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Operational Estimates of Lake Superior Evaporation Based on IFYGL Findings

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Monthly evaporation from Lake Superior was determined for individual years of a 34-year period, 1942–1975, by an improved mass transfer method. This method permits timely evaporation estimates from readily available land-based meteorological data and represents the most practical approach for determining operational evaporation estimates for the Great Lakes. Method improvements consist of refinements in the mass transfer coefficient and the land-to-lake data adjustments derived from extensive investigations conducted on Lake Ontario during the International Field Year for the Great Lakes. The mass transfer coefficient and data adjustments are based on atmospheric stability considerations applicable to Lake Superior. Because of extensive ice cover on the lake, the standard overwater mass transfer results were also adjusted for the effects of ice cover during winter. The mass transfer evaporation estimates are verified by the water budget determinations, which for Lake Superior offer firm estimates of evaporation but are impractical for operational applications because of long delays in the availability of data. In contrast to the other Great Lakes, all hydrologic components of the Lake Superior water budget are of the same order of magnitude, with comparable errors, eliminating the possibility of large residual errors in computed evaporation. Evaporation values as determined by the two methods agree reasonably well for both seasonal distribution and the annual total, with the resulting long-term annual value of approximately 500 mm. The ice cover adjustment reduced the average annual mass transfer overwater evaporation by 13% and produced much better agreement with the water budget seasonal distribution and annual values. Generally, the ice cover reduction of evaporation could be estimated by reducing the lake area by appropriate ice cover.

INTRODUCTION

Evaporation from Lake Superior removes approximately half a meter of water from the lake surface annually and represents a major water loss. This water loss has an important effect on various aspects of lake hydrology dealing with the hydrologic water balance, lake levels, and their regulation. Accurate and timely determination of lake evaporation is needed for operational applications involving these aspects of lake hydrology. Since evaporation is basically a cooling process, which constitutes a transfer of both mass and heat or energy across the air-water interface, evaporation rates can be calculated from related mass transfer and mass or energy balance determinations. Because of limitations imposed by the available data, only the first two methods (mass transfer and water balance) are normally used to compute Great Lakes evaporation. Data required to determine energy fluxes across the air-water interface are generally not available for the Great Lakes for any appreciable period of time.

All hydrologic components for Lake Superior are of the same order of magnitude and similar accuracy, greatly enhancing the reliability of water budget evaporation, but this method is not practical for operational applications because of long delays in the availability of required data. The most practical approach for operational evaporation estimates is the mass transfer method, where evaporation is determined from readily available climatological data. Basic climatological data for the Great Lakes are restricted to land stations located around the lakes and require adjustments to reflect overwater conditions. Because of their large surface areas and great depths, the Great Lakes have a tremendous heat storage capacity, which considerably modifies the overwater climate. This is particularly true for Lake Superior, which has a surface area of 82,100 km^2 and an average depth of 150 m (Figure 1). The required adjustments for variations in the atmospheric stability over land and water areas for various parameters have been refined on Lake Ontario during the International Field Year for the Great Lakes (IFYGL) and permit improved mass transfer determinations from the available lake perimeter meteorological data.

One of the primary objectives of intensive field investigations conducted during IFYGL was to improve estimates of Great Lakes evaporation. Using IFYGL results, *Phillips* [1978] and *Quinn* [1979] have presented improved mass transfer techniques for Lake Ontario which include atmospheric stability effects. This paper tests the IFYGL findings on Lake Superior, a much bigger lake subject to a more severe climate, and describes an improved mass transfer technique, based on atmospheric stability considerations, that can be applied to any large lake in different climatic conditions. Because of extensive ice cover on Lake Superior during winter, the method includes ice cover reduction of standard overwater mass transfer evaporation to indicate actual winter evaporation from the lake.

The period of record employed in the study, 1942–1975, was determined by the availability of generally homogeneous climatological data. The year 1942 corresponds to a general relocation of first-order meteorological stations (wind, air temperature, and humidity) from city to airport locations.

MASS TRANSFER METHOD

The mass transfer method of computing evaporation is based on the removal of water vapor from the lake surface by turbulent diffusion and is considered to be a function of the wind speed and the vapor pressure difference between saturation vapor pressure at the surface and ambient air vapor pressure at some predetermined level. The mass transfer equation used to compute Great Lakes evaporation during the past two decades represents a modification of the classic Lake Hefner equation [*U.S. Geological Survey*, 1954, 1958], which was ad-

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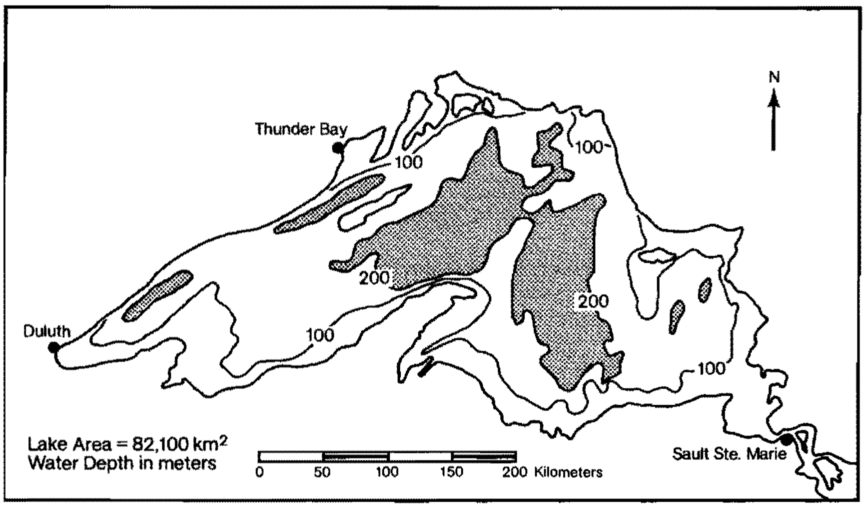

Fig. 1. Lake Superior Basin.

justed to 8 m. Expressed in its basic form, for SI units, the equation is

$$E = M(e_s - e_a)U \quad (1)$$

where

- E evaporation rate, mm d^{-1};
- M mass transfer coefficient, mm $d^{-1}(\text{mb m s}^{-1})^{-1}$;
- e_s saturation vapor pressure at lake surface; temperature, mb;
- e_a vapor pressure of ambient air, mb;
- U wind speed of ambient air, m s^{-1}.

The problem of nonrepresentative overlake climatological data was overcome by incorporating land-to-lake data adjustment terms. Variable land-to-lake adjustments for wind and air vapor pressure, based on air stability and overwater fetch criteria, were developed by *Phillips and Irbe* [1978] from the extensive IFYGL data base on Lake Ontario. These adjustments, expressed as lake/land wind ratios and land-lake air and dew point temperature differences, are grouped into five ranges of atmospheric stability and lengths of overwater fetch for six wind speed classes. Various atmospheric stability conditions (very stable, stable, neutral, unstable, and very unstable) are determined by the stability index, defined as air-water temperature difference. The stability index is determined from readily available land-based air temperatures. From the results of *Phillips and Irbe* [1978] a set of adjustment equations for all fetches was developed for Lake Superior [*Derecki,* 1980] and is used in the present study to obtain adjustments based on air stability conditions during each month. Separate adjustments have been developed for the overwater and overice atmospheric stability conditions to permit independent evaluation of the ice cover effects on various parameters.

Because of extensive ice over on Lake Superior, winter evaporation as computed by the standard mass transfer method for open water conditions may be considerably overestimated. The ice cover reduction of evaporation from Lake St. Clair, with comparable ice cover, was found to be 100 mm yr^{-1} [*Derecki,* 1979]. The ice cover reduction of evaporation was included by considering both open water and ice-covered areas of the lake during winter. Partial suppression of evaporation by ice cover was evaluated by determining ice cover effects on air stability (wind and temperatures) and vapor pressure. The stability index over ice for these evaluations was determined from ice surface temperatures. Separately determined overice values of evaporation were combined with the standard overwater data to produce overlake values reflecting actual lake surface conditions by using the extent of ice cover and areally weighting the two sets of values (overice and overwater). As indicated later by the evaporation results, the ice cover reduction of evaporation could be estimated by simply reducing the lake area by the applicable ice cover. However, separate computations for the overwater and overice conditions have been conducted in the Lake Superior study to permit evaluation of the ice cover effects on the input data and the mass transfer coefficient; the ice cover effect on these parameters is discussed in more detail by *Derecki* [1980].

MASS TRANSFER COEFFICIENT

The mass transfer coefficient used traditionally in Great Lakes evaporation studies represents the Lake Hefner calibrated constant, adjusted to 8 m (0.097 for mm d^{-1}). Because of large differences in lake size and climatic conditions, the atmospheric stability over Lake Hefner and the Great Lakes differs considerably, both in magnitude (strength) and seasonal variation. Using the classical approach of correlation between the mass transfer product and water budget evaporation, *Derecki* [1976] showed that the Lake Hefner constant is applicable to Lake Erie (0.097 versus 0.100). However, in subsequent extensive evaporation studies conducted on Lake Ontario during IFYGL, *Quinn and den Hartog* [1979] obtained considerably lower coefficient values with significant seasonal variation. *Quinn* [1979] developed a variable mass transfer coefficient, based on atmospheric stability, and presents an iterative algorithm for its derivation from the same meteorological variables that are required for normal mass transfer computations. This approach, used in the present study, defines the mass transfer coefficient as follows:

$$M = 0.622\rho(C_E/p)86400 = 53741\rho(C_E/p) \quad (2)$$

where

- M mass transfer coefficient, mm $d^{-1}(\text{mb m s}^{-1})^{-1}$;
- ρ air density, 1.25 kg m^{-3};
- C_E bulk evaporation coefficient;
- p atmospheric pressure, mb.

The above relationship shows that the mass transfer coefficient is dependent on the air density and pressure, and the latent heat flux. For average values of air density and applicable atmospheric pressure (1000 mb) the above equation for Lake Superior may be reduced to

$$M = 67.18\ C_E \tag{3}$$

Derivation of the bulk transfer coefficient for latent heat flux (C_E), dependent on atmospheric stability, was based on the analysis of nondimensional wind speed and potential temperature gradients in the surface boundary layer. The analysis involved determinations of frictional velocity, roughness length, Monin-Obukhov stability length, and stability functions for momentum and sensible heat to derive bulk transfer coefficients for momentum (drag) and sensible heat. Assuming that bulk transfer coefficients for sensible and latent heat fluxes are equal, the evaporation coefficient was obtained from the equation

$$C_E = C_H = \frac{KU_*}{U[\ln (Z/Z_0) - \Psi]} \tag{4}$$

where

C_E bulk transfer coefficient for latent heat;
C_H bulk transfer coefficient for sensible heat;
K von Karman's constant, 0.41;
U_* friction velocity, m s^{-1};
U wind speed, m s^{-1};
Z reference height, m;
Z_0 roughness length, m;
Ψ stability function for sensible heat.

Separate stability functions were determined for different atmospheric stability conditions. The stability ranges were defined by the reference height/Monin-Obukhov stability length relationship (Z/L) as

unstable

$$Z/L < 0$$

neutral

$$Z/L = 0$$

stable

$$0 < Z/L < 1$$

strongly stable

$$Z/L \geq 1$$

Known values of air and water surface temperatures, wind speed, and reference height were used to determine the bulk transfer coefficient. The reference height in the present study was standardized at 8 m (Z = 8 m) for all applications (mass transfer coefficient and meteorological data). More detailed information on the analysis and determinations of friction velocity, roughness length, stability length, and stability functions for different conditions is contained in *Quinn*'s [1979] paper. This information is needed to calculate the mass transfer coefficient for Lake Superior, but is omitted to eliminate extensive duplication.

The resulting Lake Superior mass transfer coefficient M is summarized in Table 1, which shows average monthly and annual values for the overwater stability conditions and comparable values for the approximate coefficient M_8 and the Lake Hefner constant. The average annual value of 0.071 for the Lake Superior coefficient is much lower than the 0.097 Lake Hefner value, but this large difference does not reflect actual effects on computed evaporation. During the more sensitive high evaporation season, the Lake Superior coefficient is generally close to 0.100 and agrees reasonably well with the Lake Hefner value. Still, the use of the Lake Hefner coefficient would tend to underestimate Lake Superior evaporation during the high evaporation season and overestimate it during the low evaporation season. Seasonal variation in the mass transfer coefficient is very large, increasing from a low value of 0.030 in June to a high of 0.105 in January. Reduction of the overwater coefficient due to ice cover is significant during winter. The winter overice values of the coefficient are reasonably constant (approximately 0.07) and agree closely with the perimeter values since meteorological conditions over ice and snow surfaces are similar; the overice coefficients could be estimated from perimeter data [*Derecki,* 1980]. The overice coefficients attain major importance during February and March, the months of extensive ice cover.

TABLE 1. Average Values for Lake Superior Mass Transfer Coefficient (Overwater), 1942–1975

Month	IFYGL Coefficient M	IFYGL Approximate M_8	Lake Hefner Coefficient
Jan.	0.105	0.082	0.097
Feb.	0.103	0.082	0.097
March	0.092	0.079	0.097
Arpil	0.058	0.071	0.097
May	0.036	0.068	0.097
June	0.030	0.068	0.097
July	0.032	0.069	0.097
Aug.	0.044	0.068	0.097
Sept.	0.072	0.072	0.097
Oct.	0.082	0.075	0.097
Nov.	0.098	0.081	0.097
Dec.	0.104	0.082	0.097
Annual	0.071	0.075	0.097

All values in (mm d^{-1})(mb m s^{-1})$^{-1}$. Coefficients: IFYGL M adjusted for wind and stability, IFYGL M_8 adjusted for wind only, and Lake Hefner (0.097) calibrated constant.

The inclusion of stability effects increased the Lake Superior mass transfer coefficients during winter and reduced them during summer. *Quinn and den Hartog* [1979] state that for many Great Lakes uses the available data do not justify the inclusion of the variation of the coefficient with stability and recommend simplified procedures to obtain the coefficient. This approximation, based on linear regression of the bulk transfer coefficient with wind, includes the variation of the mass transfer coefficient with wind speed for a constant value of bulk transfer coefficient. For the 8-m reference level the simplified coefficient is given by the equation

$$M_8 = 0.047 + 0.0046\ U_8 \tag{5}$$

where M_8 is the approximate mass transfer coefficient based on variation with wind speed, mm d^{-1}(mb m s^{-1})$^{-1}$ and U_8 is the wind speed at 8 m, m s^{-1}.

Tests of the above equation on Lake Superior produced a similar annual value (0.075) for the overwater coefficient (M_8), but drastically reduced seasonal variation in the coefficient (0.068–0.082). As shown later in the evaporation discussion, a large reduction of the high winter coefficients resulted in a 25% reduction of the annual evaporation values and produced overall results inferior to those obtained with the Lake Hefner constant. *Quinn and den Hartog* [1979] also present mass

TABLE 2. Average Values for Lake Superior Data Adjustments (Overwater), 1942–1975

Month	Wind Speed Ratio, R_w	Air Temperature Difference, °C, ΔT_{aw}	Dew Point Temperature, °C, ΔT_{dw}	Water Surface Temperature, °C, ΔT_w
Jan.	1.71	−4.7	−3.5	−0.1*
Feb.	1.74	−4.0	−2.6	0.2*
March	1.54	−1.8	−1.3	0.6*
April	1.09	1.1	−0.8	1.3*
May	1.02	3.9	0.9	2.8*
June	1.14	5.5	2.2	4.1
July	1.28	5.4	2.2	5.0
Aug.	1.27	2.3	0.0	2.3
Sept.	1.37	−0.1	−1.0	1.1
Oct.	1.40	−0.8	−1.1	0.9
Nov.	1.62	−2.3	−1.5	0.5
Dec.	1.75	−4.0	−2.6	0.0*
Annual	1.41	0.0	−0.8	1.6

* Temperature adjustments based on estimated water surface temperatures.

transfer coefficients determined from regression of the mass transfer product versus several other evaporation estimates. Their best coefficient from regression, based on aerodynamic evaporation estimates, was tested on Lake St. Clair [*Derecki,* 1979] and indicated results similar to those described above. The above tests indicate that simplified Lake Ontario coefficients should not be used for the other Great Lakes.

BASIC DATA AND DATA ADJUSTMENTS

Meteorological Data

Basic meteorological data and derived mass transfer parameters were obtained by averaging the records from three first-order meteorological stations located around the lake (Sault Ste. Marie, Duluth, and Thunder Bay). Records for wind speed, air temperature, and relative humidity were obtained from regular climatological publications in the United States and Canada (National Weather Service, NOAA, and Atmospheric Environment Service, Environment Canada). Individual station records were standardized at 8 m to be compatible with the equation (8-m coefficient) and to eliminate the periodic bias induced by differing measurement heights of various sensors, which ranged from 1 to 26 m. Adjustment of data to the standard height of 8 m was made with the following equation:

$$X_8 = X_m \frac{\ln (Z_8/Z_0)}{\ln (Z_m/Z_0)} \qquad (6)$$

where

- X_8 parameter value at 8 m;
- X_m parameter value at measured height;
- Z_8 reference height of 8 m;
- Z_0 roughness length, 0.0001 m;
- Z_m measurement height, m.

The perimeter wind speed for Lake Superior shows a high degree of consistency in monthly and annual values (3.8–5.0 m s^{-1}). Adjustment of the wind speed to the 8-m level reduced the average recorded values by 3% but varied throughout the period from a 9% reduction to a 2% increase. The magnitudes of the land-to-lake data adjustments for the open water conditions are given in Table 2, showing the average monthly and annual overwater values for the wind ratios and temperature differences. The overwater adjustment increased annual perimeter winds by 41%, varying seasonally from under 10% during spring to nearly 75% during winter. Actual winter adjustment was reduced to about 60% because of ice cover effects.

Average perimeter air temperature and relative humidity were used to determine dew point temperatures and ambient air vapor pressure. Perimeter air temperatures are below freezing for 5 months of the year (November–March) and vary seasonally from a low in January (−12.6°C) to a high in July (17.8°C). The average land-to-lake air temperature adjustments show that seasonal land-water air temperature differences are quite large (−4.7° to 5.5°C) but balance each other during the year. Owing to a lack of data, the overice air temperatures were assumed to be equal to the perimeter values, with a maximum of 0°C. This assumption should be valid during extensive ice cover. During limited ice cover the assumption is immaterial, since actual lake evaporation would not be changed significantly. Ice cover on the lake reduces resulting overlake air temperatures by nearly 2°C during February and approximately 0.5°C during other winter months.

Monthly humidity values for the lake perimeter are strongly consistent throughout the year, varying from about 70% during spring to about 80% during fall. The average perimeter dew point temperatures are about 4°C lower than air temperatures. The overwater adjustments of dew point temperature increased the average perimeter values by nearly 1°C, varying seasonally from a winter increase of about 3°C to a summer decrease of about 2°C. Winter overice adjustments averaged approximately −1°C but significantly reduced the resulting overlake dew point temperatures only during February.

Water Surface Temperature

The water temperature data for Lake Superior were obtained by adjusting average water temperature records from

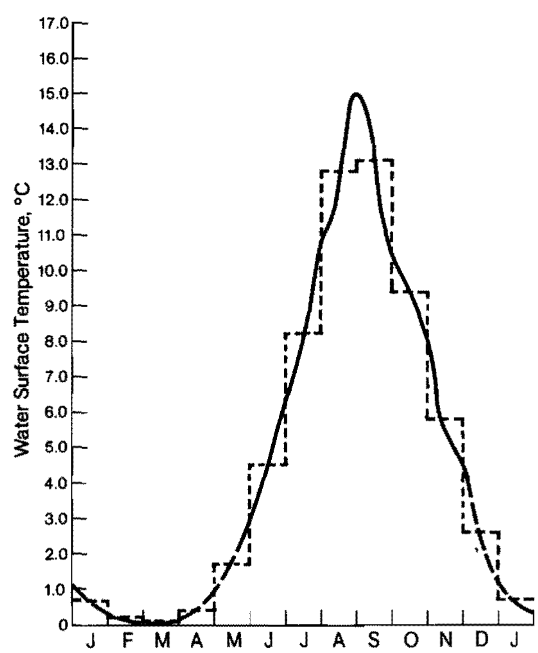

Fig. 2. Lake Superior mean seasonal water surface temperature distribution based on ART surveys, 1966–1975.

the municipal water intakes located at first-order stations (Sault Ste. Marie and Thunder Bay). Records for the Duluth water intake were omitted because they are obtained in deep water (20 m) and are insulated from the surface waters by the thermocline during most of the year. Municipal water intakes are the only source of continuous long-term water temperatures on the Great Lakes, but these records are for subsurface coastal temperatures and require adjustments to lake surface conditions. Surface temperature adjustments were derived from the airborne radiation thermometer (ART) survey measurements conducted since 1966 on the Great Lakes bordering Canada by the Atmospheric Environment Service, Environment Canada. Water surface temperatures obtained from satellites and ships of opportunity observations were also tested but exhibited poor accuracy and were discarded. Ship observations are obtained during normal passage and tend to avoid bad weather, thus biasing the data toward more frequently traveled routes and fair weather. Both ART and satellite observations are corrected for atmospheric attenuation, but available satellite data are not tied to surface observations and indicate a claimed ±2°C possible bias. Claimed accuracy for the ART temperatures is within 1°C [*Richards et al.*, 1969; *Irbe*, 1972].

Seasonal distribution of the ART water surface temperatures for the 1966–1975 period is shown in Figure 2. The ART data during individual years were normally insufficient to permit firm delineation of seasonal distribution. The ART surveys were limited to the open water season and the winter temperature distribution was estimated based on ice cover, air temperatures, and other water temperatures discussed above. The average monthly surface temperatures were obtained from the superimposed bar graph shown in the figure. Monthly water surface temperature adjustments were derived from simultaneous ART and water intake data. The adjustments indicate temperature differences similar to air and dew point temperature corrections and are listed in the last column of Table 2.

Temperature adjustments indicate that the water intake temperatures are considerably warmer than the lake water surface temperatures during summer but only slightly warmer during winter. The average monthly temperature differences vary from −0.1°C in January to 5.0°C in July. These average monthly water temperature corrections were applied throughout the study period to adjust the water intake records to the open water lake surface conditions. Because of lower winter temperatures, approaching 0°C, the use of average adjustments produced occasional negative water temperatures; therefore the minimum was preset at 0°C. The average monthly water surface temperatures vary from 0°C in March to 12.4°C in August. Additional surface temperature corrections were applied during winter for the ice-covered portion of the lake. The ice cover reduction of water temperatures is significant during the January–March period, with average monthly reductions of 1.3–4.7°C. These large reductions of water surface temperatures produce negative lake surface temperatures during winter.

The saturation vapor pressure was derived separately from the water and ice surface temperatures for the vapor pressure saturation with respect to water and ice surfaces, respectively, and combined with the corresponding ambient air vapor pressure in the evaporation computations. Resulting vapor pressure gradients were adjusted to the 8-m reference level by (6). Adjustment of the vapor pressure gradients to the standard height of 8 m increased the average vapor pressure difference values by 16%.

Ice Cover

The ice cover on Lake Superior was obtained from ice surveys conducted regularly since 1961 by the Great Lakes Environmental Research Laboratory (GLERL), NOAA, and the Ice Forecast Central in Canada. Estimates of the monthly average ice cover on the lake were determined from the individual surveys for the period of record (1961–1975) and computed by derived ice cover–air temperature relationships for the preceding years. Monthly ice cover equations were derived by multiple regression of available monthly ice cover data and perimeter air temperatures for the month and the preceding month. The equations are listed in Table 3. Statistical analysis of the equations shows strong correlation between the monthly ice cover and the 2-month air temperatures for February and March, the months of extensive ice cover. Weaker but significant correlation was obtained for January and April, the months of normally light ice cover. Computed ice cover was maintained arbitrarily, when needed, within 0–100% limits.

The observed and computed monthly ice cover estimates for the 1961–1975 period and the average monthly values for both the 1961–1975 and the 1942–1975 periods are given in Table 4. Agreement between observed and computed values is generally good, with maximum monthly differences of 10%. In the extensive ice cover months of February and March the ice covered approximately 50% of the lake area during the shorter period and 40% during the longer period, but varied from 15% to 80% during individual years. In the light ice cover months of January and April the ice cover varied from 0% to 30%, with the average value approximately 10%. Consideration of ice cover effects on computed mass transfer evaporation is particularly important during February and March because nearly half of the lake is normally ice covered. Since high evaporation from Lake Superior occurs during winter, the ice cover drastically reduces these high evaporation rates and produces a corresponding reduction in the total annual water loss from the lake.

TABLE 3. Lake Superior Monthly Ice Cover Equations

Month	Ice Cover, %			Multiple Correlation Coefficient	Standard Error, %	Mean, %
Jan.	IC = −15.30	$-1.793T_1$	$-0.313T_{12}$	0.86	2.8	11.7
Feb.	IC = −65.02	$-5.529T_2$	$-3.594T_1$	0.98	4.4	50.0
March	IC = −65.06	$-1.177T_3$	$-8.904T_2$	0.94	9.1	48.5
April	IC = −1.26	$+0.286T_4$	$-2.635T_3$	0.72	6.5	12.9

Note: Use equations to compute ice cover during 1942–1960 winter seasons.

Terms: IC = ice cover, % ($0 \leq IC \leq 100$); T_1 = January T_{ab} °C; T_2 = February T_{ab} °C; T_3 = March T_{ab} °C; T_4 = April T_{ab} °C; T_{12} = December T_{ab} °C; T_{al} = perimeter air temperature (land), °C.

TABLE 4. Estimates of Lake Superior Average Monthly Ice Cover in Percentages, 1961–1975

	January		February		March		April	
Year	Observed	Computed	Observed	Computed	Observed	Computed	Observed	Computed
1961	12	12	30	26	20	6	7	6
1962	20	15	70	68	60	66	17	9
1963	20	17	80	76	80	72	20	13
1964	3	3	15	16	18	26	10	18
1965	12	14	55	63	68	68	30	20
1966	12	15	49	49	16	28	0	7
1967	10	8	66	65	76	83	12	13
1968	15	10	60	52	60	50	6	5
1969	7	9	31	32	36	29	10	16
1970	13	14	60	65	74	64	17	17
1971	15	16	50	55	35	43	9	16
1972	13	15	71	68	77	69	27	20
1973	5	6	29	30	26	31	0	0
1974	12	12	54	56	55	59	18	17
1975	6	7	30	30	28	32	10	17
Mean	12	12	50	50	48	48	13	13
1942–1975		10		42		40		13

EVAPORATION

The monthly Lake Superior evaporation computed for the period of study (1942–1975) by the improved mass transfer method is listed in Table 5. Computed evaporation values are based on perimeter data and derived mass transfer coefficients, which were adjusted to lake surface conditions (water and ice surfaces) by atmospheric stability considerations and should indicate actual water loss from the lake. Annual mass transfer evaporation varied from a low of 405 mm to a high of 627 mm, with an average value of 483 mm.

The effects of data adjustments and standarization at the 8-m reference level are indicated in Table 6. The first two columns (E_m and E_8) indicate hypothetical perimeter evaporation, determined with measured and 8-m standardized data, intended primarily to show the effects of data standardization. Adjustment of the wind speed and vapor pressure gradient to the standard height of 8 m produced a net increase in evapo-

TABLE 5. Lake Superior Evaporation by the Mass Transfer Method

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
1942	110.7	65.9	26.3	−1.5	−6.0	−12.9	−11.2	−8.8	16.9	58.7	129.6	130.2	497.9
1943	101.0	63.0	76.3	14.0	−5.8	−11.1	−15.3	−14.1	16.9	49.0	118.2	135.5	527.6
1944	69.7	80.9	49.1	12.3	−7.6	−13.8	−24.1	−10.4	15.7	50.7	69.0	114.7	406.3
1945	88.7	49.8	9.5	5.9	.8	−6.8	−17.9	−12.2	18.9	70.9	108.5	95.0	411.1
1946	86.1	50.5	9.7	4.0	−.3	−7.1	−13.9	−6.3	15.1	38.0	124.4	125.1	425.4
1947	114.6	76.0	36.0	9.0	−5.0	−13.8	−18.0	−20.5	10.7	14.7	99.2	102.0	404.9
1948	91.3	47.5	31.5	−3.3	−.2	−6.5	−11.2	−11.7	16.9	55.4	78.4	116.4	404.6
1949	111.7	61.8	47.2	1.3	−3.2	−13.4	−14.6	−8.5	24.2	41.9	140.2	137.5	526.0
1950	127.6	52.8	62.7	26.4	−4.7	−8.1	−11.5	−.8	−.2	27.7	116.1	112.4	500.2
1951	102.3	56.8	53.9	−.4	−4.3	−11.4	−17.8	−9.3	10.8	33.4	104.4	117.2	435.6
1952	93.2	61.6	50.5	.0	−3.7	−10.2	−12.9	−11.8	−4.4	85.0	89.9	84.1	421.5
1953	116.8	62.1	34.3	12.1	−4.3	−11.9	−16.8	−9.2	49.3	49.7	86.6	130.9	499.6
1954	102.0	41.3	83.9	7.8	−2.5	−9.0	−12.3	−3.4	9.2	53.0	58.8	89.6	418.6
1955	113.5	57.1	62.7	−4.4	−3.0	−13.6	−19.3	−8.7	38.0	44.4	142.8	126.6	536.0
1956	92.4	67.8	55.2	18.6	−1.5	−7.8	−16.3	−10.7	23.9	29.0	114.3	116.0	481.0
1957	104.8	43.7	36.2	2.2	1.6	−10.0	−18.0	.5	42.8	69.3	103.6	102.2	479.0
1958	91.7	73.9	20.7	7.6	−.2	−5.7	−19.4	−6.6	4.2	38.6	119.9	120.4	445.3
1959	104.7	39.1	28.8	8.5	−9.8	−7.6	−9.1	−14.7	10.6	103.9	136.5	84.2	475.2
1960	94.7	65.2	60.8	2.8	−5.2	−6.0	−13.8	−11.6	33.4	56.3	102.9	128.5	508.0
1961	93.5	50.2	38.3	8.9	−.7	−4.3	−14.7	.1	40.0	57.4	89.6	108.7	467.1
1962	117.3	33.0	24.2	9.6	−6.2	−8.1	−9.4	−4.5	49.7	57.1	74.7	124.8	462.4
1963	94.5	27.1	25.7	3.2	−2.4	−14.3	−12.0	−1.0	10.5	27.4	109.3	137.7	405.7
1964	104.3	74.2	66.4	4.7	−3.6	−6.4	−10.9	−2.2	25.1	66.6	101.1	143.6	562.8
1965	109.7	46.6	32.8	5.1	−5.3	−5.5	−14.5	−3.3	41.1	50.7	105.1	91.2	453.5
1966	104.6	48.1	43.6	9.8	2.7	−7.2	−12.1	−4.8	47.4	84.5	110.5	107.8	534.8
1967	114.3	39.1	23.2	11.4	5.6	−12.3	−11.3	3.6	38.5	76.2	101.4	113.5	503.4
1968	94.0	66.9	29.7	4.3	−2.3	−8.8	−10.4	−8.9	15.0	54.8	113.8	124.3	472.4
1969	109.7	61.3	54.6	2.6	−1.4	−5.2	−9.3	−1.5	77.6	102.7	106.3	129.3	626.7
1970	135.8	42.0	35.9	7.0	−3.4	−5.2	−13.0	.2	51.7	61.2	107.9	112.8	533.0
1971	112.4	47.8	44.6	11.6	1.5	−7.0	−5.2	7.4	28.5	41.8	117.6	135.1	536.1
1972	119.2	46.0	35.1	17.6	−2.2	−7.4	−9.4	−6.1	49.6	63.0	76.4	140.8	522.6
1973	85.2	71.1	18.8	16.3	−.4	−15.3	−12.4	−7.3	54.3	36.3	105.6	138.5	490.6
1974	90.3	53.8	42.6	6.2	−1.5	−10.7	−14.9	−7.9	57.9	62.9	74.4	94.5	447.6
1975	118.3	62.8	70.3	26.5	−4.9	−19.2	−21.4	2.5	57.2	51.8	105.0	142.6	591.5
Mean	103.5	55.5	41.8	7.9	−2.6	−9.5	−13.9	−6.2	29.3	54.8	104.2	118.1	482.8

All values in mm.

TABLE 6. Average Mass Transfer Evaporation for Lake Superior, 1942–1975

Month	Perimeter E_m	Perimeter E_8	Overwater, E_w	Overice, E_i	Overlake, E
Jan.	45.4	50.4	115.2	5.0	103.5
Feb.	37.4	41.6	94.2	5.7	55.5
March	32.0	35.5	62.9	12.0	41.8
April	14.0	15.6	7.7	7.7*	7.7*
May	5.1	5.6	−2.6		−2.6
June	−7.4	−8.2	−9.5		−9.5
July	−2.9	−3.2	−13.9		−13.9
Aug.	8.8	9.8	−6.2		−6.2
Sept.	33.9	37.7	29.3		29.3
Oct.	43.4	48.2	54.8		54.8
Nov.	50.3	55.9	104.2		104.2
Dec.	46.3	51.4	118.1		118.1
Annual	306.3	340.3	554.2		482.8

All values in mm. Evaporation determined by the proposed mass transfer method using data indicated as follows: E_m is from the measured perimeter data (land), E_8 is from the perimeter data standardized at 8 m, E_w is the overwater data (8 m), E_i is from the overice data (8 m), and E is from the areally weighted overwater and overice evaporation.

* Estimates based on overwater results. Erroneous data (ice temperature estimates) produced irrational results of higher overice evaporation (8.5 mm) and corresponding overlake value (7.9 mm) than the overwater evaporation.

ration of 11%. Because of differences in atmospheric stability over large lake and land surfaces, perimeter data without adjustments are not suitable for evaporation computations. The hypothetical perimeter evaporation (E_8) for Lake Superior indicates a large reduction from lake evaporation (E) during the high evaporation season (about 50%) and produces a net annual reduction of 30%. Because of extensive ice cover on Lake Superior during winter, the overwater evaporation (E_w) indicates a substantial increase over the actual lake evaporation (E). During the January–March high evaporation period the average overwater evaporation (63–115 mm) exceeds the low overice evaporation (E_i) values (5–12 mm) by amounts ranging from 50 to 110 mm per month. Elimination of the ice cover effects on Lake Superior during these months, inherent in the standard overwater mass transfer computations, increases the average monthly evaporation values by 10–40 mm and the annual total by 70 mm, which represents 15% of the actual lake evaporation. The ice cover effect in April may be significant during individual years, but has little effect on the average evaporation values. An apparent anomaly of higher overice evaporation (8.5 mm) than overwater evaporation (7.7 mm) was obtained in April because of erroneous data, primarily in the ice temperature estimates, which are particularly difficult for April. Data adjustments required several assumptions discussed previously. However, this increase is small and the ice cover in April is not extensive, producing an insignificant increase in the overlake evaporation (7.9 mm). This inconsistency is eliminated by disregarding computed overice values and using the overwater values for all three conditions. Generally, the ice cover reduction of evaporation could be estimated by reducing the lake area by appropriate ice cover.

Seasonal distribution of the mass transfer evaporation for the average, maximum, and minimum monthly values is shown in Figure 3. During the high evaporation season of fall and winter, the average monthly losses from the lake normally exceed 100 mm in the November–January period. The highest monthly mass transfer evaporation, which occurs in December, yields an average value of 118 mm and a maximum of 143 mm. During the low evaporation season of spring and summer, the evaporation process is normally reversed to condensation in the May–August period. The highest monthly condensation, which occurs in July, normally exceeds 10 mm, with an average value of 14 mm and a maximum of 24 mm. Condensation also frequently exceeds 10 mm in June.

Presented Lake Superior evaporation estimates are derived from monthly values of the input data (air and water temperatures, relative humidity, and wind speed). Since evaporation is a short-period process, these estimates may contain averaging errors due to temporal nonlinearity of the meteorological data. This aspect of monthly evaporation estimates was not investigated for Lake Superior, but both temporal and spatial averaging effects on computed evaporation were investigated on Lake Ontario, using the extensive IFYGL data base. *Quinn and den Hartog* [1979] found that, while evaporation is primarily an hourly and daily phenomenon, mass transfer evaporation estimates based on weekly and monthly average values of meteorological data produced acceptable results. *Quinn* [1979] also states that there is no degradation in computed evaporation when daily averaged rather than hourly averaged meteorological parameters are used.

SENSITIVITY AND ERROR ANALYSIS

The relative sensitivity and error variance of the input parameters on computed evaporation was determined by a modified version [*Quinn*, 1979] of the sensitivity and error variance functions presented by *Coleman and DeCoursey* [1976]. Quinn's modification of the sensitivity function involves the definition of the range for the independent variables. He employs the total range (maximum–minimum) instead of the partial range (measured–minimum) used by Coleman and DeCoursey. The relative importance of the independent parameters as defined by the relative sensitivity function is

$$\Psi_{R_i} = \frac{\Delta E}{\Delta X_i} \frac{(X_{\max} - X_{\min})}{E} \tag{7}$$

where

Ψ_{R_i} relative sensitivity;
ΔE evaporation increment;

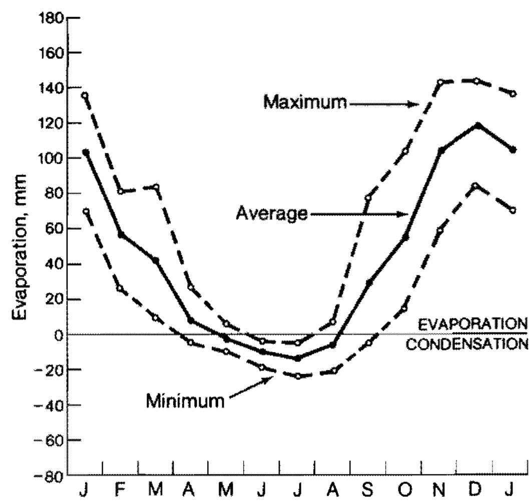

Fig. 3. Lake Superior monthly evaporation by mass transfer method, 1942–1975.

ΔX_i unit change in the independent parameter;
X_{max} maximum value of the independent parameter;
X_{min} minimum value of the independent parameter;
E evaporation.

The error variance function, designed to indicate possible error contributions from each of the independent parameters, is defined as

$$E[V(X)] = \sum_{i=1}^{n} \left(\frac{\Delta E}{\Delta X_i}\right)^2 \mathrm{Var}\,(X_i) \tag{8}$$

where

$E[V(X)]$ expected error due to variance of X;
$\mathrm{Var}\,(X_i)$ variance of the independent variable X;
Σ summation, $i = 1, \cdots, n$ variables.

The results of the relative sensitivity and error variance analysis for the annual values are given in Table 7. The relative sensitivity and error variance analyses were also tested for the high and low evaporation, both seasonal and monthly, with generally similar results. Computed evaporation is most sensitive to the dew point temperature and highly sensitive to the water surface temperature, while other parameters (wind speed, air temperature, and bulk evaporation coefficient) are relatively unimportant.

The variance is the standard error of measurement squared, which is expressed in appropriate units, except for the bulk evaporation coefficient, which is expressed as a percentage of the mean parameter value ($\bar{X}$). Indicated standard errors for the meteorological data represent generally accepted limits of accuracy for the Great Lakes (about 10%). The greatest potential error indicated by the error variance is due to the wind speed, followed by the greatly reduced influence of the water surface temperature, dew point temperature, and bulk evaporation coefficient. Air temperature is again unimportant. Similar results were obtained for Lake Ontario by *Quinn* [1979].

VERIFICATION OF MASS TRANSFER EVAPORATION

Verification of evaporation determined by the mass transfer method is provided by the water budget determinations [*Derecki,* 1980]. Since water budget evaporation is determined as a residual of the water supply and losses from the lake, computed evaporation contains the errors of input hydrologic components. Care was exercised to reduce these errors to a minimum by careful treatment of the individual hydrologic components (overwater precipitation, runoff from drainage basin, outflow from the lake, and lake storage). In contrast to the other Great Lakes, where lake outflow normally exceeds other hydrologic components by an order of magnitude, all important water budget factors for Lake Superior are of the same order of magnitude, eliminating the possibility of large residual evaporation errors due to relatively small errors in one of the inputs. Factors disregarded in the Lake Superior water budget are the groundwater fluxes and thermal effects on the lake levels, from which lake storage was determined. Direct groundwater contributions to the Great Lakes water budgets are generally considered to be negligible, which is substantiated by the more recent studies investigated for this report [*Derecki,* 1980].

The expansion and contraction of water associated with seasonal warming and cooling of the lake affect lake levels and the magnitude of seasonal water storage, and consequently monthly water budget computations. The annual evaporation is not affected, since net annual thermal changes are insignificant for the water balance considerations. Two recent studies of thermal effects on Lake Superior, by *Meredith* [1974] and *Bennett* [1978], were investigated. Both studies indicate annual balancing of monthly expansions and contractions of water, but monthly values generally compare poorly, with frequent large differences or contradicting results. Disagreement between the two sets of monthly thermal adjustments for lake levels are as large as the mass transfer and water budget evaporation differences presented in this paper. The thermal effects are relatively small, with maximum monthly values (expansion or contraction) of about 10 mm, and could be exceeded by the measurement errors for the change of lake storage determinations. Resolution of the thermal effects on lake levels with reliable long-term water temperature profile data would improve seasonal distribution of the water budget evaporation, but such data are simply not available at present.

Both the mass transfer and water budget computations indicate relatively constant long-term evaporation, which fluctuates around the 500 mm yr^{-1} value. In comparison with the water budget evaporation the mass transfer determinations agree reasonably well in the average seasonal distribution and the extremes (maximum and minimum) of the high evaporation season. During the low evaporation season, the mass transfer extremes indicate a reduced range of variation in monthly evaporation. The average 1942–1975 monthly evaporation from Lake Superior as determined by both the water budget and the mass transfer methods is shown in Figure 4. The figure also shows the ice cover reduction of the mass transfer evaporation during winter. As indicated in the figure, the ice cover adjustment produces much better agreement with the water budget evaporation values. The major disagreement between the two determinations is an apparent lag of about a month between the water budget and the mass transfer evaporation values during the increasing evaporation period, beginning in July. It might be noted that applying long-term thermal corrections, discussed in the preceding paragraph, to the water budget values would generally not improve the water budget–mass transfer evaporation comparison. However, *Bennett*'s [1978] corrections, which appear to be based on more sound water temperature profile data, would improve the comparison slightly during most months of the apparent lag period (July–October), with maximum re-

TABLE 7. Mass Transfer Sensitivity and Error Variance Analysis, 1942–1975

Parameter, X	Sensitivity, Ψ_{R_i}	Standard Error, SE	Error Variance, $E[V(X)]$
Wind speed	0.19	1.0 m/s	89.5
Water surface temperature	2.60	0.5°C	23.4
Dew point temperature	4.23	0.5°C	11.2
Air temperature	0.07	0.5°C	0.0
Bulk evaporation coefficient	0.73	10%($\bar{X}$)	9.5

duction of differences by about one quarter. Similar lag was also obtained for Lake Erie [*Derecki,* 1976] and was attributed to inaccuracies of data, particularly in the water surface temperatures. In the present study the water surface temperatures represent the weakest link in the mass transfer computations. Elimination of this weakness will be feasible when the satellite surface temperature observations become sufficiently accurate.

As a point of interest, an additional comparison of the mass transfer and water budget evaporation was made for 1973, with the water budget and energy budget estimates obtained by *Bennett* [1978] and *Schertzer* [1978], respectively, during a comprehensive Lake Superior limnology study. The agreement between the four sets of evaporation values is reasonably good and generally similar to that shown in Figure 4 for the present long-term average monthly values. Major disagreements occur during summer–fall and midwinter periods. The mass transfer estimates begin to show consistently lower evaporation during summer but eventually indicate higher water loss in late fall (as in Figure 4). During midwinter, evaporation as indicated by the energy budget estimates is higher by similar magnitude. All estimates indicate close agreement during the low evaporation period, but during the high evaporation period closest overall agreement is provided by the present determinations. Bennett's water budget estimates contain thermal water density corrections, but as mentioned previously, these corrections are relatively minor.

The comparison of average evaporation for Lake Superior in Table 8 shows the evaluation of various mass transfer coefficients. The coefficients evaluated are the IFYGL coefficient M, the approximate IFYGL coefficient M_8, and the Lake Hefner coefficient or calibrated constant, 0.097. The water budget and mass transfer with coefficient M determinations represent the best long-term evaporation estimates feasible at present from the available data. Evaporation estimates with the approximate mass transfer coefficient M_8 indicate a large reduction of evaporation during most months, reflecting the effects of neglecting air stability variations. The combined effect of reduced evaporation and increased condensation produced a 25% reduction in the annual evaporation. The overall effect, both monthly and annual, of the approximate coefficient produces evaporation results that are worse than those obtained with the Lake Hefner constant. Average evaporation estimates

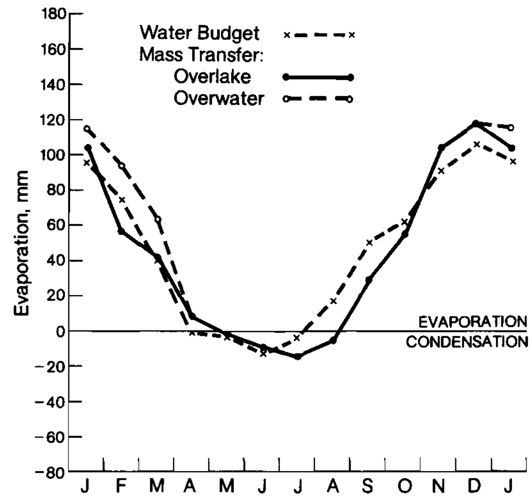

Fig. 4. Comparison of average monthly Lake Superior evaporation, 1942–1975.

obtained with the Lake Hefner coefficient appear reasonable during the high evaporation season, but are inferior to those of the approximate coefficient during the low evaporation season. The use of the relatively high Lake Hefner constant coefficient produces unrealistically high condensation values, which result in a 13% reduction of the annual evaporation. The large reduction of the evaporation estimates obtained with the approximate Lake Ontario coefficient M_8 shows that this simplified procedure is not suitable for Lake Superior and probably not for the remaining Great Lakes.

CONCLUSIONS

Evaporation from Lake Superior was determined by the improved mass transfer method, which includes atmospheric stability effects for both open water and ice-covered lake surfaces. The mass transfer computations employ individual monthly land-to-lake data adjustments and a variable mass transfer coefficient to provide realistic operational evaporation estimates, which are verified by the water budget determinations. The average annual difference between water budget and mass transfer evaporation is 7%, which is within normal limits of accuracy for the Great Lakes climatological data (about 10%).

TABLE 8. Comparison of Average Evaporation for Lake Superior, 1942–1975

		Mass Transfer Method		
Month	Water Budget Method [*Derecki,* 1980]	IFYGL Coefficient M	IFYGL Approximate M_8	Lake Hefner Coefficient
Jan.	96.5	103.5	81.0	96.1
Feb.	75.4	55.5	44.7	54.5
March	40.6	41.8	36.0	43.9
April	−1.3	7.9	8.6	11.1
May	−3.3	−2.6	−5.5	−8.0
June	−12.7	−9.5	−21.6	−30.6
July	−4.2	−13.9	−30.0	−42.4
Aug.	17.2	−6.2	−10.2	−14.6
Sept.	50.4	29.3	28.1	36.5
Oct.	62.3	54.8	49.0	62.7
Nov.	90.7	104.2	86.3	102.6
Dec.	105.8	118.1	93.8	110.2
Annual	517.4	482.8	360.3	422.0

All values in mm. Coefficients: IFYGL M adjusted for wind and stability, IFYGL M_8 adjusted for wind only, and Lake Hefner (0.097) calibrated constant.

Normal long-term evaporation removes approximately 500 mm of water from the lake surface annually but varies substantially from year to year. Monthly Lake Superior evaporation indicates a high evaporation season during fall and winter and a low evaporation season during spring and summer. During the peak of the high evaporation season, monthly water losses frequently exceed 100 mm. During the low evaporation season the evaporation process is frequently reversed to condensation (negative evaporation). Monthly condensation values during the peak condensation season are normally under 15 mm. Employment of a variable mass transfer coefficient, based on air stability, eliminates unrealistically high normal monthly condensation values during the peak condensation season. Approximate mass transfer coefficients tested in the study (simplified IFYGL and the Lake Hefner constant) produced inferior results and should not be used for the Great Lakes. Evaporation estimates with these coefficients produced average monthly condensation values from two to three times higher, while a previous Lake Superior mass transfer study [*Richards and Irbe*, 1969] indicated average monthly condensation values six times higher. The Lake Superior ice cover during winter, which normally covers from 10% to 40% of the lake area, reduces the potential overwater evaporation by a similar percentage. The ice cover reduction of evaporation, which was generally ignored by other investigators in previous Great Lakes mass transfer studies, could be estimated by simply reducing the lake area by the appropriate percentage of ice cover.

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