5	0	0
Sept. 3	11	11	10	0	1
Sept. 10	27	27	27	0	0
Sept. 17	16	16	16	0	0
Sept. 24	28	28	29	0	−1
Oct. 1	16	17	17	−1	−1
Oct. 8	18	18	18	0	0
Oct. 15	40	40	40	0	0
Oct. 22	35	36	35	−1	0
Oct. 29	13	13	14	0	−1
Nov. 5	21	21	21	0	0
Nov. 12	15	15	15	0	0
Nov. 19	27	27	27	0	0
Nov. 26	31	31	32	0	−1
Dec. 3	31	31	31	0	0
Dec. 10	22	22	24	0	−2

TABLE 4. Sensitivity and Error Variance Analysis, May–November

Parameter, X	Ψ_R	Standard Error	$E[V(X)]$
Wind speed	2.38	1 m/s	141
Dew point temperature	5.93	0.5°	26
Water surface temperature	5.69	0.5°	58
Air temperature	0.88	0.5°	1
Pressure	0.0		
C_E	0.90	0.10×10^{-3}	24

where Var X_i is taken as the square of the standard error of X_i, was also used in assessing the possible error contributions from each of the parameters. The results are given in Table 4. The most sensitive parameters are seen to be the water surface and dew point temperatures. However, the greatest potential error, indicated by the error variance, is in the wind speed followed by the water surface temperature.

WIND SPEED AND STABILITY ANALYSIS

The effect on the evaporation computations of the dependence of C_E on wind speed and stability was also investigated. The two average bulk coefficients C_E of 1.43×10^{-3} and 1.5×10^{-3} recommended by *Donelan et al.* [1974] and *Holland et al.* [1979], respectively, were compared with the recommended procedure. The variation of evaporation with the variable Z_0 but without stability effects was determined by running the recommended procedure with $\Psi_1 = \Psi_2 = 0$. A comparison is given in Table 5 on a monthly basis. The effect of the variable Z_0 is seen to be an evaporation increase during the higher-evaporation periods and a decrease during the lower-evaporation periods. These differences were further accelerated by the inclusion of the stability considerations. The inclusion of stability (Table 5, column 5) increased the May–November value by approximately 9%. It also reduced the calculated condensation in May and June to more reasonable values.

Table 5 also provides comparisons with the procedures used by *Phillips* [1978] and *Businger et al.* [1971]. Phillips' stability criteria are based solely upon air-water temperature differences. The procedure applied by Businger et al. has been discussed earlier. *Phillips*' [1978] procedure yields approximately 7% less evaporation than the recommended procedure, with the main differences appearing in the unstable months of October and November. *Phillips*' [1978] procedure gives results comparable to those of the recommended procedure without stability effects.

The procedure developed by *Businger et al.* [1971] yields approximately 34% greater evaporation than the recommended procedure. It is given here only for comparison as there is still controversy concerning its relevancy and assumptions [*Dyer,* 1974].

COMPARISON WITH MASS TRANSFER FORMULATIONS

The bulk aerodynamic procedure is closely related to the classic mass transfer formulation [*U.S. Geological Survey,* 1954]

$$E = M(e_s - e_a)U \tag{34}$$

where M is the mass transfer coefficient. Equation (34) can be written in terms of the vapor pressure as

$$E = C_E\rho\left(\frac{0.622e_s}{P - 0.378e_s} - \frac{0.622e_a}{P - 0.378e_a}\right)U \tag{35}$$

where e_s is the saturated vapor pressure at the water surface temperature and e_a is the vapor pressure of the ambient air. Equation (35) can be factored to obtain

$$E = \frac{C_E\rho 0.622}{P - 0.378e_s}(e_s - Re_a)U \tag{36}$$

where

$$R = (P - 0.378e_s)/(P - 0.378e_a) \tag{37}$$

The average value of R was computed by substituting typical Lake Ontario values for P, e_s, and e_a of 1020, 21, and 17 mbar, respectively, into (37), obtaining an R value of 0.999. The minimum value of R calculated from the overlake data was 0.996. Thus no significant error is introduced by assuming R to be 1.0.

The effect of neglecting the pressure correction term of $0.378e_s$ was also analyzed with the above mentioned values and found to be less than 1%, which is well within the limits of the data. Thus the mass transfer coefficient M is

$$M = 0.622\rho(C_E/P)86400 = 53741\rho(C_E/P) \tag{38}$$

Equation (38) shows a dependence of M on the air density ρ and the pressure P as well as on the bulk evaporation coefficient C_E. For an average pressure and air density of 1020 mbar and 1.25 kg/m^3, respectively, the mass transfer coefficient is given by

$$M = 66.05C_E \tag{39}$$

For a constant value of $C_E = 1.5 \times 10^{-3}$, as discussed earlier, M becomes 0.099 at the 3-m level rather than 0.124, the corresponding classic Lake Hefner [*U.S. Geological Survey,* 1954] value currently used in most Great Lakes studies. The use of the 0.124 coefficient gives a May–November total computed evaporation of 455 mm, an increase of 19% over the recommended procedure. The Lake Hefner coefficient overestimated the evaporation or condensation during the entire period. The bulk transfer coefficient corresponding to the Lake Hefner mass transfer coefficient is 1.9×10^{-3}.

TABLE 5. Comparison of Evaporation Procedures: Monthly Evaporation in Millimeters

	Constant C_E		$\alpha = 0.0101$,	Recommended	*Phillips*	*Businger et al.*
	1.43×10^{-3}	1.5×10^{-3}	$\Psi_1 = 0$, $\Psi_2 = 0$	$\alpha = 0.0101$	[1978]	[1971]
May	−1	−1	−1	0	−1	1
June	−7	−7	−6	−2	−5	−2
July	21	22	19	20	21	25
Aug.	52	54	47	49	51	64
Sept.	86	91	85	90	86	118
Oct.	101	106	109	118	108	159
Nov.	89	93	99	108	96	147
Total	341	358	352	383	356	512

TABLE 6. Computed 1972 Weekly Evaporation

Week Ending Date	Evaporation, mm
April 30	7
May 7	−1
May 14	2
May 21	−1
May 28	0
June 4	−1
June 11	1
June 18	−1
June 25	−2
July 2	0
July 9	3
July 16	0
July 23	2
July 30	13
Aug. 6	13
Aug. 13	17
Aug. 20	10
Aug. 27	5
Sept. 3	10
Sept. 10	27
Sept. 17	16
Sept. 24	29
Oct. 1	17
Oct. 8	18
Oct. 15	40
Oct. 22	35
Oct. 29	14
Nov. 5	21
Nov. 12	15
Nov. 19	27
Nov. 26	32
Dec. 3	31

Field Year Evaporation

The monthly and weekly evaporation estimates during the IFYGL are given in Tables 5 (column 5) and 6, respectively. They show two evaporation regimes in existence during the portion of 2 field years when overwater data were available. The first, a low-evaporation regime between April and July, accounted for 9 mm of evaporation. The second regime, a high-evaporation period between August and December, accounted for 390 mm of evaporation, 98% of the total. The daily variations in evaporation were similar to those reported by *Phillips* [1978].

Conclusions

This paper has described an aerodynamic technique that includes the variation of evaporation with atmospheric stability for large lakes. The computer algorithm that computes the variable bulk transfer coefficient is very efficient and uses the same meteorological variables required for the simplified mass transfer formulation. The inclusion of stability in the technique increases the calculated evaporation during the unstable high-evaporation months and decreases the calculated condensation during the stable late spring months to more realistic levels. The computations with the IFYGL data base show no degradation resulting from the use of daily averaged rather than hourly averaged meteorological variables.

Comparisons between the recommended and Lake Hefner mass transfer equation indicate that the mass transfer equation may overestimate Lake Ontario evaporation by approximately 20%. Similar overestimates are likely on the other Great Lakes. The two equations would yield more comparable results at higher wind speeds than are normally encountered on the Great Lakes.

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References

Bryon, J. W. (Ed.), *Final Canadian Data and Information Catalogue, International Field Year for the Great Lakes,* 1976.

Businger, J. A., J. C. Wyngaard, Y. Izumi, and E. F. Bradley, Flux profile relationships in the atmospheric surface layer, *J. Atmos. Sci., 28,* 181–189, 1971.

Charnock, H., Wind stress on a water surface, *Quart. J. Roy. Meteorol. Soc., 81,* 639–640, 1955.

Coleman, G., and D. G. DeCoursey, Sensitivity and model variance analysis applied to some evaporation and evapotranspiration models, *Water Resour. Res., 12(5),* 873–879, 1976.

Derecki, J. A., Multiple estimates of Lake Erie evaporation, *J. Great Lakes Res., 2(1),* 124–149, 1976.

Donelan, M. A., F. C. Elder, and P. F. Hamblin, Determination of the aerodynamic drag coefficient from wind set-up (IFYGL), *Proc. Conf. Great Lakes Res., 17th,* 778–788, 1974.

Dyer, A. J., A review of flux-profile relationships, *Boundary Layer Meteorol., 7,* 363–372, 1974.

Foreman, J., IFYGL physical data collection system: Description of archived data, *Tech. Rep. EDS 15,* 175 pp., Nat. Oceanic and Atmos. Admin., Washington, D. C., 1976.

Hicks, B. B., Reply, *Boundary Layer Meteorol. 10,* 237–240, 1976.

Holland, J. Z., F. C. Elder, W. Y. Chen, and J. A. Almazan, *Atmospheric Boundary Layer in IFYGL—The International Field Year for the Great Lakes,* National Oceanic and Atmospheric Administration, Rockville, Md., 1979.

Panofsky, H. A., Determination of stress from wind and temperature measurements, *Quart. J. Roy. Meteorol. Soc., 89,* 85–94, 1963.

Paulson, C. A., The mathematical representation of wind speed and temperature profiles in the unstable atmospheric surface layer, *J. Appl. Meteorol., 9,* 857–861, 1970.

Phillips, D. W., Evaluation of evaporation from Lake Ontario during IFYGL by a modified mass transfer equation, *Water Resour. Res., 14(2),* 197–205, 1978.

Pickett, R. L., Intercomparison of Canadian and U.S. automatic data buoys, *Mar. Technol. Soc. J., 9,* 20–22, 1975.

Richards, T. L., and J. G. Irbe, Estimates of monthly evaporation losses from the Great Lakes 1950–1968 based on the mass transfer technique, *Proc. Conf. Great Lakes Res., 12th,* 684–701, 1969.

Smith, S. D., and E. G. Banke, Variation of the sea surface drag coefficient with wind speed, *Quart. J. Roy. Meteorol. Soc., 101,* 665–673, 1975.

U.S. Geological Survey, Water loss: Water-loss investigations: Lake Hefner studies, technical report, *U.S. Geol. Surv. Prof. Pap., 269,* 158 pp., 1954.

Webb, E. K., Profile relationships the log-linear range, and extension to strong stability, *Quart. J. Roy. Meteorol. Soc., 96,* 67–90, 1970.

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