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Environment Canada
Water Science and Technology Directorate

Direction générale des sciences et de la technologie, eau
Environnement Canada

Physical Climatology of Lakes

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NWRI Contribution # 93-39

13-39

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Chapter 6 : **Surface Climates of Canada**

McGill / Queens University Press

Management Perspective

Canada's vast landscape encompasses many different surface types and climatic regions. Climate and climate change issues are a natural concern for Canadians where socio-economic activities such as agriculture, transportation, navigation, hydroelectric potential, fisheries etc. are so interlinked with climate. This manuscript forms one of 15 Chapters written by invited Canadian experts in physical climatology which addresses the "Surface Climates of Canada". Baseline conditions as well as the potential for climate change impacts on undisturbed (natural) surfaces and well as disturbed surfaces are described. The physical climatology of natural systems included such landform types as snow and ice, ocean and coastal zones, lakes, wetlands, Arctic Island, low Arctic and subArctic, forest and alpine surfaces. Disturbed surfaces discussions detailed physical climatology of agricultural and urban landforms.

The Chapter on physical climatology of lakes recognizes, in large part, the significant research conducted by the National Water Research Instutute as well as others in understanding the complex seasonal distributions of physical climatological and hydrologuical components of the Great Lakes in particular.

CHAPTER 6 PHYSICAL CLIMATOLOGY OF LAKES

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1. INTRODUCTION

While oceans are vital to climate on a global scale, lakes can significantly affect local and regional climate. Lakes can transport and store heat, respond to atmospheric forcing and feed-back to affect the overlying air. Air/water exchanges of heat and moisture has climatological implications not only for the water body but also on the climate of the overlying air and areas adjacent to the water mass. Understanding the interactions and implications of lakes on climate necessitates examination of the physical characteristics of water bodies and temporal and spatial variation of the momentum, heat and mass exchanges.

Much of the physical climatology research on lakes has been focused on understanding the variation of heat and moisture exchanges at the air/water interface with application to estimate lake evaporation. In Canada there are as many as 2 million freshwater lakes within it's five major drainage basins representing approximately 7.6% of the country's total area (Figure 1ab). Nearly 14 percent of the world's lakes having surface areas over 500km^2 are located in Canada. The Laurentian Great Lakes constitute the largest interconnected freshwater body on Earth consisting of Lake Superior, Lake Huron, Lake Erie and Lake Ontario bordering Canada and the USA and Lake Michigan entirely within USA. These lakes are the dominant geographical feature of the western part of the Great Lake-St. Lawrence River climatic region (see Chapter 2).

Although Canada's water resource base is large, much of the physical climatology of its 2 million lakes is largely unexplored. Recently, concern over degradation of lake water quality from pollution, acid rain, toxic rain and issues such as climate change has increased research on water resources including lakes in the vast northland (Prowse and Ommanney, 1990) and in specialized basin studies. Much physical, biological and chemical research has been conducted on the Laurentian Great Lakes which supports large socio-economic-industrial concerns in this region. Several large-scale experiments have included the physical climatology of lakes. These include the International Field Year for the Great Lakes (IFYGL, 1981) on Lake Ontario, the Upper Great Lakes Reference Study (IJC, 1977) which examined Lake Superior and Lake Huron, the Lake Erie Binational Study (Boyce et.al. 1987) and the Upper Great Lakes Connecting Channel Study (Shimizu and Finch 1988).

While it is not possible to adequately cover the physical climatology of all of the various sizes and types of lakes across Canada, this Chapter provides an overview of the primary elements of this environmental field. Emphasis is placed on description of the momentum, radiation, energy and water budget and evaporation of large lake systems using the Great Lakes as the primary focus.

2. PHYSICAL CHARACTERISTICS OF LAKES

The Canadian climate is diverse and in large measure, Canada's lakes reflect the wide diversity of the climatic regions.

2.1 Formation and Type : Canada has **tectonic** (tectonic impacts and subsidence), **barrier** (volcanic) and **excavated** (glacial action) lakes. Morphometric characteristics of area, depth and volume greatly influence the lake physical responses and air/water exchanges. The diversity in the Great Lakes is shown in the depth profiles (Figure 1c). Water volume (V) is greatest in Lake Superior (11600 km^3) compared to 3580, 1710 and 545km^3 for Lake Huron, Lake Ontario and Lake Erie respectively. The flushing time (F) of water $F=V/(Q-E)$ where Q is discharge and E is evaporation, is longer for **closed** lakes in which water leaves primarily through evaporation compared to **open** lakes in which loss also occur through drainage and seepage. Flushing times for the Great Lakes (Schertzer 1980) are Lake Superior 165.0 yrs., Lake Michigan 69.5 yrs., Main Lake Huron 10.6 yrs., Georgian Bay 5.7 yrs., Lake Erie 2.5 yrs., and Lake Ontario 7.5 yrs.

Climate conditions are often reflected in the general thermal and circulation characteristics of lakes. Polar and mountain lakes with permanent ice cover are termed **amictic**. Lakes which only circulate fully during the summer are **monomictic** while temperate lakes with complete turnover during both the spring and autumn, such as the Great Lakes, are **dimictic**.

2.2 Physical Properties : Density of water accounts for many of the physical changes (temperature stratification) occurring in large lakes. Maximum density of freshwater is about 4°C and decreases both at warmer temperatures and as it cools to its freezing point at 0°C. The temperature of maximum density is affected by pressure, decreasing by 0.2°C for every 100-metre depth of water.

Water is significantly different from other surfaces. It has one of the highest specific heat capacities of all substances (4.186 $kJkg^{-1}C^{-1}$), about twice that of ice (2.04 $kJkg^{-1}C^{-1}$). This produces a small diurnal temperature range. Compared to other surface types, water has a low thermal conductivity of 0.57 $W\ m^{-1}\ K^{-1}$. Consequently, heat transport in water occurs almost entirely by water movement. Solar radiation penetrates through the surface and as a consequence, energy absorption is distributed through a large volume. Heat distribution throughout the water volume occurs by wind mixing, convection and advection. Open water surfaces have unlimited water availability for evaporation. Water is not compressible like air, however it can be deformed resulting in surface waves. Dynamical processes such as internal waves, upwelling/downwelling and Seiches redistribute heat within the water column.

2.3 Seasonal Thermal Cycles : Lake temperature and wind mixing significantly affects exchanges at the air/water interface. Summer water surface temperatures (Figure 2) are higher for shallow Lake Erie compared to the other Great Lakes, while the thermal inertia of Lake Huron and Lake Ontario allows higher surface temperatures in winter. Characteristics of the seasonal thermal cycles vary among lakes of different physical dimensions and latitudinal location.

Large dimictic lakes mix (overturn) twice yearly, in the spring and the fall, when the surface water temperature reaches maximum density and increased wind speed and rate of surface heat flux help to stir the lake (Schertzer, 1980). Winter temperatures are generally isothermal below the temperature of maximum density. In spring, shallow nearshore water warms more rapidly forming a convergence zone (thermal bar); its disappearance marking the onset of summer stratification, which in deep lakes such as the Great Lakes takes as much as six to eight weeks (Rodgers, 1987; Schertzer 1987). Thermal stratification proceeds at varying rates forming a warm upper layer (epilimnion), and a cold lower layer (hypolimnion) separated by the thermocline which inhibits heat and particle transfer (Schertzer et.al. 1987). Maximum heat storage occurs towards the end of summer. Storms with high wind stresses are responsible for intense energy transfer from the lake to the air and vertical mixing which breaks down the thermocline culminating in fall overturn. Wind mixing and heat losses continue to cool the lake. By the end of winter, the entire water mass cools down to below 4°C with the coldest water remaining close to shore.

Satellite imagery shows non-uniform characteristics of surface temperature. Nearshore areas warm faster. Wind stresses may act to tilt the thermocline resulting in upwelling of colder mesolimnion or hypolimnion waters on one shore and downwelling of warmer water on the opposite shoreline (Simons and Schertzer, 1987). Dynamical circulation processes such as internal waves and Seiches (Boyce 1989) result in increased variability of lake temperatures. For a large lake such as Lake Ontario, variable wind (and other meteorological fields) complicates sampling and averaging of observations for developing a representative climatology over short time scales.

Reduced thermal inertia in small, shallow lakes results in faster spring heating and fall cooling and more uniform temperature distributions. Penetration of solar radiation to the lake bottom becomes an important factor in the lake heat budget as the water column is heated from above and from below. In very shallow lakes, bottom vegetation can also enhance warming. Net advection of heat from tributaries may not be minor and conduction between water and the land can become important. Shallow lakes often have complete ice cover. Due to their smaller size, small lakes generally do not modify surrounding land to the extent felt by large lakes the size of the Great Lakes.

2.4 Measurement Systems : Figure 3a depicts a standard meteorological buoy used on the Great Lakes which typically includes instrumentation to observe air temperature, surface water temperature, relative humidity, wind speed and direction and can accommodate measurement of radiation fluxes. Other systems such as towers (Figure 3b) offer stable platforms from which specialized measurements can be conducted.

3. MOMENTUM TRANSFER

Surface wind stress (τ) is a critical component in lake studies since it drives lake circulation, lake set-up, surface waves and turbulent transfer at the air/water interface. Spatial and temporal variability in τ can be high for large lakes since detailed overlake wind observations are often lacking. Occurrences of overlake fog, lake-land breeze circulations and modifications of vertical profiles under warm air or cold air advection (Chapter 13) often complicates estimations.

Surface wind stresses in excess of 0.2 Nm^{-2} on Lake Erie resulted in significant vertical mixing of heat especially in spring and fall months under unstable thermal stratification (Schertzer, et.al. 1987). Surface wind stress influences upwelling and downwelling events in large lakes affecting horizontal and vertical temperature distributions (Lam and Schertzer 1987), currents and heat transports (Simons and Schertzer, 1987).

Accurate estimates of the surface wind stress over lake surfaces requires information on the surface roughness characteristics and the drag coefficient.

The aerodynamic characteristic of the surface is related to the height, shape and spacing of roughness elements and defines a roughness length (Z_o) representing the height above the surface at which the mean wind speed becomes zero. For water surfaces, Z_o is complicated by mobility of the roughness elements ranging from 10^{-6} to 10mm over smooth and rough water surfaces (Donelan 1990).

The drag coefficient C_D shows weak linear dependence on wind speed, and in non-steady inhomogeneous surface conditions both Z_o and C_D must also account for the wave field. The effect of nonstationarity of the wind field on the value of the drag coefficient is significant. Over surface conditions of ultrasmooth to fully rough, drag coefficients, C_D measured at 10m height above the water, may vary by an order of magnitude ($3x10^{-4}$ to $3x10^{-3}$ dependant on wind speed, fetch, stability and wave conditions (Donelan 1990).

4. RADIATION EXCHANGES

Important elements of physical climatology in lake studies include the assessment of evaporation and surface heat exchange which is a source term for many important processes such as thermal stratification and biological productivity. The primary components of the lake water and energy budget is summarized in Figure 4. Determining the surface heat flux requires estimation of the radiation and turbulent heat flux components as well as the lake heat storage. The radiation balance provides a statement on the net amount of energy available to the lake surface resulting from atmospheric and surface radiative exchanges.

The net radiation flux, Q^*, is the algebraic sum of the incoming and outgoing radiative fluxes at the surface of the lake expressed as,

$$Q^* = (1 - \alpha)K\downarrow + (1 - \epsilon)L\downarrow - \epsilon\sigma T_s^4$$

where $K\downarrow$ is incoming solar radiation, $L\downarrow$ is incoming longwave radiation, α is surface albedo, ϵ is infrared emissivity, σ is the Stefan-Boltzmann constant ($5.57 \times 10^{-8} Wm^{-2}K^{-4}$) and T_s is surface temperature. Lake surface characteristics such as temperature play an important role in the radiation balance. Other important surface controls include the surface emissivity and albedo.

4.1 Surface Controls : If a surface at a given temperature emits the maximum possible amount of radiation per unit of its surface area in unit time its emissivity is unity. The emissivity of water and ice ranges from 0.92 to 0.97. On Lake Ontario, Robinson et.al (1972) determined ϵ = 0.972 with a standard deviation of 0.21, except for a thin film of oil which reduces emissivity by about 3 percent. The emissivity of snow ranges from 0.82 to 0.99 for old and new snow respectively (Oke, 1983).

The reflectivity of a surface to solar radiation defines its albedo. Estimates of backscattered light from subsurface layers range from 2-3% for turbid water to 6% for clear ocean water.

Albedo over a water surface increases with increasing zenith angle and is consistently larger under rough (waveheights > 30cm) conditions (Nunez et.al 1971, 1972). For large zenith angles and under cloudless skies, diffuse radiation may lower the albedo while waves tend to increase the albedo. Albedo under clear sky and scattered cloud (Figure 5a) increase similarly as zenith angle increases. Diurnal changes for broken cloud are considerably less and under overcast conditions the mean albedo averages about 7.5%. Daily values of albedo over the period July to November (Figure 5d) ranged from 7% in early July to 11% in the middle of November.

Albedo ranges from 77-95% and 40-70% for fresh and old snow respectively. Sea ice albedo ranges from 30-45% (Oke, 1983), while higher reflectivities occur at lower wavelengths for all ice types with highest reflection over densely consolidated brash and lowest for thin skim ice (Leshkevich, 1984). Decay of the ice surface results in significant diurnal variations in albedo, however, Bolsenga (1977) suggested that an average mid-day albedo under average sky conditions would suffice for computation of K↑ in most lake energy budget studies. Daily mean value of albedo for winter months have been estimated at 16% (Davies et.al. 1975).

4.2 Solar Radiation Penetration : Solar radiation also penetrates the water surface and is greatest at the shorter wavelengths 0.2 to 0.6μ while longer wavelengths are absorbed in the upper few centimetres. The energy is used for heating and also for biological productivity (Schertzer 1978). The diminution of light intensity between two irradiances at the surface I_o and at some depth I_z, follows Beer's law.

The rate of light attenuation (α_v) is defined as the mean vertical extinction coefficient. For pure water, (α_v) is estimated at 0.1 m^{-1}. Spatial and temporal variability in (α_v) over the Great Lakes ranges from 0.15 to 1.0 m^{-1} (Schertzer et.al. 1978). Recent investigations of lake optics (Bukata et.al 1988) have detailed spectral characteristics of light attenuation in the Great Lakes.

4.3 Diurnal/Seasonal Variations of Radiation Components : Figure 6a shows the diurnal variation of radiation budget components for Lake

Ontario for cloudless conditions on 28 August 1969 (Davies et.al. 1970). Under cloudy conditions solar radiation at the water surface may be significantly reduced. Diffuse-beam ranges from 25-75% of solar radiation and varies depending on solar altitude (Oke 1983). Compared to incoming solar radiation, the reflected solar radiation is relatively small. The net diurnal solar radiation flux ($K^*=K\downarrow-K\uparrow$) is positive.

Incoming longwave radiation is dependent on atmospheric absorbers and emitters such as water vapour, clouds, and carbon dioxide. Under cloudless skies $L\downarrow$ varies diurnally with short-term variations of up to 20 percent (Robinson et.al. 1972). $L\downarrow$ increases in the presence of medium to low cloud, however, high cloud has little influence. Emitted longwave radiation ($L\uparrow$) is dependant on the water surface temperature alone with constant emissivity. Surface water temperature varies diurnally and is affected by sea state, surface sensible and latent heat exchange, mixing and advection. Water intakes, airborne radiometric techniques (ART), satellite measurements, and ship surveillance are common sources of water temperature estimates. Comparison of ART and satellite temperature measurements with in situ observations show accuracies in the range of 1.5°C under controlled conditions (Robinson et.al. 1972). Comparisons of $L\uparrow$ determined using surface water temperatures from float thermistor ($L_{TS}\uparrow$) and from infrared radiometer "skin temperature" techniques ($L_R\uparrow$) showed $L_R\uparrow$ more variable diurnally than $L_{TS}\uparrow$ which was relatively constant through the day (Robinson et.al. 1972). Comparison of L↑ determined from the surface and at radiometer height indicated significant divergence in summer and convergence in fall months because of temperature differences ($\Delta T=T_s-T_a$). Flux errors may be introduced (~1%) if water surface temperatures are not accurately observed. The reflectivity of a water surface for atmospheric radiation is approximately 3 percent, consequently, reflected longwave radiation (L_r) is a small component of the surface radiation budget for lakes. As a result of small diurnal ranges in bulk air and water surface temperatures, both incoming and outgoing longwave radiation fluxes are relatively constant (Robinson, et.al. 1972). The net diurnal longwave flux ($L^*=L\downarrow-Lr-L\uparrow$) is negative (Figure 6).

Seasonal variation of radiation budget components for the Great Lakes is shown in Figure 7a. As expected, reflected components of both solar and longwave radiation are small relative to incoming fluxes. For all lakes, the incoming solar radiation is minimal during the winter months and maximal during June and July. Monthly mean K↓ on Lake Huron during May, June and July is greater than the other Great Lakes implying lower levels of cloudiness. Occurrences of overlake fog on Lake Ontario during May and June results in substantially lower K↓ compared to other lakes. Incoming longwave radiation shows a similar seasonal pattern for the Great Lakes ranging from 20-25 $MJm^{-2}d^{-1}$ during the winter months to 30 $MJm^{-2}d^{-1}$ in August. Lower L↓ values for Lake Superior result from lower air temperatures and humidities compared to the lower Great Lakes.

Outgoing longwave radiation losses for the Great Lakes is strongly related to the seasonal pattern of surface water temperature. During winter months, $L\uparrow$ is similar for all the lakes approximately 27 $MJm^{-2}d^{-1}$. During the summer months, $L\uparrow$ is lowest for cooler Lake Superior and is highest for shallow Lake Erie.

Monthly means of K^* for all of the Great Lakes is similar to the seasonal cycle of $K\downarrow$. Monthly mean L^* is negative with values ranging from -8 to -2 $MJm^{-2}d^{-1}$.

Direct measurements of Q^* over large lakes is costly and therefore depends on estimates of the individual components. Large errors are possible due to inaccuracies in estimating overlake meteorological conditions from land station data (Phillips and Irbe 1978). Figure (6) clearly shows that the diurnal net radiation budget is dominated by net solar radiation ($K^*=K\downarrow-K\uparrow$) during the day and is equal to net longwave radiation ($L^*=L\downarrow-Lr-L\uparrow$) at night. Robinson et.al (1972) concluded that the influence of L^* on Q^* is variable and generally small.

Comparison between Q^* determined from measurements (Davies and Schertzer 1974) and numerical model calculations (Atwater and Ball 1974) showed that overlake fog can significantly affect overlake estimates. During spring, unaccounted overlake fog had the effect of reducing weekly averaged net radiation estimates on Lake Ontario by as much as 40 percent. Variation in cloud height (surface to

The largest annual heat budget for world lakes is that of Lake Baikal, Russia (approx. 3.1 GJm^{-2}). Figure 9a shows heat incomes for the Great Lakes. Analysis of the thermal regime of Lake Superior (Bennett 1978; Schertzer 1978), indicated that the average Spring heat income of 1.47 GJm^{-2} accounted for a rise in mean lake temperature from a minimum of 1.4°C to the average temperature of maximum density 3.82°C while the summer income of 1.26 GJm^{-2} accounts for the remaining rise to the maximum mean lake temperature of 5.88°C. For Lake Superior, approximately 54% of the annual heat income is used for spring warming of the water to maximum density. In general, Figure 9a shows that as lake surface area, depth or volume increases, there is a rise in heat uptake and storage. The large heat incomes for the Great Lakes is related to the dimictic nature of these lakes, which allows the entire lake volume to be involved in heat exchange semi-annually.

The seasonal cycle of heat content for the Great Lakes is shown in Figure 9b estimated from Boyce et.al (1976), Schertzer (1978; 1987), and Bolsenga (1975). On the Great Lakes, minimum heat storage occurs in late winter while maximums occur in late summer/early fall. Substantial differences in the timing of maximum and minimum heat content in each lake is largely related to the variation in volume.

5.2 Advected Heat/ Hydrological Balances : Net advected heat (Q_A) is due to the heat content of water entering the lake (Q_1) from tributaries, runoff, precipitation and waste heat (Q_w) and water leaving the lake (Q_o) through evaporation and outflow and snow melt heat loss (Q_m). In contrast to large deep lakes, small shallow lakes/ponds may have a significant sediment heat budget requiring information on the structure and conduction characteristics of the sediments.

Since water has a low thermal conductivity, molecular transport of heat energy over small distances is negligible, and heat transport takes place almost entirely as a result of water movement. Whereas the transport of heat (Q_i-Q_o) is negligibly small in a soil column, the net heat advection due to flows can be significant especially for smaller lakes. Connecting channel flows of the Great Lakes are large (i.e. 5000-7000 m^3/s), however, the net impact of Q_i-Q_o on a whole lake basis is reduced due to the similar magnitude of T_i and T_o and is generally small compared to the lake heat content.

During summer, overlake precipitation adds heat to the lake, however, during late-fall or winter, frozen precipitation on open water affects the heat storage as the heat of fusion is extracted from the lake for melting. Heat gain or loss through ice formation or decay also affects the lake heat balance and ice cover has the effect of reducing heat losses through the surface (Schertzer, 1978, Pinsak and Rodgers 1981). Inclusion of precipitation within the energy budget of Lake Ontario decreased evaporation estimates by an average of 1 percent (April through October), and 5 percent during November through March during IFYGL. Omission of ice formation and decay changed evaporation in the range of 1 to -2.4 mm week^{-1} (Pinsak and Rodgers, 1981).

Waste heat inputs from point sources to the Great Lakes can represent a significant heat input on the local scale (Boyce et.al. 1993). Modifications to the seasonal heat storage can occur through artificial regulation of flows and flow changes due to introduction of ice booms etc.

For large lakes, the advective component represents a small contribution to the lake energy budget especially over longterm averages (Derecki 1975; Bolsenga 1975; Schertzer, 1978,1987; Elder et.al. 1974). For smaller bodies of water, net advection may represent a significant component of the energy budget.

5.3 Turbulent Heat Fluxes : Observations and models are used to determine Q^*, Q_S, Q_A, and Q_G leaving the residual $Q_H + Q_E$ (Figure 8b). Direct observation of the turbulent heat fluxes over spatial scales of large lakes is difficult and is not practical. Bulk aerodynamic methods are commonly used to estimate turbulent transfer components (Derecki 1975) in which bulk transfer coefficients for sensible heat and latent heat fluxes are generally assumed equal under near neutral conditions averaging 1.5 x 10^{-3} (± 20%) (Phillips 1973). Bulk transfer coefficients are observed to vary as a function of friction velocity, stability and sea state etc. (Henderson-Sellers, 1986).

Alternatively, energy partitioning of the residual (Q_H + Q_E) can be done by application of the Bowen ratio, (Appendix 1), from which the latent heat flux can then be determined based on independently derived overlake temperature and vapour pressure.

The Bowen ratio assumes that the transfer processes for heat and water vapour are similar. Difficulties over large lakes are related to estimation of overlake T_a and T_s and least confidence in values occurs for periods when water temperature approaches air temperature (Quinn, 1978). The Bowen ratio is very sensitive to small changes in moisture and temperature and daily estimates derived from hourly averaged ratios may vary significantly compared to estimates derived from hourly averaged input data (Pinsak and Rodgers, 1981). For temperature ranges encountered for large dimictic lakes, a critical value of β=-1 results in a very large (i.e. ∞) computed latent heat flux. Standard deviation of Bowen Ratios for Lake Ontario during IFYGL were found to be highest in the heating season (i.e. 1.4 to 19) and smallest during the cooling season (<0.6) (Pinsak and Rodgers 1981). Figure 10c shows wide variation in the climatological monthly mean β for the Great Lakes over the Spring months. On Lake Ontario 95% of the evaporation occurs during the cooling period (August through February). While the spatial and temporal representativeness of the Bowen ratio is of concern for very large lakes, the Bowen ratio technique is probably adequate for determining evaporation over longer time scales with least confidence on daily computations especially during the heating season (Pinsak and Rodgers, 1981).

Figure (10ab) shows monthly mean Q_H and Q_E for the Great Lakes. On an annual basis, Q_H is approximately 30-35% of Q_E on Lake Ontario (Pinsak and Rodgers, 1981) and L. Huron (Bolsenga 1975) and ranges from a low of about 5-10% for Lake Erie (Derecki 1975; Schertzer 1987) to 60% on Lake Superior (Schertzer 1978). Q_H should conform to the air-water temperature gradient in direction. Negative values of Q_E observed in all of the Great Lakes indicate episodes of condensation primarily during the spring months. Large errors in Q_H and Q_E occur under near neutral conditions when e_s approach e_a and when β approaches infinity. During the fall, decreased lake heat storage results primarily from increased heat losses through Q_H and Q_E and lower Q^*.

Under conditions of local advection through changes in air mass characteristics, Phillips (1973) observed that a dry air mass enhanced evaporation rate (because water to air humidity gradient is increased) and humid air suppresses it. Similarly the introduction of a relatively cold air mass enhances Q_H because it causes the water to air temperature profile to become more lapse and accordingly increases convective instability whereas a warm one has a dampening effect. Phillips (1973) reports an example of enhanced Q_H over Lake Ontario in January. At this time the lake water is considerably warmer than the cold continental air traversing it. Q_H may be as large as 20 $MJm^{-2}day^{-1}$ and since Q^* is very small at this time, the energy output to the atmosphere must be derived from the heat storage (Oke, 1983).

Figure 8b shows variation of the monthly mean total turbulent exchange on the Great Lakes in which positive values indicate net gain of heat to the atmosphere. In general, the total turbulent exchange is directed to the atmosphere during winter and fall months as the lake temperatures are higher than air temperature and evaporation is highest especially during the fall. Heat gains to the lake occur during the spring and early summer months as the air temperature is warmer than surface water temperature and heat gains also occur through condensation. The seasonal pattern of the total turbulent exchange is similar for all of the Great Lakes with the notable exception of shallow Lake Erie during the heating season. On Lake Erie, turbulent exchange losses are much greater due to higher water surface temperatures which has a large effect on losses through Q_E. Maximum losses of heat through turbulent transfer components occurs at the time of minimal gains through net radiation. Examination of the spatial distribution of turbulent heat exchange on Lake Ontario (Phillips 1973) showed substantial differences across the lake and as expected, patterns resembled spatial distribution of surface temperature.

5.4 Surface Heat Flux : The change in lake heat storage can be determined using detailed observations of the vertical temperature structure between two survey periods, (Boyce and Killins 1978). The limiting factor is availability of detailed temperature profile measurements over the lake. The alternative is to observe or compute the other components of the lake energy budget and to determine the surface heat flux as the residual.

Figure (8c) shows longterm monthly mean surface heat flux (ΔQ_S) for the Great Lakes. Positive changes in lake heat storage ΔQ_S (lake heat gains) occurs from February to September primarily through net radiation heating as Q_E and Q_H are relatively small. During the lake cooling period in the fall, negative changes in the lake heat storage for the Great Lakes occurs primarily through the turbulent exchange components (Q_E+Q_H) since net radiation (Q*) is small during these months. Maximum heat loss occurs in the period December through January and maximum heat storage occurs in late August and early September (Figure 9b) coinciding with the period in which the surface heat flux turns negative.

6. LAKE EVAPORATION

Water resources are important for such purposes as irrigation, navigation, industrial cooling water, hydro-electric potential and other socio-economic activities. Management of water resources (quantity and quality) is an important issue and water loss is an important concern in which physical climatological methods has played an important role. Evaporation is a common component linking the lake energy and hydrological balance (Figure 4). Large lake evaporation is primarily estimated through the water budget, energy budget, the empirical mass transfer or pan methods.

Both the water budget and energy budget approaches are generally more applicable for lakewide longterm evaporation averages. Where inflow and outflows are very large and similar, small errors may have significant impact on smaller components of the water balance such as evaporation (de Cooke and Witherspoon, 1981). The energy budget approach is relatively accurate and practical to use on lakes of moderate size with small net advection. Empirical techniques such as the mass-transfer and the evaporation pan are applicable for short time periods, however, may require calibration for each lake necessitating independent evaporation estimates. The advantage of the mass transfer approach is that relatively accurate evaporation rates can be derived from few simple measurements.

Comparison of monthly mean evaporation on Lake Ontario during IFYGL using three common methods (Figure 11a), shows similar seasonal patterns, however, except for a few months, differences in the evaporation estimate can be large. Considering that the IFYGL represented one of the most intensive investigations of a large lake, the differences highlight the difficulty in determination of evaporation from a large lake. The most apparent differences occurred in the Spring and Winter months. Significant condensation is indicated in May, June and July from the energy budget method with slight condensation from the aerodynamical approach while the water budget indicated significant evaporation.

6.2 Great Lakes Seasonal Evaporation : Seasonal distribution of evaporation for the Great Lakes (Figure 11b) shows highest evaporation during the late summer, fall and early winter (generally higher winds with large vapour pressure gradients) while minimum evaporation occurs during the late winter and early spring (ice cover suppression of evaporation and cold surface water temperatures). Summer and early fall evaporation for Lake Superior deviates significantly from the tendency shown by the other lakes of this system due primarily to surface water temperature differences. On temporal scales smaller than monthly, lake evaporation can be highly episodic. IFYGL buoy data indicates that 51% of evaporation occurred during only 15% of the days. Highest daily evaporation of 13mm occurred after the passage of a cold front on October 9, 1972 (Quinn and den Hartog, 1981) and 17% of the highest evaporation occurred on only eight days, when evaporation exceeded 7mm.

Spatial variability of evaporation on very large deep lakes can be very high and reflects the variability in the seasonal temperature distribution. Based on a dense buoy network, analysis of spatial variability of evaporation revealed an increase of evaporation from the west to the east of the lake ranging from 289mm to 488mm (Quinn and den Hartog, 1981).

6.3 Evaporation from Small Lakes : As indicated previously, lakes have a wide range of physical dimensions and latitudinal locations which affects the seasonal energy and evaporation regime. In general, shallow lake evaporation is controlled primarily by variations in the daily energy input and consequently, the evaporation seasons tend to correspond with the ice-free period from May through October. In large deep lakes (i.e. Great Lakes) maximum evaporation occurs over the fall months with energy derived from the heat storage from summer heating.

Lake Diefenbaker, Saskatchewan and Perch Lake, Ontario differ markedly from the physical dimensions of the Great Lakes. Lake Diefenbaker, created by damming of the South Saskatchewan River, is an elongate lake 225km long, averaging 2.5km in width and has a mean depth of 24m. In contrast, Perch Lake is a small circular lake (451,000 m^2) with a mean depth of 2m and a maximum depth of 3.7m situated near the Atomic Energy of Canada's laboratories at Chalk River, Ontario. Extensive studies have been conducted to determine energy and water budget and evaporation characteristics on both lakes.

Figure 12a shows the seasonal energy budget and evaporation for Lake Diefenbaker, 1973 (Cork, 1974). The energy budget of Lake Diefenbaker is considered typical of smaller lakes in the temperature an subarctic regions during the ice-free season (Rouse, 1979). Maximum Q* occurs during June and July, however, Q_E is maximal during the period August through October. During the months May through August, Q_H is small and negative indicating transfer of heat from the warmer air to the lake surface. After maximum heat storage (July-August) the warmer lake surface heats the overlying cooler air and Q_H is positive. Advection, while a small component, is important primarily during the spring months. Heat storage losses especially after August occur with decreases in Q* and high rates of losses through the turbulent transfer components, largely dominated by Q_E. The pattern of heat storage losses for Lake Diefenbaker (24m depth) is comparable (although lagged in time) to shallow Lake Erie (Figure 8c), as the energy used for evaporation during the August to October period is derived partly from the heat storage. The seasonal evaporation cycle is significantly shifted, peaking earlier in the year on Lake Diefenbaker compared to the Great Lakes.

Figure 12b shows the seasonal energy budget and evaporation for Perch Lake 1970 (Barry and Robertson, 1975). Measurements indicate that within two weeks after ice melt, Perch Lake heats rapidly with mean surface water temperatures exceeding mean air temperature. On this lake, mean water temperature continues to exceed mean air temperature throughout the summer and fall months and vertical temperature profiles show generally weak stratification for all months except in August near peak heat content. In a small shallow lake such as Perch Lake, heat storage not only increases rapidly in the spring and decreases rapidly in the fall, it also tends to oscillate through a narrow range throughout the summer and fall months in response to energy input. Q_H is positive throughout the summer and fall as the warm water surface heats the overlying air. Q_E is the dominant turbulent transfer component on Perch Lake and Figure 12b dramatically shows that nearly all of the absorbed incoming radiation (Q*) is expended in evaporation and heating the air passing over the lake. Cumulative curves (May through mid-October 1970) of the individual energy components (Barry and Robertson, 1975) showed that of the total energy input (Q*=1.62 GJ), 80% was used to evaporate water, 19% was used to heat the air and the remainder was accounted for by heating of the sediments and advection losses. The seasonal evaporation (mass transfer approach), while not as high as Lake Diefenbaker, is similarly displaced seasonally compared to the larger Great Lakes. For a shallow water body such as Perch Lake, the rate of evaporation averaged over a week or longer is nearly proportional to the net radiation at the surface.

6.4 Annual Lake Evaporation : Longterm mean annual evaporation from Great Lakes (Figure 13a) shows evaporation ranging from 500-750mm yr^{-1}. Highest evaporation occurs in warmer and shallower Lake Erie while lowest evaporation occurs on lakes with cooler surface temperatures. The wide range in evaporation over each of the Great Lakes is a reflection of weather variability.

Mean annual lake evaporation across Canada (Figure 13b) was summarized by Phillips (1990). Lake evaporation is greatest in the dry Prairies, where it can be as much as 1000mm a year. Lowest evaporation occurs in the extreme north, averaging around 100mm as a consequence of suppression due to nine or more months of ice and snow cover. Prowse and Ommanney (1990) reported that for much of the Canadian North, lake evaporation is equal to or greater than 50% of precipitation.

The Pacific coast has the longest evaporation season with annual losses averaging from 400 to 700 mm. At high elevations in the Cordillera, the season is short and lake evaporation is low because of the colder temperatures and longer ice season. July is the month when maximum evaporation occurs in the south, and August is the peak month in the Arctic islands. On the Prairies evaporation losses are high in May. Across eastern Canada lake evaporation ranges from about 400mm in the north to 800 mm in southwestern Ontario. In the east there is a regular seasonal cycle with a fairly well-marked maximum in July for smaller lakes.

7. PHYSICAL CLIMATOLOGY AND LAKES RESEARCH

An understanding of the physical climatology of large lake systems (momentum transfer, heat and moisture exchange) plays a fundamental role in assessing important environmental water quantity and quality concerns. Lake modelling investigations have incorporated aspects of physical climatology for assessment of eutrophication, the effects of acid rain, the fate and pathways of toxic contaminants and recently the potential effects of climate change on water resources. Monitoring the longterm physical climatology of large deep lakes and physical changes such as thermal stratification responses has been suggested as one method for detection of regional changes in climate. Continuing research is required to adequately specify surface exchange coefficients for momentum, heat and moisture exchanges under variable surface conditions. Spatial and temporal scales of analysis are often dictated by the adequacy of overlake meteorological fields. Physical climatological research on the Great Lakes has often been large-scale, providing essential physical components for models of the circulation, temperature structure or water quality. In such cases, flux computations are generally determined through use of occasional ship surveillance (with aliasing problems) and through use of stationary buoy or shoreline data. A major problem with shoreline data is the requirement to apply transformations to derive overlake values. Current methods are generally lake specific and require improvement. Recent use of satellite data for such variables as surface water temperature has greatly improved flux computations in the Great Lakes.

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FIGURES

Figure 1. (a) Selected large lakes in Canada ordered according to surface area: (b) Location of selected large lakes and major drainage basins: (c) Profile of the Great Lakes system.

Figure 2. Monthly mean surface water temperatures of the Great Lakes (Based on Phillips 1990).

Figure 3. Examples of measurement systems deployed in large lakes (a) Meteorological Buoy and (b) Tower platform.

Figure 4. Energy and water budget components of lakes.

Figure 5. Lake surface albedo (a) Albedo in response to cloud, sea state and zenith angle, (b) Range of observed mean daily surface albedo. (Based on Nunez et.al. 1971).

Figure 6. Diurnal variation of Lake Ontario radiation budget components for cloudless conditions on 28 August 1969. (After Davies et.al. 1970)

Figure 7. Seasonal variation of Lake Ontario radiation budget components for (a) solar radiation components, and (b) for net solar and longwave radiation. Sources : L. Superior (Schertzer

1978); L. Huron (Bolsenga 1975); L. Erie (Derecki 1975; Schertzer 1987) and L. Ontario (Atwater and Ball 1974).

Figure 8. Long term monthly mean (a) net radiation, (b) total turbulent exchange and (c) surface heat flux for the Great Lakes. Sources : L. Superior (Schertzer 1978); L. Huron (Bolsenga 1975); L. Erie (Derecki 1975; Schertzer 1987) and L. Ontario (Atwater and Ball 1974).

Figure 9. Heat storage in the Great Lakes, (a) Heat income per unit surface area versus mean depth (after Bennett 1978) and (b) Great Lakes seasonal heat content cycle (Based on Boyce et.al (1976), Schertzer (1978; 1987), and Bolsenga (1975).

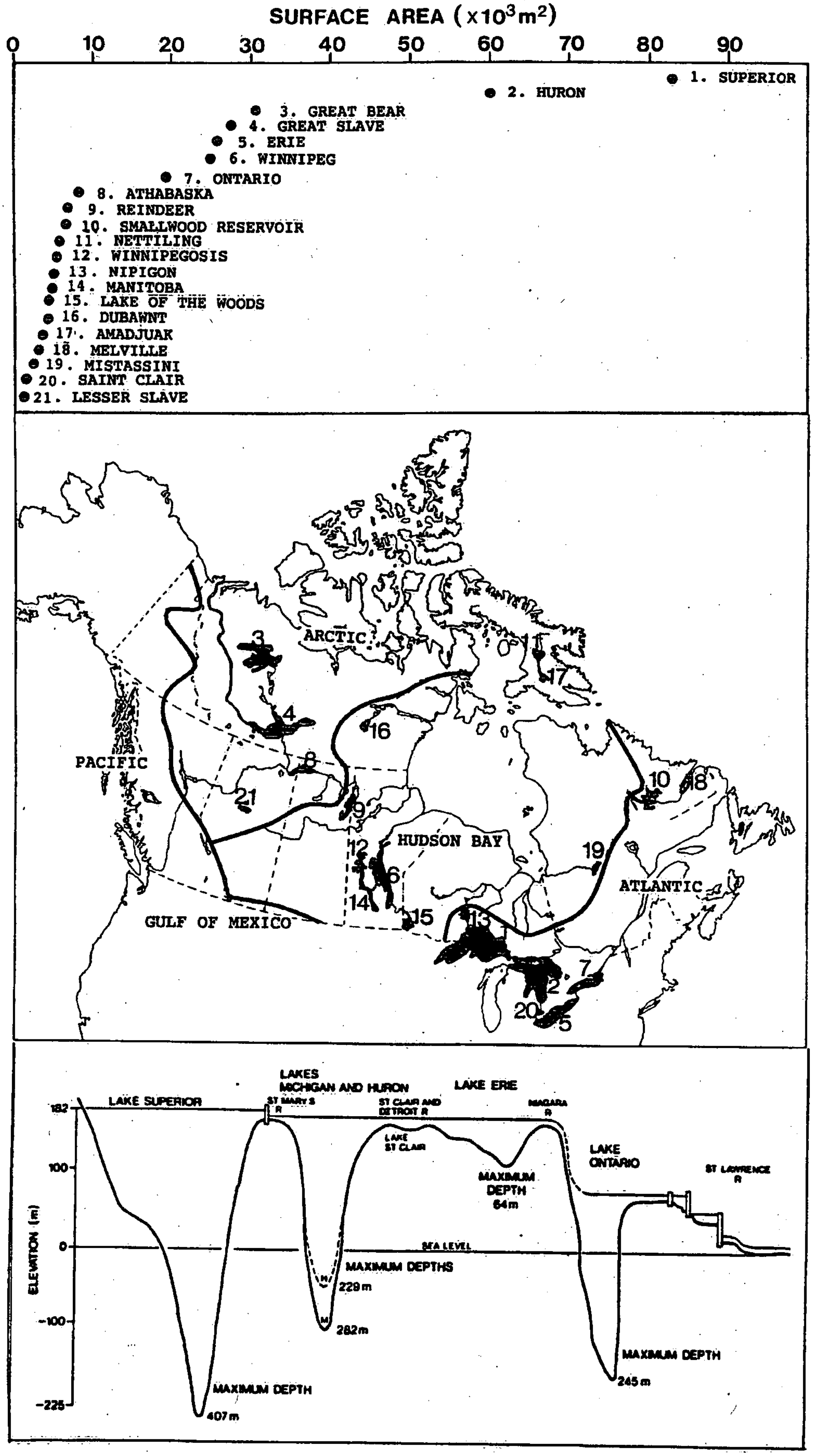
Figure 10. Long term monthly mean sensible and latent heat flux and computed Bowen Ratio for the Great Lakes. Sources : L. Superior (Schertzer 1978); L. Huron (Bolsenga 1975); L. Erie (Derecki 1975; Schertzer 1987) and L. Ontario (Atwater and Ball 1974).

Figure 11. Monthly mean evaporation from the Great Lakes depicting (a) comparison of evaporation estimates for Lake Ontario (IFYGL) from the water budget, energy budget and mass transfer approaches, and (Pinsak and Rodgers (1981), and (b) long term monthly mean mass transfer evaporation (Atmospheric Environment Service).

Figure 12. Seasonal energy budget and evaporation of Lake

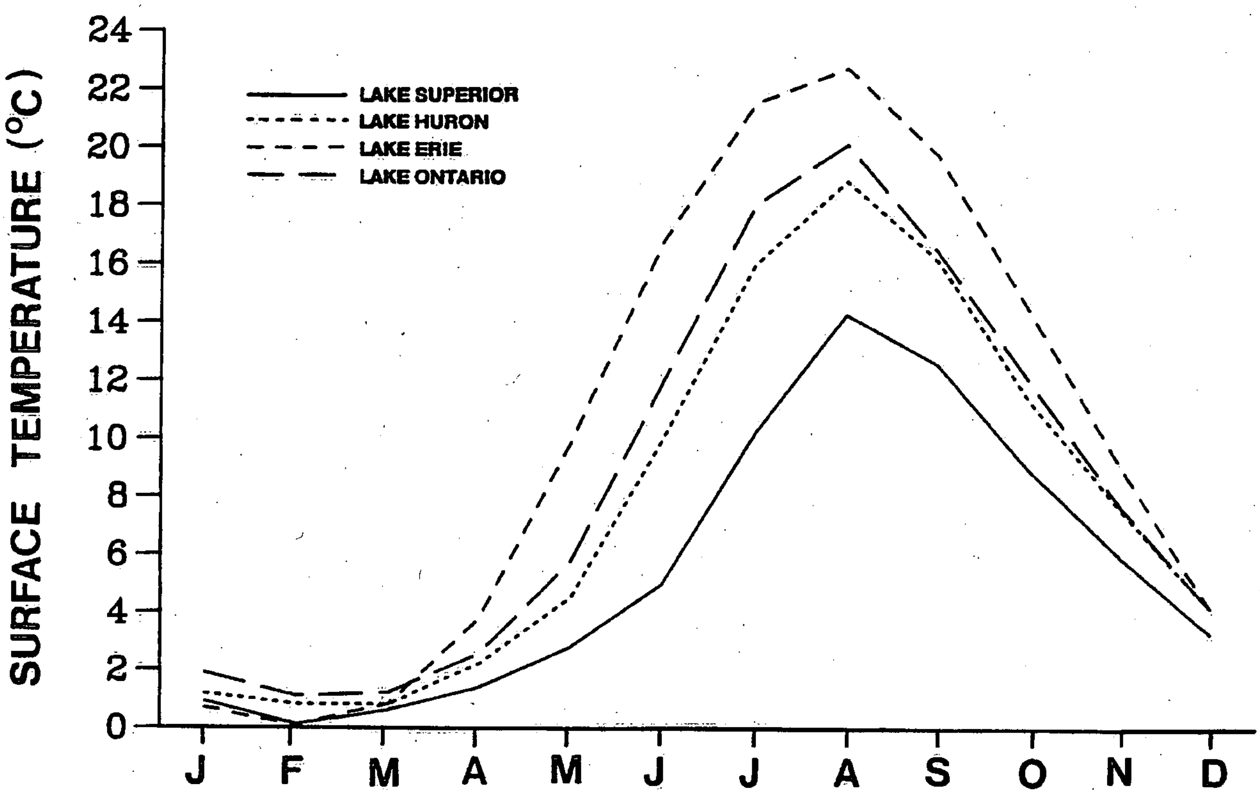
Diefenbaker, Saskatchewan and Perch Lake, Ontario (data abstracted from Cork, 1974 and from Barry and Robertson 1975).

Figure 13. Average annual lake evaporation (mm) for (a) the Great Lakes by mass transfer technique (Atmospheric Environment Service), and (b) for Canada estimated using pan evaporation (Phillips, 1990).



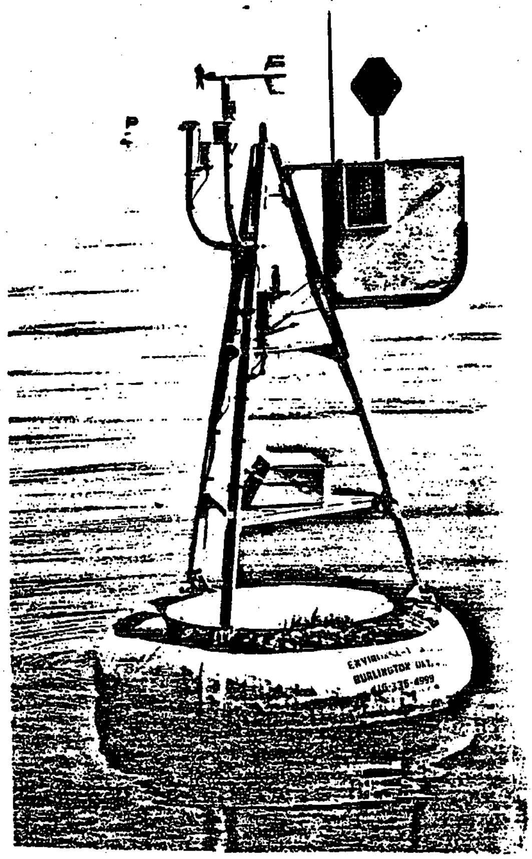
SURFACE AREA (x10³m²)
0
10
20
30
40
50
60
70
80
90
1. SUPERIOR
2. HURON
3. GREAT BEAR
4. GREAT SLAVE
5. ERIE
6. WINNIPEG
7. ONTARIO
8. ATHABASKA
9. REINDEER
10. SMALLWOOD RESERVOIR
11. NETTILING
12. WINNIPEGOSIS
13. NIPIGON
14. MANITOBA
15. LAKE OF THE WOODS
16. DUBAWNT
17. AMADJUAK
18. MELVILLE
19. MISTASSINI
20. SAINT CLAIR
21. LESSER SLAVE
ARCTIC
PACIFIC
HUDSON BAY
ATLANTIC
GULF OF MEXICO
LAKES MICHIGAN AND HURON
LAKE ERIE
LAKE SUPERIOR
ST MARY'S R
ST CLAIR AND DETROIT R
NIAGARA R
LAKE ST CLAIR
LAKE ONTARIO
ST LAWRENCE R
MAXIMUM DEPTH 64m
SEA LEVEL
MAXIMUM DEPTHS
229m
282m
MAXIMUM DEPTH
245m
MAXIMUM DEPTH
407m
ELEVATION (m)
182
100
0
-100
-225

F1



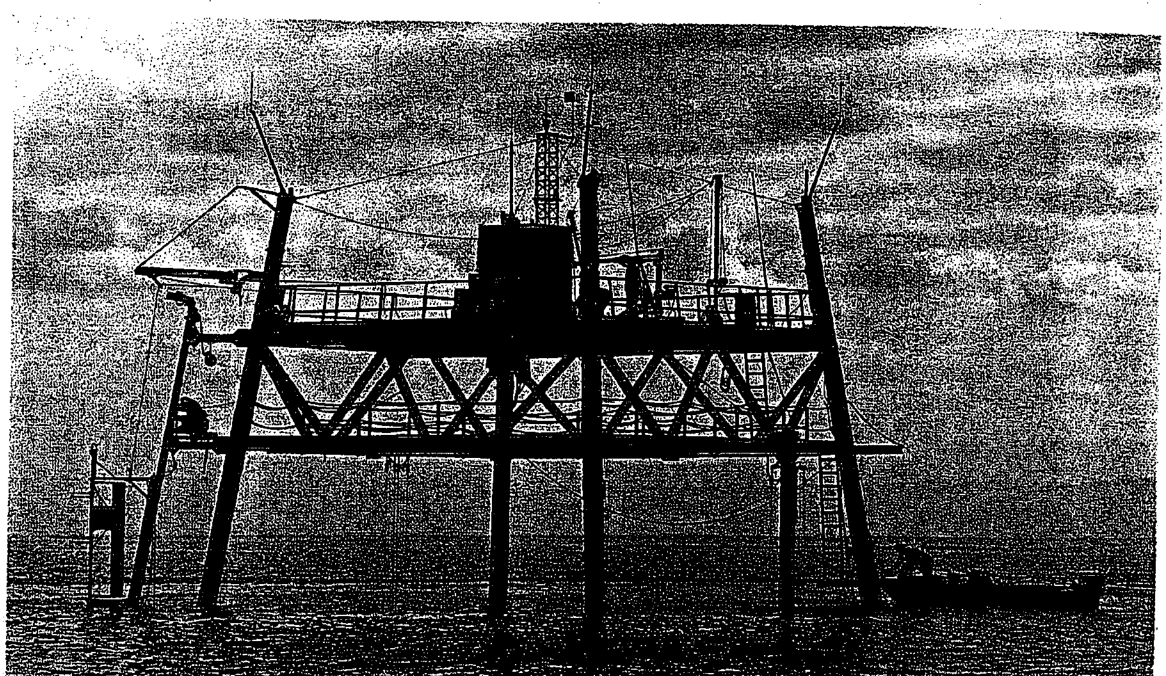
SURFACE TEMPERATURE (°C)
24
22
20
18
16
14
12
10
8
6
4
2
0
LAKE SUPERIOR
LAKE HURON
LAKE ERIE
LAKE ONTARIO
J F M A M J J A S O N D

F2

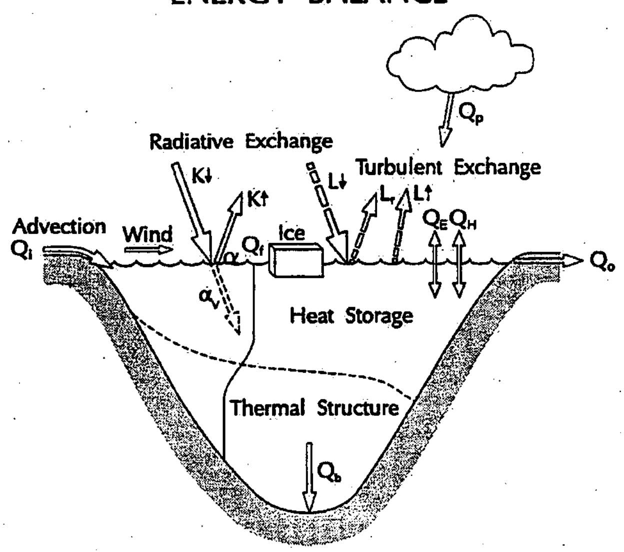


Note: Final prints can be made of these or other photos.

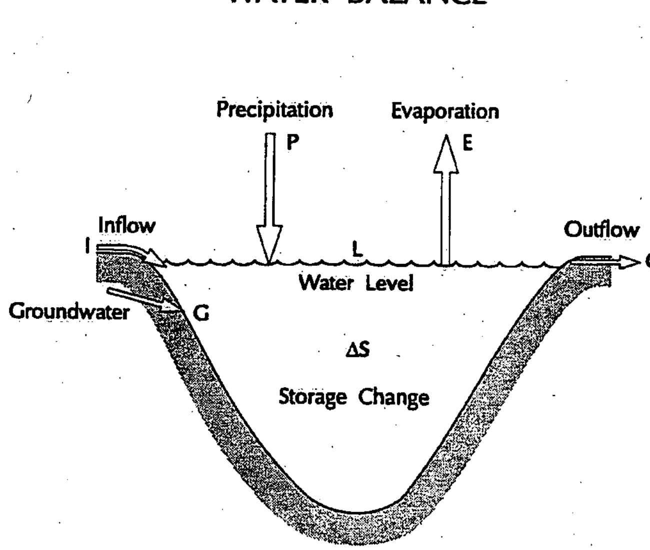
Perhaps some instrumentation can be labelled.

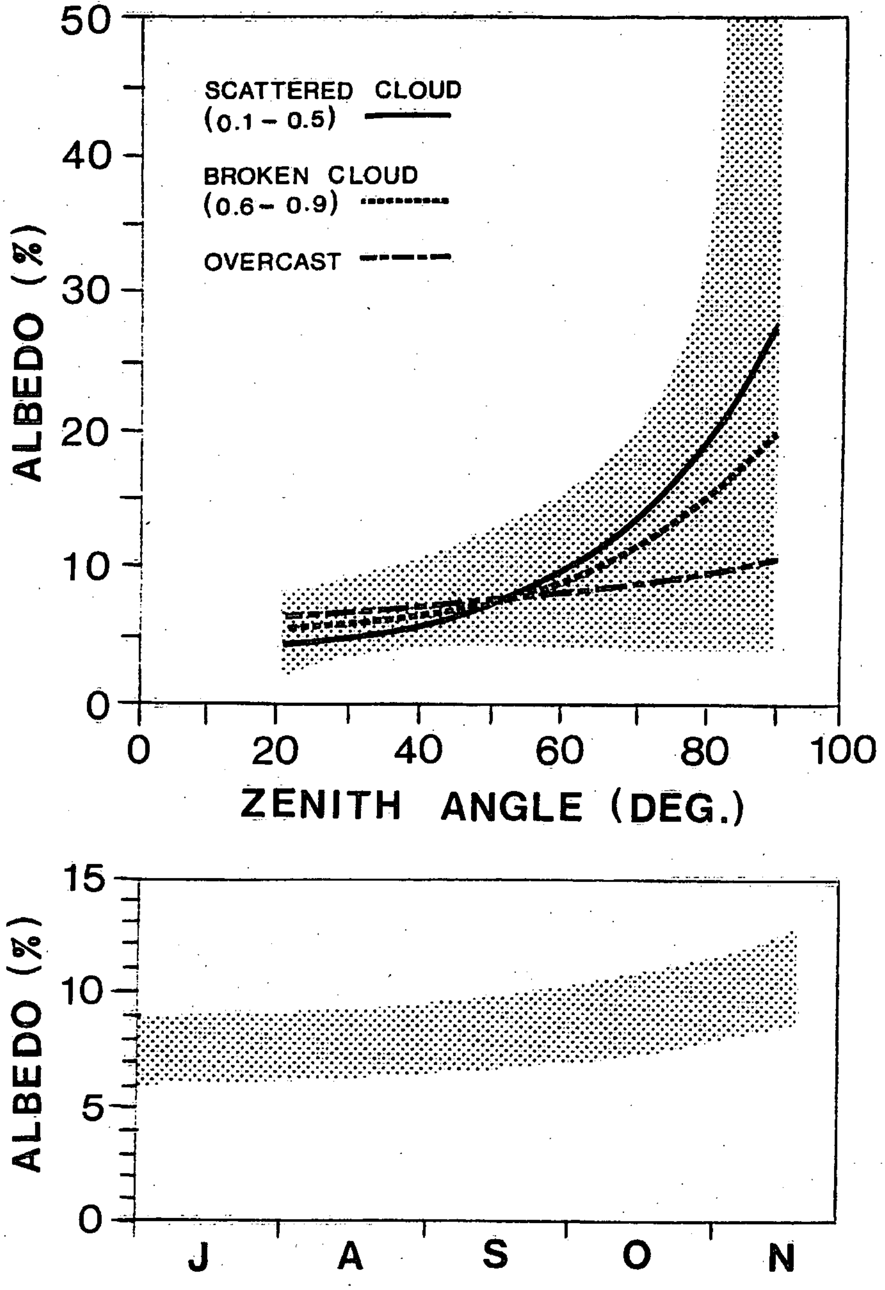


ENERGY BALANCE

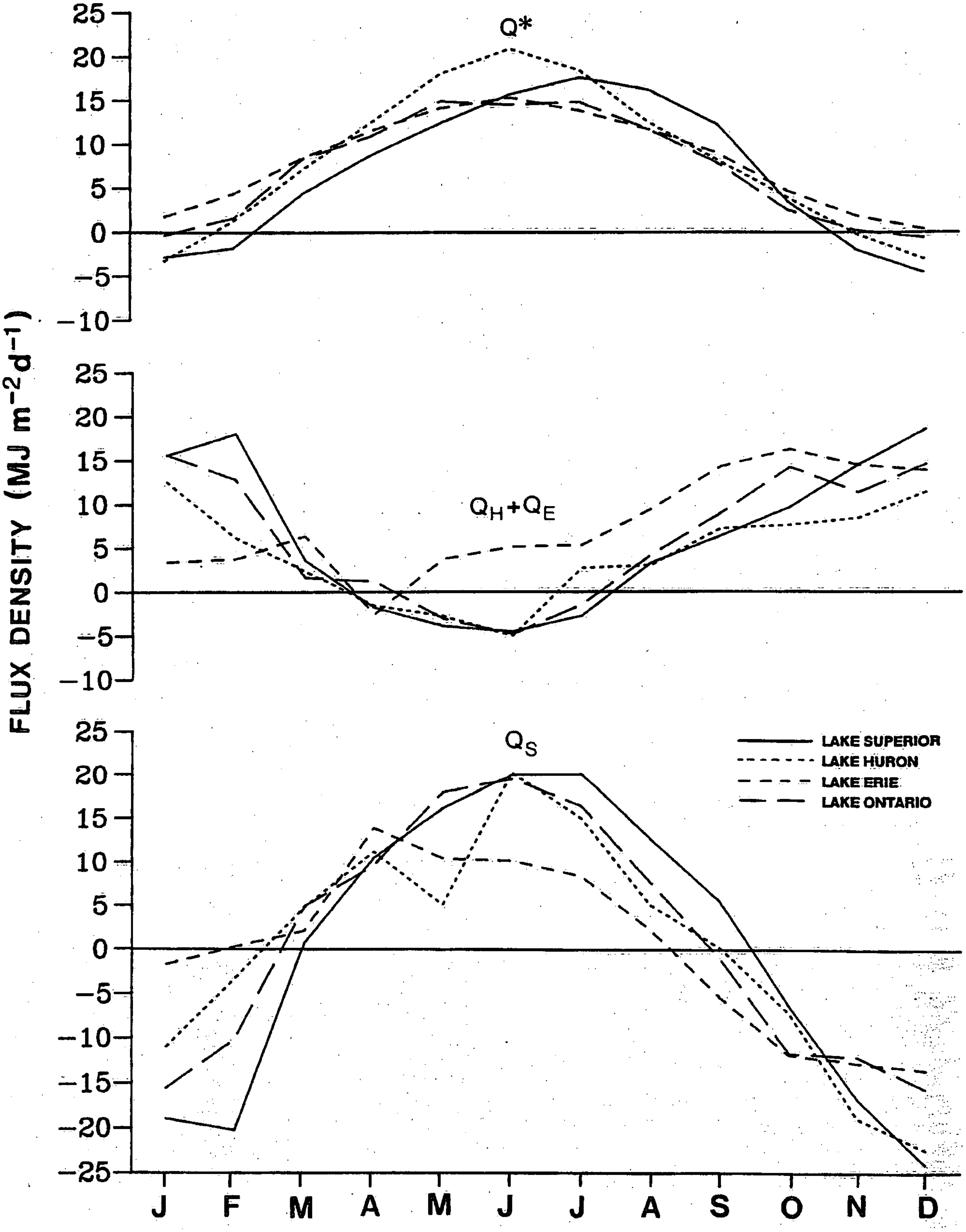


WATER BALANCE



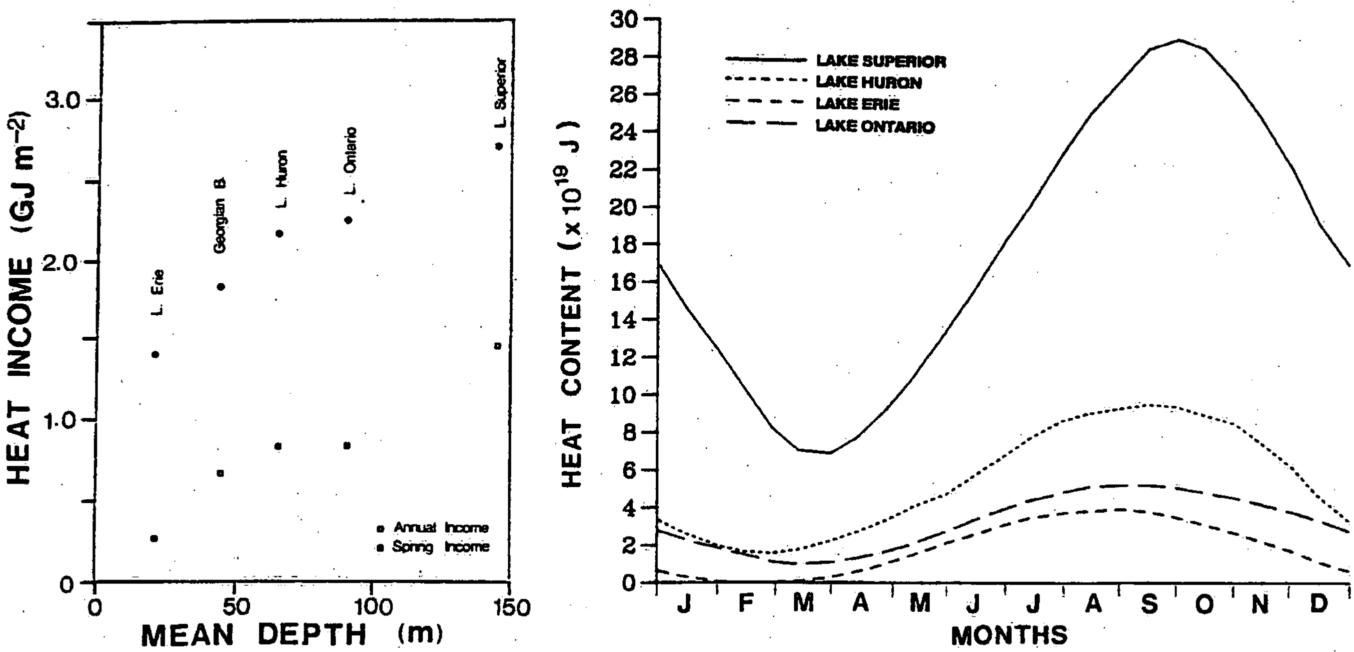

50
40
30
20
10
0
ALBEDO (%)
SCATTERED CLOUD
(0.1 – 0.5)
BROKEN CLOUD
(0.6 – 0.9)
OVERCAST
0
20
40
60
80
100
ZENITH ANGLE (DEG.)
15
10
5
0
ALBEDO (%)
J
A
S
O
N

F5

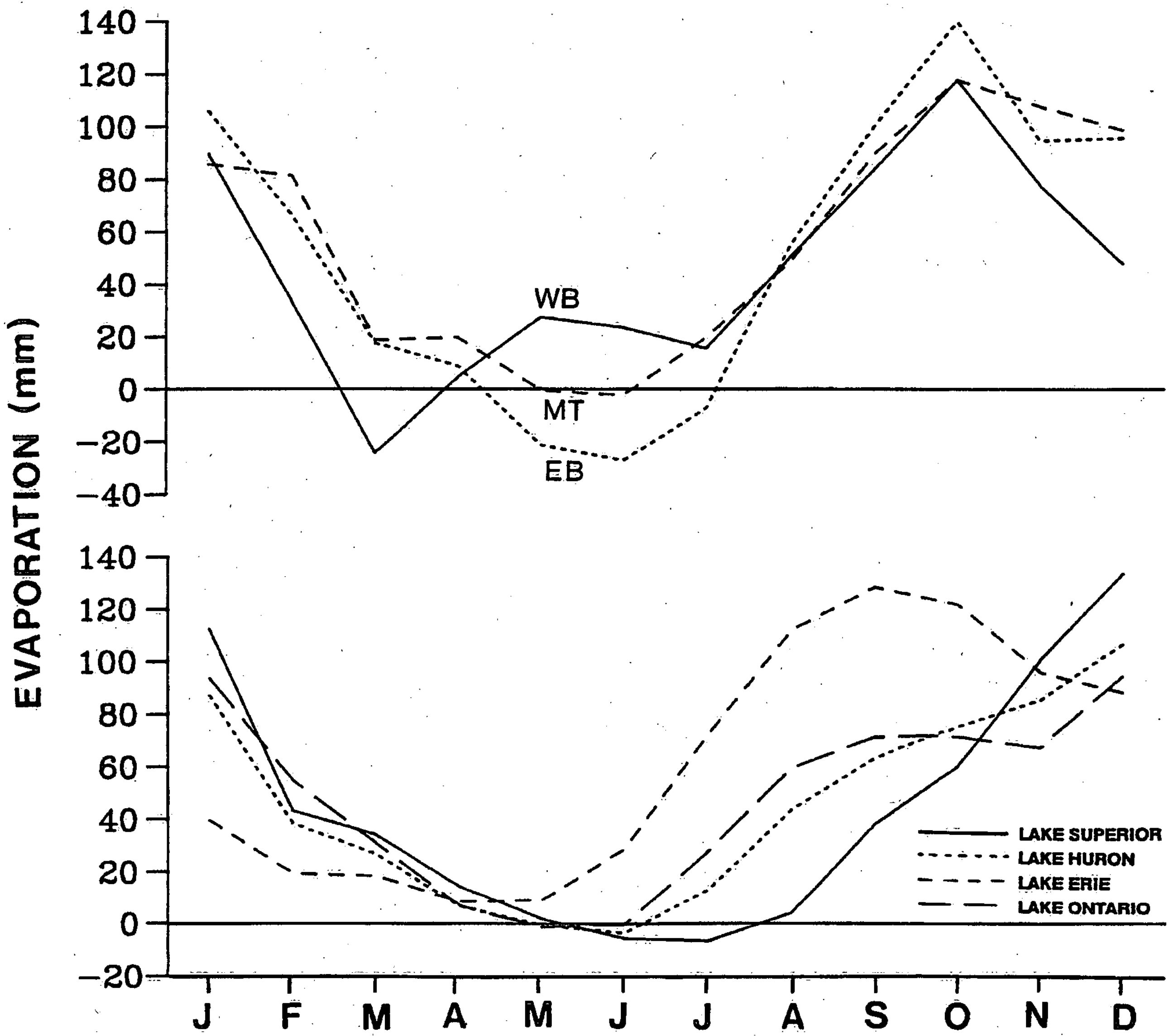


FLUX DENSITY (MJ m-2 d-1)
Q*
QH+QE
QS
LAKE SUPERIOR
LAKE HURON
LAKE ERIE
LAKE ONTARIO
25
20
15
10
5
0
−5
−10
−15
−20
−25
J F M A M J J A S O N D

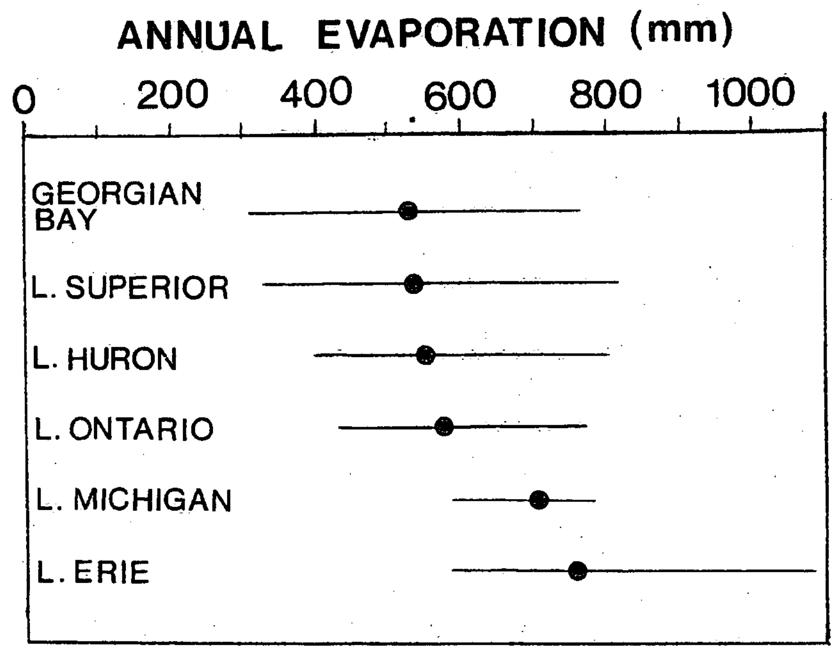
F8

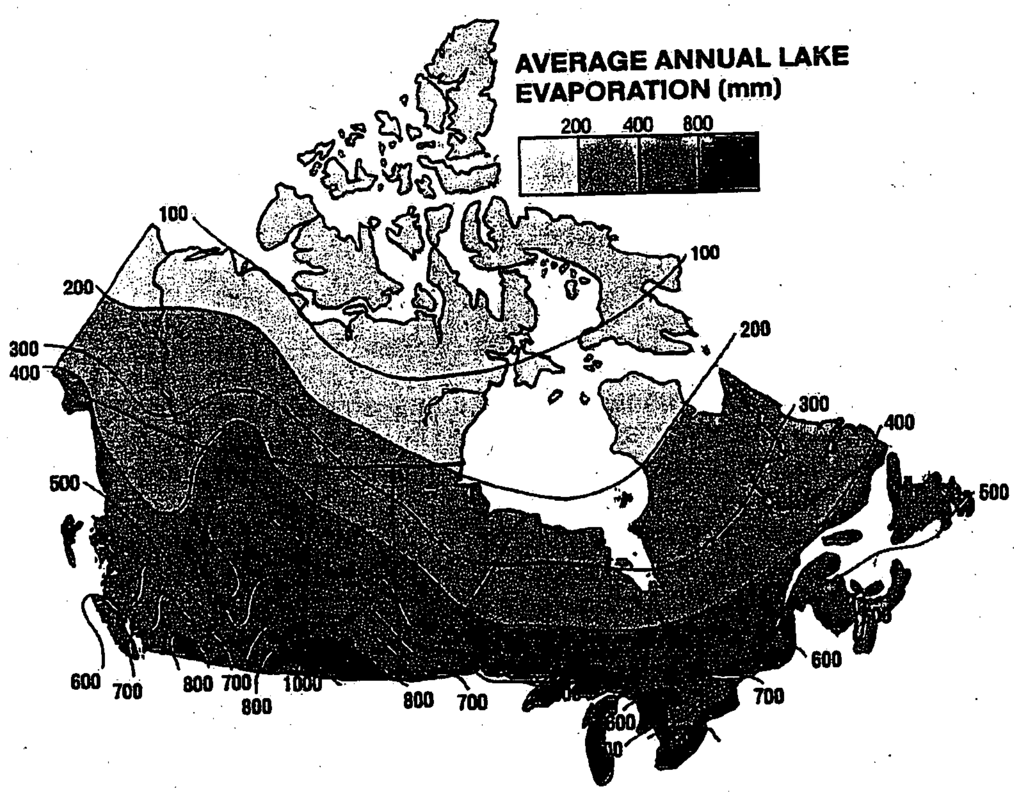

HEAT INCOME (GJ m^{-2})
MEAN DEPTH (m)
L. Erie
Georgian B.
L. Huron
L. Ontario
L. Superior
Annual Income
Spring Income
HEAT CONTENT (x 10^{19} J)
MONTHS
LAKE SUPERIOR
LAKE HURON
LAKE ERIE
LAKE ONTARIO
J F M A M J J A S O N D

F9


EVAPORATION (mm)
140
120
100
80
60
40
20
0
−20
−40
WB
MT
EB
140
120
100
80
60
40
20
0
−20
LAKE SUPERIOR
LAKE HURON
LAKE ERIE
LAKE ONTARIO
J F M A M J J A S O N D

F11


ANNUAL EVAPORATION (mm)
0
200
400
600
800
1000
GEORGIAN BAY
L. SUPERIOR
L. HURON
L. ONTARIO
L. MICHIGAN
L. ERIE


AVERAGE ANNUAL LAKE EVAPORATION (mm)
200
400
800
100
200
300
400
500
600
700
800
1000

Environment Canada Environnement Canada

Canada

Canada Centre for Inland Waters
P.O. Box 5050
867 Lakeshore Road
Burlington, Ontario
L7R 4A6 Canada

National Hydrology Research Centre
11 Innovation Boulevard
Saskatoon, Saskatchewan
S7N 3H5 Canada

St. Lawrence Centre
105 McGill Street
Montreal, Quebec
H2Y 2E7 Canada

Place Vincent Massey
351 St. Joseph Boulevard
Gatineau, Quebec
K1A 0H3 Canada

Centre canadien des eaux intérieures
Case postale 5050
867, chemin Lakeshore
Burlington (Ontario)
L7R 4A6 Canada

Centre national de recherche en hydrologie
11, boul. Innovation
Saskatoon (Saskatchewan)
S7N 3H5 Canada

Centre Saint-Laurent
105, rue McGill
Montréal (Québec)
H2Y 2E7 Canada

Place Vincent-Massey
351 boul. St-Joseph
Gatineau (Québec)
K1A 0H3 Canada