

An Overview of the Characteristics of Lake Superior Meteorology, Hydrology and Physical Limnology

William M. Schertzer* and Yerubandi R. Rao

*Water Science and Technology Directorate, Environment
Canada, Canada Centre for Inland Waters, 867 Lakeshore Rd.,
Burlington, Ontario L7R 4A6*

**Corresponding Author: william.schertzer@ec.gc.ca*

Keywords: climatology, hydrology, energy budget, thermal regime, mesoscale wind forcing, coastal dynamics, circulation, eddies, climate impacts

Introduction

The bulk of physical limnological research on Lake Superior has been focused on establishing baseline characteristics and advancing knowledge on the dynamics which also have relevance and impact on ecosystem components. There have been a number of larger-scale investigations of the lake. These include the Upper Lakes Reference Study (ULRS), the Keweenaw Interdisciplinary Transport Experiment (KITES), and the International Upper Great Lakes Study (IUGLS). In addition, monitoring of meteorological and hydrological variables such as water levels and flows as well as lake surveys through the Great Lakes Surveillance Plan continues to provide time-series data necessary for understanding and managing this freshwater resource.

*State of Lake Superior
Edited by M. Munawar
© 2008 Ecovision World Monograph Series
Aquatic Ecosystem Health and Management Society*

The Upper Lakes Reference Group study (Munawar, 1978) was the result of the 1972 Great Lakes Water Quality Agreement in which the International Joint Commission (IJC) undertook a massive effort to assess the impact of economic and land use activities on Lakes Huron and Superior. This was an important large-scale investigation with a goal to describe the baseline physical, chemical and biological conditions of the lake. With respect to the topics in this paper, the ULRS on Lake Superior included studies related to environmental climatology and hydrology, thermal-heat budget and heat storage characteristics, water transparency and hydrodynamics.

The Keweenaw Interdisciplinary Transport Experiment (KITES), funded by the National Science Foundation (NSF) as part of the Coastal Ocean Processes Program (CoOP), began in 1997 and included researchers from six institutions that conducted one of the largest studies ever undertaken on Lake Superior. The five-year study was focused on the Keweenaw Current, one of the strongest currents of its kind in the world. The KITES project investigated the physical processes that control the position and strength of the Keweenaw current, how the current affects the distribution of nutrients, and therefore, plankton and fish, in a 150-mile study region. The investigation was important because the current is the primary means by which materials are transported from the lake's western to eastern basins, and is therefore likely to be important in processes throughout Lake Superior.

The International Upper Great Lakes Study (IUGLS) is a four-year study which began in 2007 to examine hydrological issues of the Great Lakes. Specifically, the IJC is interested in determining whether the regulation of Lake Superior outflows can be improved to address the evolving needs of the upper Great Lakes from Superior to Niagara Falls, Ontario. Major topics for investigation include determining the factors that affect water levels and flows, developing and testing potential new regulation plans and assessing the impacts of these potential plans on the ecosystem and human interests.

Through these and smaller-scale investigations, there has been advances in concepts, observational techniques and the development and applications of models to simulate current responses and to project potential impacts due to changing environmental conditions. The object of this paper is to provide an overview of both past and current research investigations on Lake Superior. The focus is to

describe relevant characteristics of the physical regime such as: climatology, hydrology, thermal structure, air-water heat exchanges, water movements and hydrodynamic modelling. These processes can impact on the state and transport of biological and chemical constituents in the lake. Also included is a brief discussion of some of the current research on the potential impact of climatic changes on Lake Superior.

Lake Superior physical characteristics and measurement network

Lake Superior has the largest surface area (82,100 km²) of any freshwater lake in the world and is the third largest in volume (Table 1). Its trophic status is oligotrophic with some indicators suggesting ultra-oligotrophic conditions. The lake itself comprises nearly 40 percent of the basin area and it is situated entirely on the Canadian Shield (Matheson and Munawar, 1978). The international boundary between Canada and the United States partitions the lake and basin with approximately 35 % of the lake surface and 66% of the basin area within Canada. The lake has a maximum depth of 406 m and a volume of 12,100 km³ with a total annual water level fluctuation in the order 1.2 m.

Sly and Thomas (1974) summarized the post glacial development of the Great Lakes Basin. A series of lake stages with varying water levels occurred in conjunction with glacial retreats, ice fronts and isostatic adjustments. The outlet sill at Sault Ste. Marie was formed about 2000 B.P. and established Lake Superior in its present form (see Thomas and Dell, 1978). The geometry of the lake is similar to that of a shallow oceanic basin (Fig. 1). The average slope of the

Table 1. Lake Superior physical dimensions (based on Gray et al., 1994).

Surface Area	82,100	km ²	Average Outflow	2,240	m ³ /s
Volume	12,100	km ³	Residence Time	170	years
Maximum Depth	406	m	Seasonal Fluctuation ¹	0.3	m
Mean Depth	147	m	Total Fluctuation	1.2	m
Shoreline Length	4,400	km	Catchment Area ²	128,000	km ²

¹ regulated, ² not including the lake surface area.

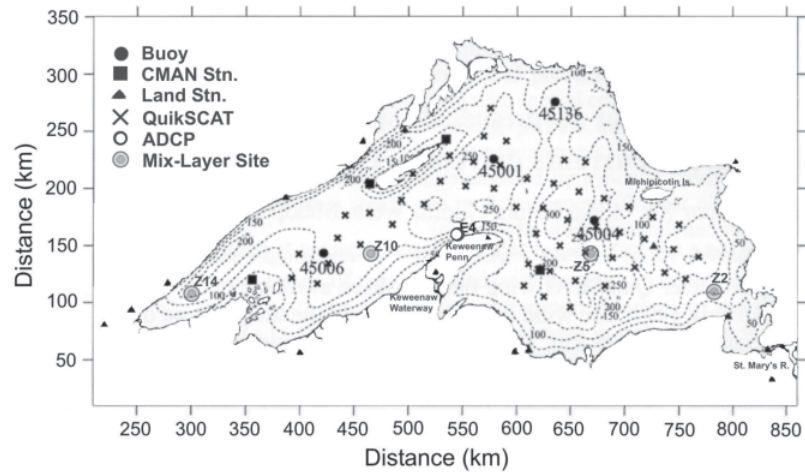


Fig. 1. Bathymetry of Lake Superior and an example of the locations of meteorological stations, lake buoys and specialized observations. (modified from Chen et al., 2004a).

shelf is about 1 m per 100m on the northern coast and about 0.5m – 3.5m per 100m on the southern coast (Chen et al., 2001). The narrowest shelf is located along the Keweenaw Peninsula where the slope exceeds 0.5m per 100 m.

The basin has a relatively dense meteorological observation network. During the ULRS, the spatial resolution of meteorological observations was one station for every 1000 km² for some elements (Phillips, 1978). Weather observations from islands, lighthouses, and shoreline stations and almost 80,000 weather reports from commercial and research vessels were used to describe the climatology of Lake Superior. In addition, meteorological buoys have been established in the lake and are operational during the ice-free period. Figure 1 shows an example of the spatial distribution of shoreline meteorological sites, lake buoys and specialized observations applied on the lake based on a recent investigation (Chen, 2004).

Environmental climatology

Lake Superior has a dominating influence on the hydroclimatology of the Great Lakes region owing to its immense size. The overlying

atmosphere plays a major role in affecting the magnitude and seasonal variability in many lake physical processes. The lake itself has its own climatological characteristics and responses to the local weather and extremes. A cross-section of selected meteorological and lake processes of significance for physical limnology and ecosystem components of the lake is provided below.

Selected meteorological components and climatology

The Lake Superior basin does not have a spatially homogeneous climate (Phillips, 1978). The climate of the sub-basins is influenced by regional and local environmental factors such as latitude, topography, lake effects, surface cover, anthropogenic activities, and proximity to the lake. The lake spans 7 ° of longitude, 85° W to 92°W, and 2.5° of latitude, 46.5 °N to 49.0 °N. In general, the highest elevation is less than 600 m and more than half of the region is under 300 m elevation and as such there are no major topographical barriers affecting the regional climate. The large volume and heat storage of the lake is a dominant influence on the lake climate, for example, moderating air temperature, augmenting or suppressing precipitation, influencing the formation of fog and the magnitude of over-lake wind speed.

Investigations conducted during the ULRS provided baseline characteristics of key meteorological variables (e.g. Table 2). Phillips (1978) provides an in-depth description of the seasonal variation of meteorological observations as well as an approximation of the spatial distribution of climatological variables based on the database available at that time.

Atmospheric stability, fog, high winds and waves

The historical data, based on combined airborne radiometric temperature and vessel bucket temperature observations, indicate that Lake Superior is warmest during late summer reaching a maximum lakewide monthly mean temperature of 12 °C in mid-September. Cooling in the fall months proceeds slowly compared to the shallower Great Lakes and the mid-winter has surface temperatures < 1°C. Differences between air temperatures on land and on the lake can vary diurnally, seasonally, and during storm events. Based on air-water temperature difference, Phillips (1978) derived a simplified

Table 2. General characteristics of selected key monthly means of over-lake variables (air temperature, surface water temperature, dewpoint, wave height and wind speed) based on 1973 experiments (Phillips, 1978). Symbol * refers to missing monthly mean.

Over-Lake		J	F	M	A	M	J	J	A	S	O	N	D
air temp.	°C	-7.6	*	*	3.1	5.7	8.7	12.6	15.3	12.8	8.4	2.1	-3.3
water temp.	°C	2.3	*	*	1.4	2.6	4.1	7.4	11.3	11.5	9.0	5.9	4.1
dewpoint	°C	*	*	*	0.1	0.3	2.0	8.5	10.9	9.5	1.7	0.4	-4.6
wind speed	m/s	9.8	*	*	8.1	7.5	6.8	6.6	6.5	7.8	8.6	9.3	9.7
waveheight	m	2.7	*	*	1.7	3.5	1.1	1.0	1.3	1.9	2.3	2.3	2.5

atmospheric stability index ranging from very unstable through very stable conditions. Neutral or stable conditions tend to dominate from March through October. Very stable air often occurs from May through August. High frequency occurrence of fog is common in June and July which results in condensation on the surface of the lake. Fog occurs on average 10% of the time on Lake Superior. Unstable conditions are seen to begin in November and continue through to March. Data from 1973 suggested that northwest to southwest winds blow about 40% of the time averaging 8 to 15 m s⁻¹. Intense cyclonic storms on Lake Superior can generate winds in excess of 30 m s⁻¹. Historical data indicate the occurrence of extreme 1 minute peak winds, some of the highest being recorded on ship at > 40 m s⁻¹. Wind magnitude and direction influence the generation of waves. Significant wave height, defined as the average height of the highest one third of all the waves sampled, indicates that for Lake Superior, waves in excess of 6 m are more probable in November (approximately 1% of the time) and low waves less than 0.5 m are more favourable in July occurring 70% of the time (Phillips, 1978).

Recent meteorological and limnological observations have included data from NOAA and Canadian meteorological buoys. Satellite observations are also available, and can be used to estimate lake surface temperature during cloud-free conditions. Figure 2 provides an example of the magnitude and temporal variation of daily averages of air temperature, water surface temperature, wind speed and relative humidity, based on intensive observations from an over-lake meteorological buoy on Lake Superior in 2005. Over-lake observations are critical for development and verification of lake physical models and are essential for forecasting of lake conditions.

Impact of an extreme wind event

Hultquist et al. (2006) provided a re-examination of the 9-10 November, 1975 intense storm on Lake Superior which produced extremely hazardous winds with wind gusts in excess of hurricane force, and wave conditions linked with the sinking of the Edmund Fitzgerald (U.S. Coast Guard, 1977). High resolution numerical simulations were conducted to assess the likely wind and wave conditions during the storm. Simulations suggested that weather conditions deteriorated rapidly during the afternoon over the lake and

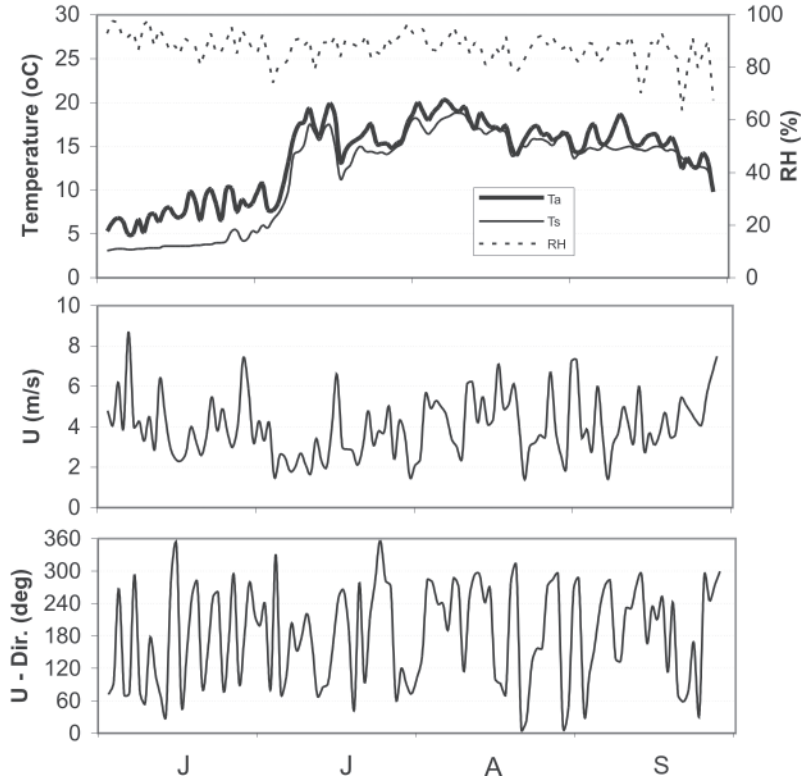


Fig. 2. Variation of air temperature, water temperature, wind speed and relative humidity from in situ meteorological buoys in Lake Superior during 2005 experiment.

by evening, sustained winds near 25 m s^{-1} encompassed southern Lake Superior with local sustained winds in excess of 30 m s^{-1} . These winds generated waves in excess of 7.5 m. The waves moved from west to east across southeast Lake Superior and, in the case of the Edmund Fitzgerald, were nearly perpendicular to its reported track. Hultquist et al. (2006) indicate that based on wind climatology, the winds experienced during this storm were not typical, but neither were they a rare occurrence. Evaluation of model simulations against observed wind and waves indicated that the models were capable of capturing the synoptic scale conditions associated with the storm very well. Modern forecast techniques have significantly improved warnings of such extreme events. In this case, high-resolution

simulated output served as a close approximation of observational data on the lake throughout the storm and was of sufficient quality to drive wave model simulations which produced output representative of the marine conditions.

Turbidity and horizontal eddy diffusivity

The level of turbidity of the water column among other factors, impacts on the depth that light in the photosynthetic wavelengths (400 – 700 nm) will penetrate into the water column and thus impact aquatic constituents. Knowledge of the mean vertical attenuation coefficient in a water column is also required as a governing variable within thermal models.

In Lake Superior, clay turbidity plumes have been observed in the extreme west arm region (Shuter et al., 1978). This is particularly evident during the spring runoff period from rivers which transport suspended material. Sediment loads generally consist of red clay particles which may average 1 micron in diameter, where a large percentage of the load (e.g. 80%) may be less than 4 microns in diameter. Such particles can be used as natural tracers from which to compute the horizontal eddy diffusivity. Shuter et al. (1978) investigated the dispersion within turbid areas ranging from 2.5 – 5.4 km in diameter and having average velocities of 0.4 to 6.1 cm s⁻¹. The coefficient for horizontal eddy diffusivity ranged from 0.6 – 1.5 m² s⁻¹ with average values of 1.1 m² s⁻¹. Such values are comparable to those found in other Great Lakes (Rao and Murthy, 2001).

Light attenuation and photic depth

Knowledge of the chlorophyll *a* concentration in the lake is important since it is an indicator of the level of productivity. Generalizing the chlorophyll *a* concentration over a lake the size of Lake Superior is difficult based on only a small number of discrete samples taken several times a year (e.g. on lake surveillance surveys). Bukata et al. (1978) developed a semi-empirical model from which both the suspended inorganic and organic (represented by suspended mineral and chlorophyll *a*) concentrations of a water mass could be determined through simultaneous measurements of the optical transmission and near surface volume reflectance of the water column. The volume reflectance is defined as the ratio of the upwelling to the downwelling

irradiance at a given point. Comparison between predictions of chlorophyll *a* concentration from the optical model and measurements from the Great Lakes showed excellent agreement for waters of Lake Superior, while comparisons were less satisfactory for lakes of greater optical complexity.

Secchi disk depth (*S*) has been a routine historical observation in the Great Lakes and transmissometer data (*T*) have also been utilized to approximate mean vertical extinction coefficient. Bukata *et al.* (1988) formalized relationships among Secchi disk depth, beam attenuation coefficient and irradiance attenuation coefficients for the Great Lakes. The relationship between total attenuation coefficient, $c = 2.85(S^{-1})^{0.80}$, and the irradiance attenuation coefficient (k_{PAR}) for Lake Superior was $k_{PAR} = 0.195C + 0.08$. Linear relationships between k_{PAR} and *S* were also derived for Lake Superior over different Secchi depth ranges. For example the following relationship $k_{PAR} = 0.67S^{-1} + 0.10$, was valid over a range of Secchi disk depths ($3\text{m} < S < 10\text{m}$).

Using 15 lake zones (Fig. 3a), representative of nearshore and offshore regimes, (Schertzer *et al.*, 1978) determined the mean vertical light extinction coefficient (MVEC) based on Vollenweider (1955) (Fig 3c) and photic depths for Lake Superior. An example of the spatial distribution of MVEC is given in Fig. 3b for the spring period May 12-24, 1973. Based on survey data over the year, considerable spatial and temporal variability in water transparency between areas and, in some lake zones was found with large departures from the lakewide mean. MVEC ranged from approximately 0.15 to 0.25 m^{-1} for the majority of zones and relatively good qualitative correspondence was observed with seasonal changes in chlorophyll *a* concentrations. Considerably lower rates of light attenuation occurred in the deeper mid-lake zones where computed photic depths characteristically ranged from 20 to 30 m. The photic depth (Fig. 3d) is defined as the depth at which 1 percent of the surface irradiance intensity in the photosynthetic wavelengths would be received in a water column. Reduced water transparency generally predominates along coastal areas (e.g. Duluth, Thunder Bay, the southwest red clay zone etc.). Photic depths in these areas are generally less than 20 m throughout the photosynthetic production period. In the vicinity of Duluth, mean photic depth values of less than 10 m were observed with extreme minimums approaching 2 m.

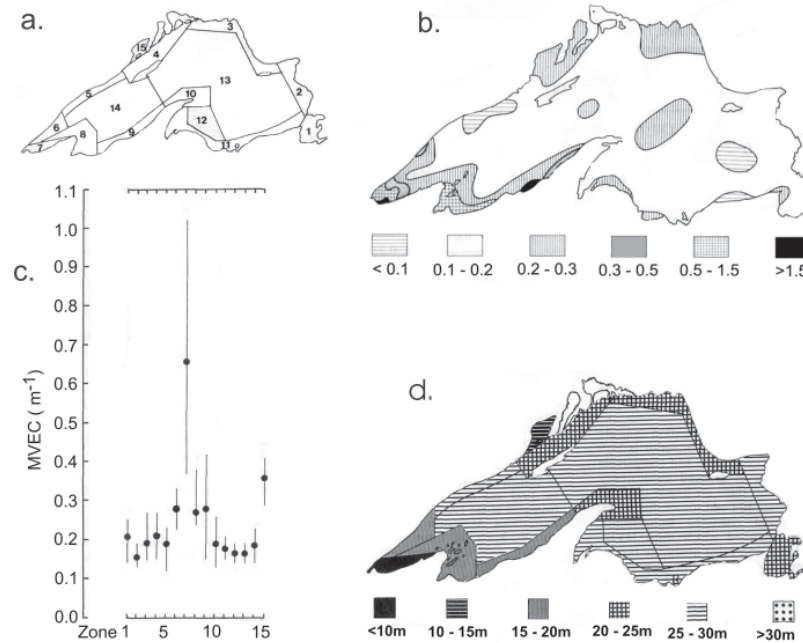


Fig. 3. Mean vertical extinction coefficient and photic depth in Lake Superior in 1973 showing (a) 15 lake zones representing nearshore and offshore regimes, (b) spatial distribution of MVEC for high runoff period May 12-24, 1973, (c) mean and range of MVEC in the lake zones, and (d) mean potential photic depth in the 15 zones (based on Schertzer et al., 1978).

General hydrological characteristics

Lake Superior is the headwater of the Laurentian Great Lakes system with discharges through the St. Mary's river influencing hydrological conditions in the downstream lakes. Understanding the magnitude and fluctuation of the major hydrological components of Lake Superior is essential for managing the freshwater resource and for regulation of its outflow. Variations in inflows and outflows affect water levels and impact on such socio-economic factors as navigation, but also the aquatic ecosystem components located especially in the nearshore zones. Much research has been conducted in terms of monitoring water levels, understanding the lake water budget and evaporation, and the development of regulation plans.

Long-term water level fluctuations

There are nearly 200, mostly small rivers which are tributary to Lake Superior. The largest tributary to the lake is the Nipigon River in Canada which drains a sub-watershed of 4,850 km³ which includes Lake Nipigon (Matheson and Munawar, 1978). Tributaries are the second largest source of water to the lake. Annual water level fluctuations on the lake occur as a result of the inflow from tributaries and precipitation and losses through evaporation from the lake surface and outflow through the St. Mary's River. Man-made structures at the major outflow can also affect water level in the lake.

Clites and Quinn (2003) provide a history of Lake Superior water level and regulations. Prior to 1888, the St. Mary's River was naturally controlled (Hartmann, 1988). In 1901 the Canadian Lake Superior Power Company constructed "compensating works" consisting of four sluice gates and Canadian ship and power canals were added in 1914, resulting in significant reductions in outflow. The Boundary Waters Treaty of 1909 between Canada and the United States established the International Joint Commission (IJC) and the Orders of Approval of 1914. The treaty directed the IJC to ensure that there were no "uses or obstructions" that would deleteriously impact on domestic, sanitary, navigation, power and irrigation purposes.

The extreme natural range of Lake Superior levels based on the period 1860 to 1913 was from 182.8 to 183.8 m. In an effort to moderate water level, outflow regulation through the St. Mary's River was instituted (IJC, 1979) with initial limits of 183.4 to 183.86 m referenced to the International Great Lakes Datum, 1985 (IGLD85). Later, these limits were adjusted based on the 1900 to 1976 period to 182.76 to 183.86 m. The series of regulations that were applied, including the current regulation Plan 1977-A has altered the levels and flows of Lake Superior and, consequently, the lower Great Lakes. Figure 4 shows the long-term (1918-2008) mean and range of Lake Superior water level.

The regulation of Lake Superior has prevented some very low levels from falling lower, but resulted in some higher levels (10-20 cm) compared to pre-regulation natural levels. Some research has suggested that the use of Lake Superior as a storage basin to prevent downstream flooding creates as many problems as it solves (Clites and Quinn, 2003). Review of regulation of Lake Superior is in progress

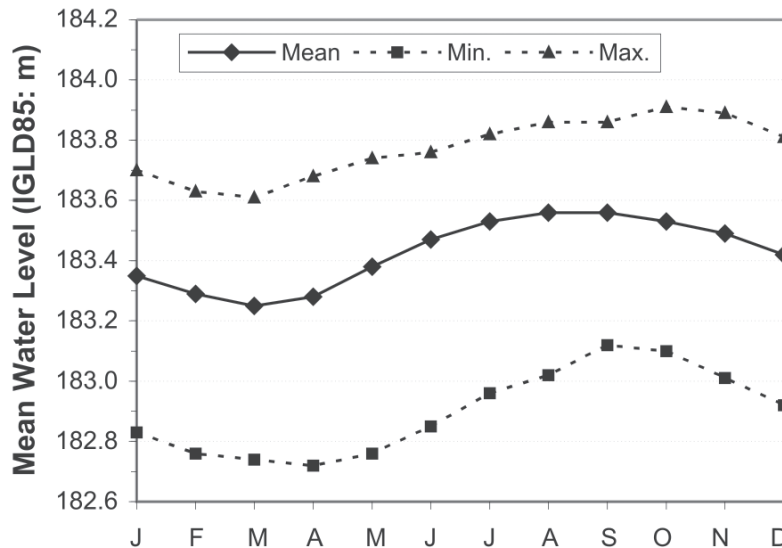


Fig. 4. Long-term (1918-2007) variation of the mean and range water level in Lake Superior referenced to IGLD85.

and offers an opportunity to consider interests and concerns that were not explicitly considered in the current regulation plan.

Water budget

Lenters (2004) examined the longterm water budget of Lake Superior. Figure 5 shows longterm monthly mean water budget components for the period 1948 to 1999 based on Hunter and Croley (1993). In terms of seasonal cycle, evaporation shows the largest amplitude of all water budget components ranging from a maximum of 11.7 cm in December to a minimum of -0.4 cm in June. The second largest component is runoff which has a maximum input of 9.6 cm in May and a minimum of 3.2 cm in February. Over-lake precipitation has a moderate seasonal cycle with maximum values in June through September and minimum values during the winter. The net basin supply has considerable seasonal variability since evaporation is out of phase with precipitation and runoff.

Outflow through the St. Mary's River is the largest contributor to the annual water budget at 86.1 cm, however, the seasonal range

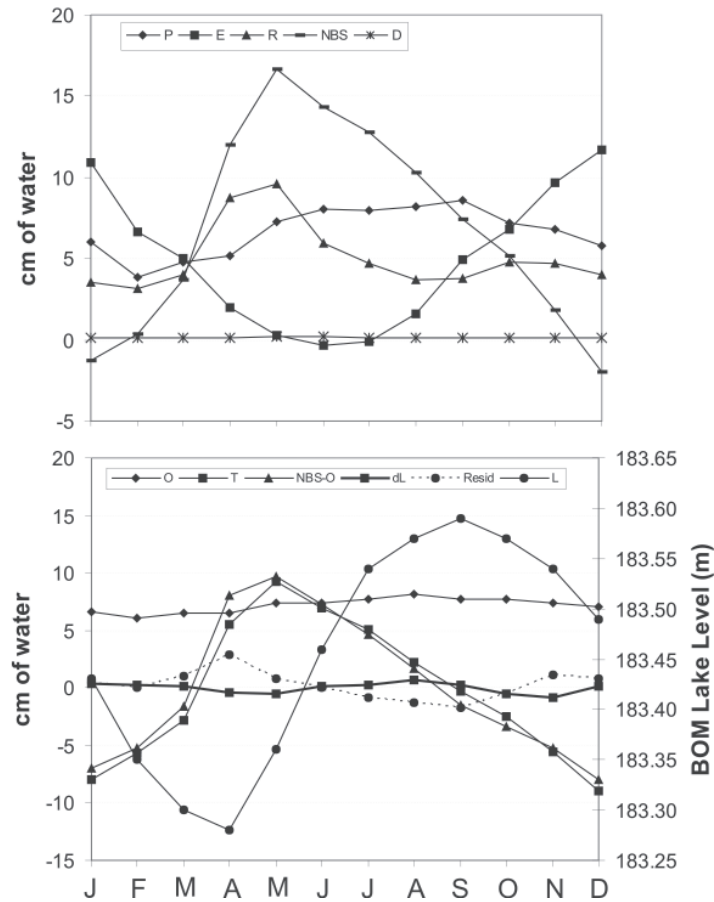


Fig. 5. Longterm monthly mean water budget for Lake Superior (1948 – 1999). Fluxes are in units of cm of water per month (1 cm/month $\sim$ 320 m³/s (based on Hunter and Croley, 1993; Lenters, 2004). P = precipitation, E = evaporation, R = runoff, NBS = net basin supply, D = Long Lac Diversion, O = outflow, T = thermal expansion, NBS-O = change in lake level, resid = residual term, and L = lake level above IGLD85. dL is change in BOM (Basis of Measurement) lake level.

of values is small compared to evaporation, precipitation or runoff. Influx from the Long Lac Diversion is small at 1.6 cm of lake level (Lenters, 2004). Thermal expansion of freshwater is an approximate cyclic quantity (Bennett, 1978b). The maximum expansion of 0.7 cm occurs in August during the highest heat content, while the maximum contraction 0.9 cm takes place during November.

Bennett (1978b) found that relatively high rainfall and land drainage to Whitefish Bay can result in a reduction of flow from the main body of Lake Superior which can impact on the concentration of materials in Lake Superior and water entering the St. Mary's River. Lake Superior has the longest flushing time from the Laurentian Great Lakes at 170 years (Table 1). A comparison between the longterm (Lenters, 2004) and the single year water budget in 1973 (Bennett, 1978b) shows a similar pattern of the water budget components. Estimated annual total inflows ($R + P$) is 140.4 cm (longterm) and 139.3 cm (in 1973). Estimated total outflows (St. Mary's River + E) is 145.2 cm (longterm) and (132.2) in 1973; the difference here being related to lower discharge of the St. Mary's River in 1973.

Lake evaporation

Evaporation is an important term within the hydrological budgets of lakes. Several methodologies are available from which to compute evaporation varying from simple to complex methods (e.g. Schertzer and Taylor, 2008). One of the major challenges is to provide lake-representative inputs to the evaporation models. This is particularly critical for lakes with large surface areas and volumes such as Lake Superior.

Initial estimates of the evaporation from Lake Superior were conducted using mass transfer (e.g. Snyder, 1960; Richards and Irbe, 1979; Lofgren and Zhu, 2000), by water budget (e.g. Morton, 1967; Bennett, 1978a), and by energy budget (Schertzer, 1978). Evaporation derived by the energy budget (Schertzer, 1978) was completed during the ULRS in 1973. Monthly mean water budget evaporation (Bennett, 1978a) is determined as a residual from the water budget, and consequently, its determination incorporates all of the possible errors from the water budget components. Earlier mass transfer studies by Snyder (1960) and Richards and Irbe (1979) applied formulations similar to those derived from the Lake Hefner investigations (e.g. Harbeck, 1962) although Richards and Irbe (1979) introduced lake-land ratios to modify land data to over-lake values. Derecki (1981) derived monthly evaporation over a 34-year period (1942-1975) based on an improved mass transfer method (Quinn, 1979; Quinn and den Hartog, 1979) derived during the intensive observations conducted in

the International Field Year for the Great Lakes (Aubert and Richards, 1981). More recent computations conducted by Lofgren and Zhu (2000) are similar to the mass transfer approach except that it incorporates diffusion coefficients that depend on the static stability and wind shear in the atmospheric boundary layer.

Figure 6 shows a comparison between mean monthly evaporation estimates. Immediately apparent is that mass transfer evaporation (Snyder, 1960 and Richards and Irbe, 1969) differs significantly from the energy budget, water budget and modified mass transfer approaches in May to August and especially in July when the over-lake atmospheric boundary layer is very stable. Based on the other studies, the annual evaporation for Lake Superior is in the order of 50 cm. Research on measuring representative lake evaporation for Lake Superior has begun through application of fast response eddy covariance methods (P. Blanken, Department of Geography and Environmental Studies University of Colorado at Boulder, Boulder, CO., per. comm., 2008) as part of the IUGLS investigation.

Thermal structure

In a general sense, the annual temperature cycle of Lake Superior is a response to the climate / weather / hydrological conditions. The magnitude and variability of the temperature cycle and associated dynamical processes in a large deep lake such as Lake Superior is fundamental for understanding lake physical, biological and chemical processes. Much research has been conducted on Lake Superior to define the base characteristics of the surface temperature, ice extent, the thermal bar dynamics, and understanding the summer and winter thermal structure.

Surface temperature

Surface water temperatures on Lake Superior have been observed using vessel surveys, airborne radiometer, and satellite observations. Most of the historical observations of surface temperature have been conducted during the spring and summer months (e.g. Millar, 1952; Richards et al. 1969; Irbe, 1972; Feit and Goldberg, 1976).

Since the Great Lakes have such expansive surface areas, surveys

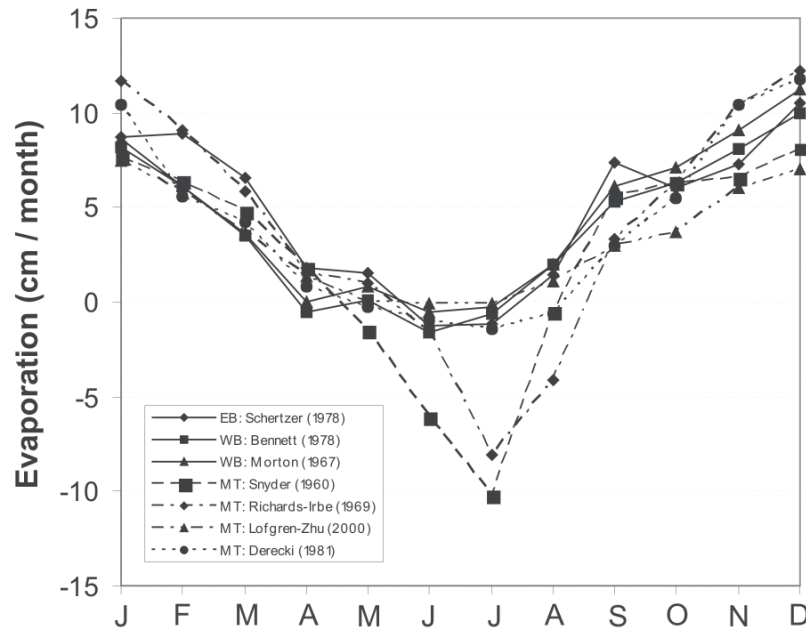


Fig. 6. Comparison of evaporation estimates for Lake Superior based on energy budget, water budget and mass transfer methods.

by vessels may take as long as a week or more and can introduce error through aliasing of the observations. Although in situ observations are indispensable for developing critical time-series of water temperature, spatial representation is often limited. Satellite observations offer the ability to remotely sense surface temperatures over large spatial areas. On Lake Superior, Budd et al. (1999) examined spatial water surface temperatures using AVHRR (Advanced Very High Resolution Radiometer) satellite imagery provided through the NOAA CoastWatch Program.

Schwab et al. (1999) developed a procedure for producing daily cloud-free maps of surface water temperature in the Great Lakes including Lake Superior based on satellite-derived AVHRR. The AVHRR data are processed to generate a 512 x 512 pixel image of the entire Great Lakes region on a Mercator geographic projection at ~ 2.6 km resolution and provide as complete as possible coverage of the lakes on a daily basis by using previous imagery to estimate temperature in cloud covered areas. Based on comparisons between

the satellite-derived water surface temperature estimates and eight NOAA weather buoys distributed over the Great Lakes, differences were less than 0.5 °C at each location. The root mean square differences range from 1.1 to 1.8 °C. The analysis procedure provides a more realistic depiction of the spatial distribution of temperature especially in spring and late fall when traditional methods using meteorological buoys or lake surveillance are not available. Introduction of ice cover digital data from the National Ice Centre based on satellite imagery, aircraft reconnaissance and surface observations is anticipated to improve estimates of mean lake surface temperatures during winter months when measurements are not available in the historical observations.

Ice cover

Ice cover on the Great Lakes has a considerable economic (e.g. affecting the length of the shipping season) and environmental impacts (e.g. shoreline damage). Aerial ice reconnaissance of the Great Lakes, including Lake Superior, was processed to develop composite maps of ice coverage (e.g. Rondy, 1969). Assel et al. (1983) have synthesized the data to provide statistics on the ice characteristics and ice cover extent on the lakes. Figure 7 shows the maximum, minimum and normal ice concentration distribution pattern for Lake Superior for nine half-month periods starting December 16-31 (D2) and ending April 16-30 (A2) based on a 20 year period (1960-79).

Interannual variability in the ice cover of Lake Superior can be large (Assel et al., 1983). The ice cover can vary from 0% to 90 – 100% ice over the lake for most of the 9 half-month periods (Fig. 7) depending on the severity of the winter. In normal winters, ice is usually confined to bays and harbours at the perimeter of the lake through the end of January with little lakeward advance of the ice boundary until February. During the first half of February, ice forms in the western basin of the lake and in the eastern basin ice may extend some 20-40 km away from the Keweenaw Peninsula to Michipicoten Island. Normally, the mid-lake area of the eastern lake basin first develops an extensive ice cover during the latter half of February. Ice formation continues through the end of March and at the same time ice cover can also begin to be lost in parts of the lake attributed to wind action and vertical mixing. The ice cover in mid-

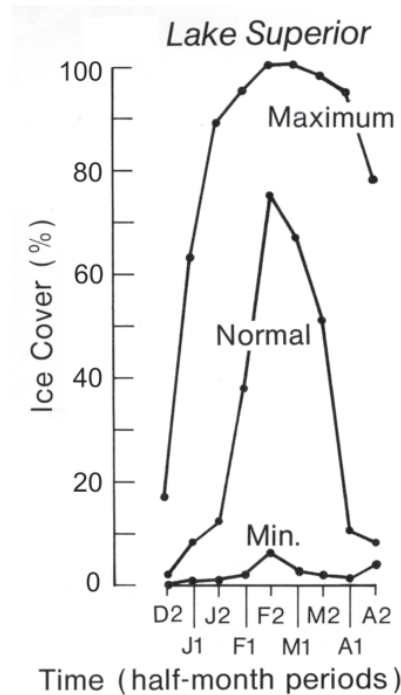


Fig. 7. Maximum, minimum and normal ice concentration distribution pattern for Lake Superior for nine half-month periods starting December 16-31 (D2) and ending April 16-30 (A2). (based on Assel et al., 1983).

lake areas is dissipated during and prior to the first half of April and continues to melt during the month of April. Minimum ice cover can extend to nearly 10 % of the lake surface area in mid-February with some persistent areas of cover in the near-shore areas. In contrast, Lake Superior can achieve 100% ice cover in the mid-February period compared to only 75% coverage in a normal winter. In severe winters, there is often persistent ice cover remaining on the lake mostly in the nearshore past the second-half of April.

Thermal Bar

During early spring, the temperature in a large temperate lake is more or less constant and below 4°C (temperature of maximum water density). As the spring heating proceeds, the water is heated convectively, leading to a faster increase of temperature in shallow

areas compared to deeper waters. When the temperature reaches 4°C, a stable stratification develops at the shore. Between this region and deep part of the lake that still experiences convection and a temperature below 4°C, there is a zone of sinking water in the vicinity of a temperature corresponding to maximum density. This zone is referred to as thermal bar. This process has been investigated (e.g. Tikhomirov, 1963) and also described for the Great Lakes (Rodgers, 1965; Ullman et al., 1998; Rao et al., 2004; Auer and Gratzke, 2004). Heat and mass exchange is limited due to the presence of the thermal bar and the shoreward water trapped by the thermal bar, can become enriched with materials from tributary discharge and spring runoff. In addition, biological materials and nutrients can also be trapped in the nearshore. Compared to the other Great Lakes, the thermal bar in Lake Superior forms latest and the temperature gradients between nearshore and deep offshore waters are generally maintained throughout the summer (Ullman et al., 1998).

Auer and Gratzke (2004) examined the thermal bar formation in Lake Superior and quantified its role in mediating cross-margin transport. It was found that the spring runoff event has the potential to increase levels of terrestrially-derived substances, including biologically important material in the nearshore waters by as much as a factor of 30 which could be important in mediating wood-web dynamics. Comparison between the timing of the spring runoff event and thermal bar formation determined from satellite images, showed significant interannual differences in both the magnitude of the runoff event and the timing of the thermal bar formation. The thermal bar generally forms between early April and late May. On average, thermal bar formation lags the spring runoff event by 3 weeks. Consequently, such variability can result in considerable year-to-year differences in the response of bacterio-plankton and phytoplankton communities.

Spring- Summer-early Fall thermal structure

Most of the investigations of the thermal structure of the lake have been conducted during spring and summer, (e.g. Smith and Ragotskie, 1971; Smith, 1972; Hubbard and Spain, 1973; Ragotskie and Niebauer, 1975; Spain et al., 1975; Bennett, 1978a).

Budd et al. (1999) used AVHRR satellite imagery which was

capable of providing detailed data on the surface “skin” temperature of Lake Superior. Comparison of AVHRR derived surface temperature and that from NDBC buoy bulk temperatures showed excellent correspondence ($r = 0.971$) with a slight negative bias $< 1^{\circ}\text{C}$ at low temperatures. Pixel resolution from the AVHRR are approximately $1.1 \text{ km}^2 \times 1.1 \text{ km}^2$ which was adequate for resolving broad-scale patterns of surface skin temperatures, including large-scale eddies off the Keweenaw Peninsula. Based on AVHRR data it was possible to describe the progression of the seasonal temperature cycle. An example of satellite imagery of the surface temperature of Lake Superior is provided in Fig. 8 for four periods from spring to fall, 1993.

During early March to April, the whole surface of the lake is still below 4°C and RADARSAT ScanSAR images indicate parts of the lake, especially near the lake periphery, remain covered with floating ice (e.g. see Assel, 1986). By early April, the lake water column is essentially isothermal with a temperature of approximately 2°C . During the initial warming stage, the shallower shore areas heat faster and receive warmer water from tributaries. In later June, a horizontal temperature gradient is typically evident with western shore locations ranging 5 to 11°C compared to the mid-lake which may be only 2 to 3°C . The May to early July period generally show large temperature differences between the rapidly warming nearshore and the cooler offshore areas. Disappearance of the thermal bar in July, signals the thermal stratification period of the lake and the appearance of upwelling and downwelling along the central and eastern shorelines characteristic of Lake Superior. The heat content of the lake continues to increase in response to the surface heat gains (Schertzer, 1978; Bennett, 1978a), however, the highest temperatures remain in the nearshore compared to the offshore regime. Major increases in central offshore temperatures are typical in mid- to late-August as the lake is strongly thermally stratified to mid-September.

Late-Fall and winter thermal structure

Significant cooling of the lake occurs during mid-September to November, also reflected in the increased heat losses from the lake to the atmosphere (Schertzer, 1978). From mid-December onwards, the lake temperature decreases rapidly to below 4°C . As cooling

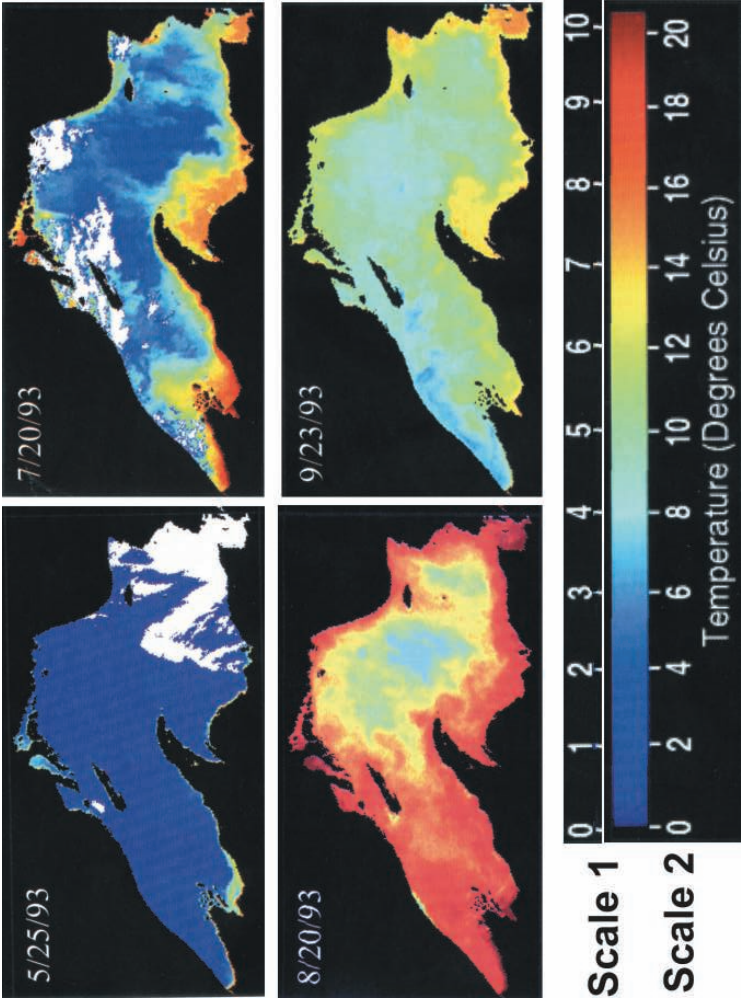


Fig. 8. Examples of the spatial distribution of surface temperature based on AVHRR for Lake Superior (a) May 25, 1993, (b) July 20, 1993, (c) August 20, 1993 and (d) September 23, 1993. (selected AVHRR images based on Budd et al., 1999). Clouds are masked as white colour.

progresses, the lake develops an ice cover whose extent is dependent on the severity of the winter (Assel et al., 1983).

Assel (1986) investigated fall and winter thermal structure of Lake Superior using temperature surveys along the upbound (westward) and downbound (eastward) shipping lanes. Instrumentation aboard ore carriers, included portable bathythermograph (BT) and expendable BT probes used during surveys of 2 to 4 day duration; 21 surveys during winters 1973 to 1976 and 25 surveys during the falls of 1976 to 1979. Examples of the variation of the mixed layer depth at 4 locations in the lake are shown in Fig. 9. The main characteristics found were an approximately exponential increase in fall mixed layer depth through early to mid-November.

Maximum values of averaged mixed layer and upper 25-m layer temperatures occurred between the end of August and mid-September,

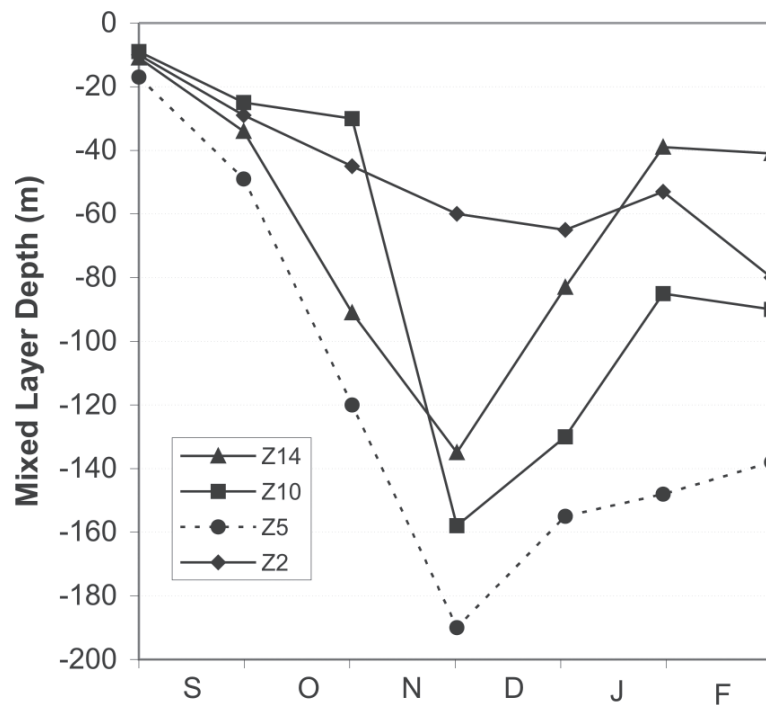


Fig. 9. Monthly mixed layer depths (ML, m) determined over 4 locations in Lake Superior (see Fig. 1). (redrawn from Assel, 1986).

while the maximum value of average water column temperature occurred in late September. Isothermal conditions are prevalent between mid-November and mid-December. Fall overturn through the temperature of maximum density (4°C) occurs in December and temperature induced reverse stratification in winter occurs in December of January depending on the winds. The average winter (January to March) mixed layer depth is between 60m and 100m depth. The minimum value of average water column temperature occurs in late March. Deeper areas of the lake with longer wind fetch in both the nearshore and mid-lake thermal regimes, have the deepest mixed layers during winter showing a correspondence with bathymetry, wind fetch and circulation patterns. Areas having a combination of greater depth and larger wind fetch, usually mid-lake areas, also tended to have higher column temperatures and ice cover of short duration in winter and lower column temperatures in summer relative to adjacent areas.

Air-water Heat Exchange and Heat Content

The surface temperature as outlined above is part of a “feed-back” loop between the atmosphere and the lake surface due to radiative and turbulent exchanges of heat. Accurate estimates of magnitude of and seasonal variation of the surface energy exchange is required for understanding many physical processes of large lakes. Components of the total heat flux forms part of the surface boundary condition for thermal and hydrodynamic models. The solar radiation itself is an important component of biological significance in lakes.

Heat fluxes

An initial estimate of heat gains and losses at the air-water interface was done by Schertzer (1978) using shoreline meteorology, hydrological inflows and outflows, surface water temperature and a series of 19 temperature surveys. The total heat flux was determined as the net radiation, minus the sensible and latent heat fluxes (evaporative) plus the heat advection due to hydrological balances. Errors in the 1973 flux computations were minimized based on the extensive shoreline stations available and the relatively high number of lake surveys

during the experiment. Even so, data were lacking for treatment of such terms as fog which can have a notable effect during early spring (e.g. Derecki, 1976). A measure of uncertainty in the 1973 computations was derived by examining the difference in the computed total heat flux determined from the energy budget (Schertzer, 1978) with substitution of the latent heat flux component determined from the mass transfer approach (Schertzer, 1978) and the water budget (Bennett, 1978b). For ice-free periods, the largest difference occurred using the mass transfer evaporation estimates ($+24.2 \text{ W m}^{-2}$) compared to ($+10.7 \text{ W m}^{-2}$) using the water budget evaporation.

Recent evaluation of the surface energy fluxes on the Great Lakes, including Lake Superior was conducted by Lofgren and Zhu (1999; 2000) which included surface temperature from the Great Lakes Coastwatch Program which uses AVHRR taken from the NOAA Polar Orbiting Satellites. Lofgren and Zhu (2000) followed methodologies described by Croley (1989). Ice cover has not been included in the computation of the surface fluxes, and consequently, results for winter months with ice cover may be inaccurate.

The general characteristics of the surface energy budget for Lake Superior is shown in Fig. 10 as monthly spatial averaged means (Lofgren and Zhu, 2000). Net radiation was positive for most of the year reaching a maximum in June. Sensible heat was generally lost to the atmosphere during the fall and winter but added to lake heating during May through August with the maximum contribution occurring in May. The energy utilized for evaporation was low in May through July and higher in the fall months owing to higher wind speeds and larger air to water vapour pressure differences. The heat fluxes derived by Lofgren and Zhu (2000) for periods 1992-95 are in general agreement with the seasonal patterns generated by Schertzer (1978) for 1973. Differences are attributable to use of different datasets, various assumptions and use of more spatially representative surface temperatures in the recent computations.

Heat content

Bennett (1978a) derived the first approximation of the longterm heat content for Lake Superior based on 19 temperature surveys conducted from 1964 to 1973 with vertical temperature profiles determined from electronic bathythermograph. The number of stations varied

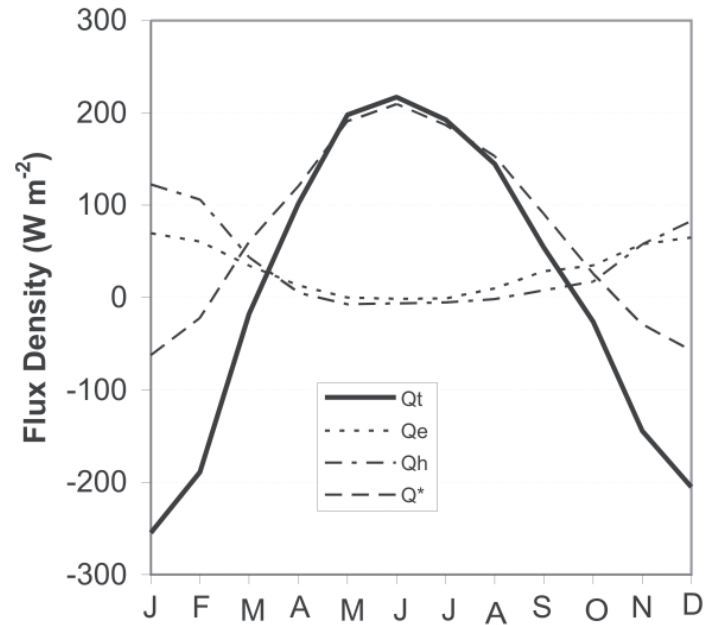


Fig. 10. Characteristics of the surface energy budget of Lake Superior (redrawn from Lofgren and Zhu, 1999; 2000). Q_t is total heat flux change, Q_e is latent heat flux, Q_h is sensible heat flux and Q^* is the net radiation.

from 80 to 200 stations per survey. Most surveys extended up to 2 weeks and were more intense during the thermally stratified period.

General characteristics of the heat storage of Lake Superior are given in Fig. 11 based on observations during the ice-free period and with subjective interpolation over the winter months. The heating cycle in Lake Superior extends from March to late-September and the lake experiences heat loss during the remainder of the year. The long-term range between minimum and maximum heat storage is estimated at 2.89×10^{20} J (Schertzer, 1978).

Water Movements and Modelling

The first map of mean summer currents in Lake Superior was presented over 100 years ago by Harrington (1894) who deduced flow patterns based on drift bottles that were recovered along the

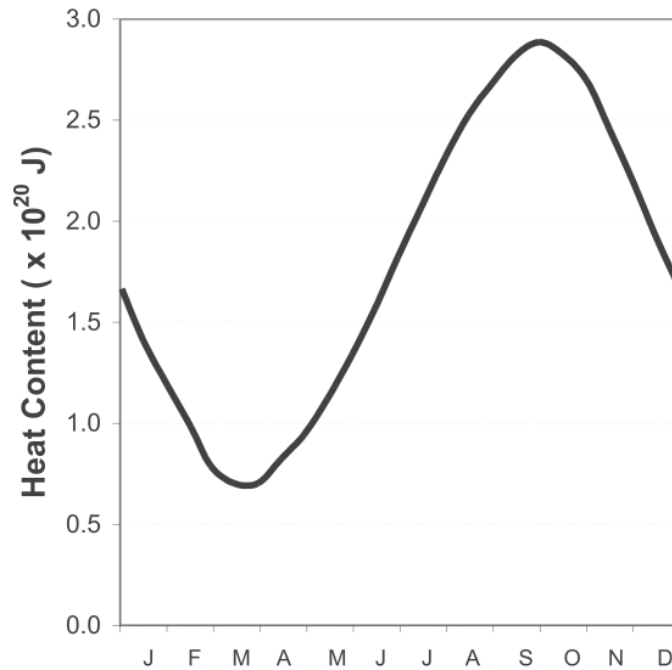


Fig. 11. Characteristics of the lakewide heat content of Lake Superior (recomputed from Schertzer, 1978)

lake coasts during the summers of 1892 and 1893 (Fig. 12). His interpretation of the tracks of drift bottles in Lake Superior indicated counterclockwise currents in the main lake, including the Keweenaw Current. Lake Superior is a typical closed basin in which currents and stratification are controlled by the external surface wind forcing and surface heat flux (Chen et al., 2004a). Both observations and numerical models are used to understand characteristics of the lakewide circulation pattern, and dynamics of nearshore and mid-lake exchange processes.

Generalization of water circulation

Knowledge of the mean circulation of lakes is important for ecological and management issues. The patterns provide an indication of transport pathways of nutrients and contaminants on long-time scales. Currents have been measured by several investigators; however, it is often

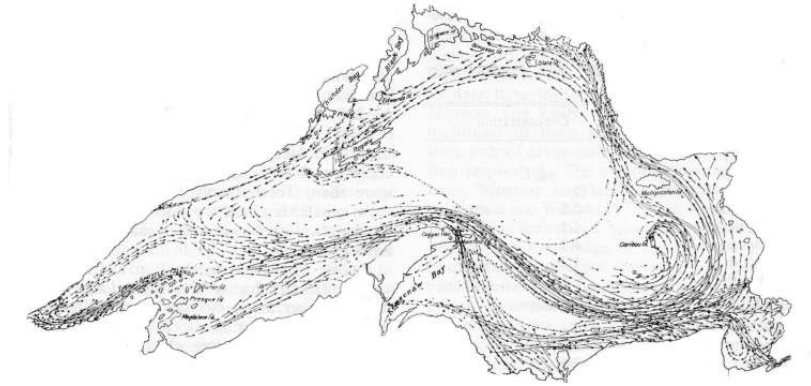


Fig. 12. Surface currents in Lake Superior (from Harrington, 1895).

very difficult to derive a mean circulation pattern from limited data. Most observations have been taken during summer ice-free periods and these are generally associated with research studies in a specific region in the lake. In the last three decades there has been progress in measurement of currents. Beletsky et al. (1999a, 1999b) collated all available information from historical and existing circulation maps in a consistent manner in an effort to describe the general circulation characteristics during summer, winter and annual periods in the Great Lakes, including Lake Superior.

Current measurements are measured in one of two approaches, either Lagrangian (based on tracking moving particles), or Eulerian (based on time-series at a fixed point). Examples of Lagrangian experiments conducted on Lake Superior are those of Harrington (1894) and Ruschmeyer et al. (1958), and Eulerian methods were used by Sloss and Saylor (1976) and others. A third method has been the development and application of hydrodynamic models on the Great Lakes (Lam, 1978; Chen et al., 2004).

Figure 13a shows the mean summer and winter circulation pattern in Lake Superior (Beletsky et al., 1999) based on data from both historical and present day observations. In general, the more recent observations of currents in Lake Superior have similar characteristics to that determined by Harrington (1894). Observations have shown that the summer circulation is mostly cyclonic and the Keweenaw

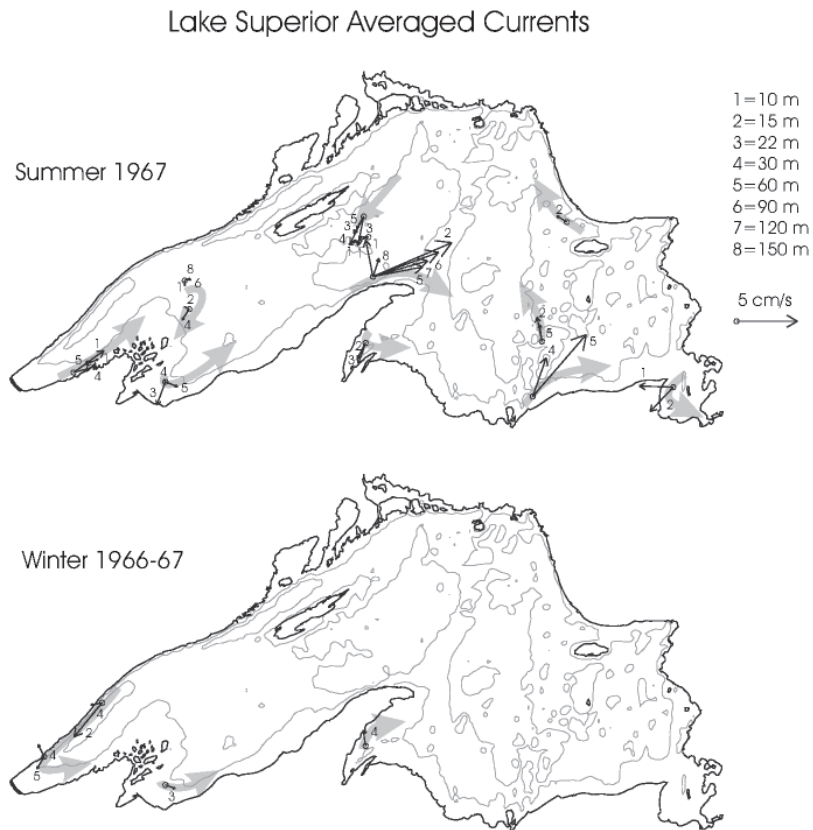


Fig. 13. Summer and winter circulation in Lake Superior (based on Beletsky et al., 1999a). Numerical values (1 to 8) represents the depth category of observation from the surface.

current is a persistent feature of summer circulation in Lake Superior. It is a warm coastal current that flows northeastward along the northern shore of the Keweenaw Peninsula (see Stn. 8, Fig 13a). Interannual variability in the current patterns has not been systematically studied. In general, there is insufficient observational data to discern the large-scale winter circulation in Lake Superior. Winter data are only available for the western part of the lake with some general circulation arrows provided on Fig. 13b.

Numerical modelling of lakewide water circulation

Description of the lakewide circulation in large deep lakes such as Lake Superior are based on steady-state dynamics (e.g. Rao and Murty, 1970), numerical models (e.g. Hoopes et al., 1973; Lam, 1978; Diehl et al., 1977; Schwab and Bedford, 1994; Chen et al., 2001) or based on broad-based data sets (e.g. Beletsky et al., 1999a, 1999b).

In one of the earliest applications of hydrodynamical model to Lake Superior, Lam (1978) simulated the summer water circulation and transports in Lake Superior based on Simons (1973) model. The horizontal grid size used in the model was 10 km x 10 km with four vertical layers. A chloride transport model (Lam and Simons, 1976) was used in conjunction with the hydrodynamic outputs to simulate the chloride movements in the lake. The hydrodynamic model showed reasonable agreement with observed currents over the June-September 1973 thermally stratified period. Figure 14 shows mean computed currents (thin vectors) for the epilimnion case. Comparison with observations (thick vectors) shows reasonable agreement with observed currents. The general counter-clockwise circulation and some features of the Keweenaw current were shown in the simulations. The transport model adequately simulated characteristics of the transport of the conservative substance of chloride despite small areal variations (Lam, 1978).

Recently, Chen et al. (2001) and Zhu et al. (2001) developed a high-resolution non-orthogonal grid 3D model for Lake Superior. The model was able to simulate the important dynamics in the lake including the large scale counter-clockwise circulation, inertial oscillations and dynamics of Keweenaw coastal current. Further, model simulations clearly indicated spring warming and formation of the thermal bar and importance of horizontal temperature distribution on the intensification of coastal current.

Coastal upwelling / downwelling cycles

Coastal upwelling and downwelling are frequently observed in the Great Lakes (e.g. Mortimer, 1971; Simons and Schertzer, 1987). The upwelling and downwelling cycles contribute to water mass exchange between the nearshore and offshore zones and are fundamental to

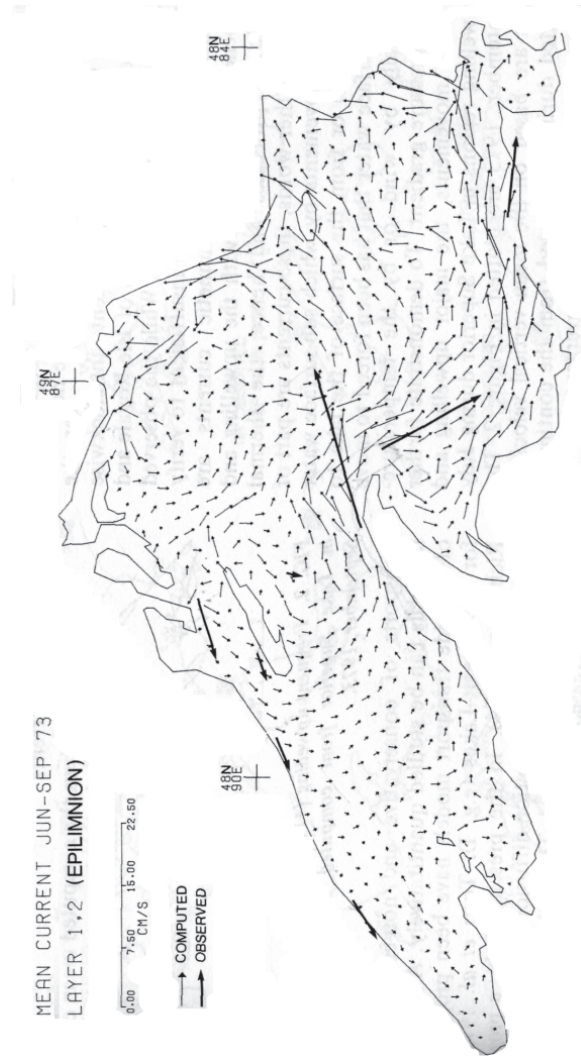


Fig. 14. An example of the mean computed currents (thin vectors) for the epilimnion case in June-September, 1973 (based on Lam, 1978). Observations are shown as thick vectors.

understanding of lake dynamics. The coastal current off the Keweenaw Peninsula has been investigated by several researchers (e.g. Ragotskie and Bratnick, 1965; Smith and Ragotskie, 1970; Green and Yeske, 1974; Yeske and Green, 1975 and Niebauer et al, 1977). Niebauer (1976) confirmed the existence of two upwelling/downwelling regimes. Based on 1972 experiments, minor upwelling events and major downwelling events may be observed in July with major upwelling events and minor downwelling events in later July and in August in the Keweenaw. The cycle of upwelling and downwelling events may have periods of about 4-5 days.

Keweenaw Current: observations and modelling

Niebauer et al. (1977) found that longshore wind components associated with passing low-pressure systems cause Ekman transport and coastal downwelling along the Keweenaw Peninsula. The Keweenaw current can reach speeds of 75 cm s^{-1} . Passing high-pressure systems north of Lake Superior can cause offshore Ekman transport and coastal upwelling. In such cases, surface water offshore are found associated with average coastal upwelling velocities of $15 \times 10^{-3} \text{ cm s}^{-1}$ with maximum values of $60 \times 10^{-3} \text{ cm s}^{-1}$ in the nearshore.

Chen et al. (2004a) investigated the response of Lake Superior to mesoscale wind forcing over the lake grid based on wind data recorded at 4 lake buoys and from land-based meteorological stations around the lake (Fig. 1). Over-land winds were modified to over-water winds (Schwab, 1978; Hsu, 1986). Additional wind information was available through satellite scatterometer observations from QuikSCAT. Winds measured by QuikSCAT every 12 hours in Lake Superior represent a snapshot of the wind field generally 35 km away from the coast. A comparison between QuikSCAT and buoy winds showed excellent correspondence with correlation coefficients of 0.95 for speed and 0.91 for direction (Chen et al., 2004a).

Comparisons between observed and model-predicted temperatures and currents clearly showed that application of the QuikSCAT wind field provided better simulation of the low frequency variation of water temperature and currents in Lake Superior (Chen et al., 2004b) during upwelling and downwelling events. The investigation indicated that the spatial resolution of wind measurements influences the accuracy of model-based predictions of currents in this lake –

especially in the nearshore (Fig. 15). Model simulations showed an underestimation of the magnitude of the eastward velocity of the Keweenaw current especially during downwelling events. Both limited meteorological buoy data and satellite-based QuikSCAT observations are not of sufficient resolution required for high-accuracy simulations on this lake especially for coastal processes. It is suggested that improvements in the simulation of the Keweenaw current requires improvement in the coastal wind field possibly through coastal radar measurement systems or with mesoscale meteorological models.

Scales and structures: prevalence of eddies

Ralph (2002) examined scales and structures of open-lake circulation in Lake Superior. Observations of currents and near surface temperatures were conducted over more than 5000 km of survey tracks. Vertical profiles of horizontal currents were measured with a 150 kHz vessel mounted acoustic Doppler current profiler. Near surface temperatures were measured every 0.125 seconds to an accuracy of 0.003 K with a hull mounted platinum resistance thermometer. The velocity data showed a complex mid-lake circulation filled with eddies. From a dynamical perspective, Ralph (2002)

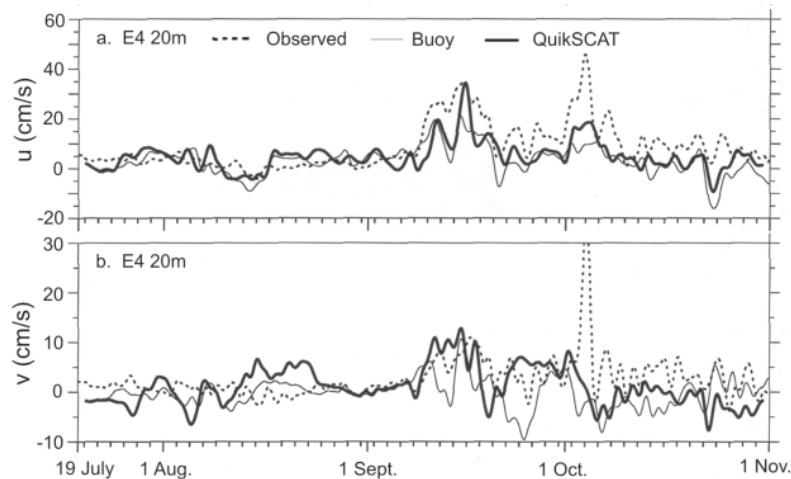


Fig. 15. Comparison between observed and model predicted Keweenaw currents. (based on Chen et al., 2004a)

surmised that the vertical structure of the eddies suggests that they were probably barotropic while their horizontal scales and current temperature fluctuations suggest that the eddies were generated by baroclinic processes. Two-dimensional sea surface temperature (SST) data from AVHRR provided insight into the generation and distribution of the eddies. Modelling studies (e.g. Chen et al., 2001) suggest that eddies are preferentially found near baroclinically-unstable coastal jets and due to topographic features. Throughout the AVHRR images, Ralph (2002) has found large temperature anomalies whose persistence suggests that eddies are continuously being generated. Examination of the spatial spectra of the temperature field suggests that the temperature anomaly field is a signature of mesoscale dynamics and stirring motion of the eddies. The importance of understanding such eddies is that they may provide a mechanism for motions that are responsible for moving lake sediments as well as nutrients and contaminants that adhere to the sediments. From a modeling point of view, understanding eddy circulation may improve hydrodynamic model ability to predict 3-D velocity fields.

Keweenaw Waterway water exchange

Embayments and waterways can be conduits for the movement of water and materials to the lake. In Lake Superior, the Keweenaw Waterway is a 38 km navigational passage that cuts across the Keweenaw Peninsula (Fig. 1). Churchill et al. (2004) described the frequency and hydrographic character of Lake Superior water intrusions into the waterway, and investigated mechanisms by which the surface wind affects the movement of water into and out of the waterway. The analysis indicated that movement of water into and out of the waterway is strongly influenced by surface winds but the wind effect diminishes with depth. Pressure gradients arising from wind-driven changes in coastal water levels is another important mechanism influencing the flow. Upwelling favourable winds can have a tendency to depress coastal lake level, and create a pressure gradient driving flow out of the waterway into the shore regions together with materials. The exchange can also significantly enhance biodiversity within both the waterway and the coastal zone of Lake Superior (Churchill et al., 2004).

Material dispersal by ice rafting – an additional transport mechanism

In addition to direct transport mechanisms such as the Keweenaw Current, ground studies and aerial observations have shown that nearshore ice may incorporate large sediment loads which can be subsequently transported to midlake areas via ice rafting (Budd et al., 1999). The bulk of the mass of sediments incorporated into nearshore ice complexes occurs through several mechanisms including the action of waves that can entrain sediments into the water column or directly onto the ice, through ice kneels that scour and entrain sediments, or from sediment-laden anchor ice detached from the lake bottom. Shallow bathymetry, abundant source materials and strong onshore winds and waves are the main environmental factors that facilitate the sediment entrainment-transport process.

RADARSAT (radar satellite) and ScanSAT (Synthetic Aperture Radar) imaging were used to analyze material dispersal through ice rafting (Budd et al., 1999). Nearshore ice incorporates large amounts of coastal sands and deeper water sediments as indicated above. During spring break-up of ice in the coastal regions, large drifting ice packs, which contain large sediment loads, are pushed by prevailing westerly winds and currents. This is often particularly evident around the tip of the Keweenaw Peninsula.

Benthic Nepheloid Layer role in sediment transport

A layer of water immediately above the sediments with increased concentration of suspended sediments is known as the benthic nepheloid layer (BNL). As part of the KITES project, Urban et al. (2004a) analyzed the extent and the composition of the BNL in Lake Superior off the northwest coast of the Keweenaw Peninsula using transmissometer profiles and samples analyzed for major elements. It was found that the BNL develops each summer when the lake is stably stratified and disappears in early fall as the thermocline erodes. Particles within the BNL are enriched in biogenic elements relative to material in surface sediments and are enriched in phosphorus relative to epilimnion seston. Along a 20 km stretch of the Keweenaw Peninsula, the inventory of solids suspended particles (e.g. 100,000 metric tons) appears quickly in summer and disappears quickly in the fall.

The BNL has been thought to be important for cycling of carbon, nutrients and contaminants. Numerous theories exist on the dynamics of the BNL and its role in sediment transport based on observation in other lakes such as sediment resuspension (Eadie et al., 1984) and upwelling (e.g. Bell and Eadie, 1983). In Lake Superior, Urban et al. (2004b) found that the evidence for a large role of the BNL in sediment transport was contradictory. A simple mixing model using nearshore sediments and BNL particles (Urban et al., 2004) suggested that the BNL particles could contribute up to 50% of the carbon measured in the sediment traps. In contradiction, settling rates determined from sediment traps suggest that it would take 1 – 12 months to remove August total solids from the BNL. Alternately, sediment radioisotope analysis from KITES sediment traps indicate particle settling at rates greater than 2.5 m d^{-1} (Chai and Urban, 2004). The current analysis indicates that BNL particles play a role in settling fluxes, but they indicate that transport of larger particles is occurring concurrently, and that the transport of the larger particles is masking movement of particle sizes located within the BNL. More research is required to analyze this problem to determine the sediment transport mechanism in the BNL and importance in Lake Superior.

Climate change considerations

There have been theoretical and modelling studies that have considered potential responses of large lakes to climate variability and change, however, there is a paucity of direct observations from which to understand the actual responses. Modelling studies on large lakes have explored the potential lake temperature changes in a number of approaches such as climate analogs (e.g. Schertzer and Sawchuk, 1990), through use of both steady-state and transient climate scenarios (e.g. Croley, 1990; McCormick, 1990; Boyce et al., 1993) and through novel approaches such as climate transition (e.g. Croley et al., 1996). Based on Lam and Schertzer (1999), there are uncertainties in the current GCM climate scenarios and a lack of coupling between lake and atmospheric models which preclude other than general descriptions of potential climate change effects on such processes as the lake circulation.

Potential impacts on selected meteorological variables and hydrological components

Austin and Colman (2007) investigated trends in summer water temperatures using three surface buoys from Coastal Marine Automated Network (CMAN) stations around the lake periphery and on islands operated by the National Data Buoy Centre (NDBC, National Oceanic and Atmospheric Administration (NOAA) (Fig. 1). Hourly observations have been available over the months April through November since the 1979 to 1981 period. Ice cover data were derived from NOAA reports (Assel et al., 2003; Assel, 2005) and represented spatially and temporally averaged ice cover percentage over the entire lake between December 1 of the previous year and 31 May. Their analysis showed a trend $(5.3 \pm 2.1) \times 10^{-2} \text{ }^{\circ}\text{C yr}^{-1}$ in annually averaged air temperature from 1979 to 2005. Ice cover in Lake Superior has been declining over that period at a rate of $(-0.42 \pm 0.20) \% \text{ yr}^{-1}$. Near surface water temperature, based on the buoy data, has been increasing at about $(11 \pm 6) \times 10^{-2} \text{ }^{\circ}\text{C yr}^{-1}$. In addition, all three buoys showed a general trend towards higher average wind speeds in the order of $0.05 \text{ m s}^{-1} \text{ yr}^{-1}$. Weaker trends were observed at the CMAN sites in Lake Superior likely due to coastal location or measurement height.

Lofgren et al. (2002) evaluated the potential impacts on Great Lakes water resources based on scenarios from the Canadian Centre for Climate Modeling and Analysis (CGCM1) and the United Kingdom Meteorological Office Hadley Centre (HadCM2). Climate scenarios derived from these two models can produce different future states for such variables as air temperature and precipitation (Lofgren et al., 2002). With respect to hydrological changes, the CGCM1 projects significantly reduced runoff and higher lake evaporation than HadCM2. The base water level in Lake Superior (183.42 m, above sea level) is projected to decrease by -0.42 m by 2090 using the CGCM1 drier-warmer climate scenario compared to an increase ($+0.11 \text{ m}$) using HadCM2 wetter-warmer climate.

Potential Impacts on thermal structure and ice cover

Over the last 27 years, the date at which summer thermal stratification begins has been occurring increasingly earlier in the year at a rate

roughly a half day per year (Austin and Coleman, 2007). At the same time, the fractional ice coverage in Lake Superior has been decreasing over the last several decades (Assel et al., 2003). The investigation concluded that there is a clear relationship between the whole-lake fractional ice cover in a given season and the subsequent onset of summer stratification and hence the mean summer temperature. The relationship is complex. The increase in summer water temperatures is hypothesized to be a combination of reduced ice cover leading to a progressively earlier start of the stratified season and higher summer air temperatures with which the lake is equilibrating.

Base ice duration days for the west basin, east basin, and for Whitefish Bay are 111 days, 108 days and 115 days, respectively, however, based on the CGCM1 model, the ice duration is expected to decrease significantly to 32, 28, and 40 days, (respectively) and to 66, 68 and 77 days (respectively) through the HadCM2 model. The percent of ice-free days is significantly higher using the CGCM1 compared to the HADCM2 scenario. The earlier ice break-up and longer stratified period will have an effect on the air-water heat exchange and thermal response, however, there are currently no observations to quantify such changes. It is postulated that longer stratified seasons with warmer summer water temperatures are likely to have a significant impact on the ecology of Lake Superior at all trophic levels (e.g. Brooks and Zastrow, 2002; Hill and Magnuson, 1990; Jones et al, 2006; Lehman, 2002; and Magnuson et al., 1997).

Summary

This paper provides an overview of historical and current knowledge regarding important characteristics of the surface meteorology, hydrology, thermal structure, energy fluxes and heat storage, water movements and hydrodynamic modelling in Lake Superior. Relevant descriptions of measurements and model simulations are also presented. The following provides a brief synopsis of the state of the research, and the effect of improvements in measurements and techniques on the current knowledge-base on selected physical components of the lake.

Advances and Required Improvements in Measurements

Historical analyses of the environmental climatology were largely based on shoreline meteorological observations. Current investigations have the benefit of in situ lake observations from NOAA and Canadian buoys. Combined observations from adjusted shore-based observations, limited lake buoys and use of satellite observations (e.g. AVHRR and QuikSCAT) have improved interpretations and analyses of dynamical processes. There are surveillance programs operational on the lake with meteorological buoys in ice-free periods, however, the spatial resolution of data is limiting for developing climatological time-series databases with high spatial resolution. Model simulations of nearshore dynamics such as upwelling and downwelling events and currents near the Keweenaw Peninsula have shown improved accuracy. Further improvements may be achieved through improvement in the coastal wind field possibly through coastal radar measurement systems or with mesoscale meteorological models. Modern observations and forecast systems for the Great Lakes over the past decades have allowed for greater advanced notice of storms with potentially extreme winds and waves which has increased the safety of ship transport on Lake Superior. Similarly, advancements in procedures to derive daily representations of water surface temperature using satellite as well as ice cover information as shown using GLSEA, will provide increased spatial information applicable to lakes research than traditional methods alone.

Advances in Hydrological Components and Evaporation Technique

An understanding of the magnitude and variation of lake hydrological components and response to climate is essential for lake management and for realistic regulation of the lake water levels and outflow. Early research (e.g. Bennett, 1978) established the baseline characteristics for water budget components. Over-lake evaporation on a lake the immense size of Lake Superior is a challenge to measure. Initial estimates of lake evaporation using constant transfer coefficients similar to the Lake Hefner studies resulted in very high levels of condensation (negative evaporation) during the months with very stable atmospheric conditions. Energy budget, water budget, and improved mass transfer formulations using variable transfer coefficients

and recently including corrections for static stability and wind shear have shown a convergence of monthly mean evaporation values. Application of advanced techniques such as fast response eddy covariance methods will provide improved estimates of the lakewide evaporation as well as an understanding of the dynamic responses of the evaporative process to atmospheric forcing.

Improved Knowledge of Surface Heat Exchange and Heat Content

Knowledge of the surface heat exchange is an important boundary condition input to thermal and hydrodynamic models. Schertzer (1978) provided baseline values of the radiative and turbulent exchange components largely based on shoreline meteorological data. Bennett (1978) used water temperature profiles based on long-term lake surveys to approximate the lake heat storage for Lake Superior. Intensive observations conducted in 2005 from shoreline meteorological sites, lake meteorological buoys and in situ lake temperature profiles, offer a possibility to re-evaluate the magnitude and seasonal variation of both the heat flux components and lake heat storage.

Requirement for Improved Long-term Current Observation

As indicated previously, advancements in simulation of the lakewide circulation and dynamics of the water movements in the nearshore, requires improved resolution of boundary conditions such as the wind field. Current investigations have focused on generalization of the summer, winter and annual circulation based on historical data (e.g. Beletsky, 1999a: 1999b). Observations have shown that the summer circulation is mostly cyclonic and the Keweenaw current is a persistent feature of summer circulation in Lake Superior. In general, there is insufficient observational data to discern the large-scale winter circulation in Lake Superior.

Advances in Keweenaw Circulation Modelling and Role of Eddies

Recent research has seen the application of hydrodynamic models in the lakewide case and nearshore to simulate currents in the complex zone of the Keweenaw Peninsula. Mid-lake eddy fields have been observed in Lake Superior throughout the water column and throughout the lake during all seasons. Findings of Ralph (2002) raise several

questions regarding the mechanisms of formation, propagation and dissipation as well as the complex 3-D structure. Eddies are postulated to have implications on how nutrients, sediments and contaminants are transported within large lakes and as well as the patchiness of phytoplankton and zooplankton.

Important Role for Remote Sensing

AVHRR imagery, supplemented by higher resolution SPOT-HRV, LANDSAT TM, or RADARSAT scenes, can provide high spatial coverage and real-time detail necessary to monitor dynamic changes associated with many meso-scale lake phenomena (Budd et al., 1999). Application of AVHRR and SeaWiFS instrumentation may provide further remotely sensed data to investigate dynamic temperature and circulation processes.

Requirement for Research on Climate Change and Development of Improved Scenarios

Modelling techniques for the simulation of climate change effects on lake hydrodynamics are available, however, data input to such models do not always exist (Lam and Schertzer, 1999). In the case of Lake Superior, Austin and Coleman (2007) indicated that while there have been theoretical and modelling studies on the lake, there is a paucity of direct observations from which to understand the actual responses. Annually averaged air temperatures from 1979 to 2005 have shown a positive trend, influencing a decreasing trend in ice extent in Lake Superior. Wind speeds from over-lake meteorological buoys show a tendency to higher average winds. These changes have an effect on the lake thermal response. July to September water temperatures have increased ~ 2.5 °C over the period 1979 – 2006 which is significantly in excess of regional atmospheric warming and has had the effect of lengthened the ice-free season resulting in earlier temperature stratification. Changes of this magnitude may have implications on other physical responses of the lake including aquatic ecosystem components.

Climate scenarios have been applied to Lake Superior largely focused on hydrological responses and ice extent. In general, these simulations have shown a potential toward reduced runoff, higher

lake evaporation and a decrease in lake water level and decrease in ice extent. The magnitude of change can be quite different using scenarios from different GCM models. Simulation of potential climate impacts on water movements in Lake Superior requires improved climate scenarios and advances which may include lake-atmospheric coupling.

Our understanding of physical limnological processes has increased greatly over the last three decades. Lake-wide models have been developed and the influence of meteorological and hydrological events are much more defined. However, further development in models at higher resolution in the nearshore will require an increase in spatial distribution of meteorological observations. Furthermore, these models have to be validated with hydro and thermodynamic variables in the lake. Analysis of trends in lake temperatures, water levels and ice cover strongly suggest that we are already experiencing warming in Lake Superior. Although future climate scenarios are uncertain, potential effects of climate change needs to be assessed using the deterministic lake models.

Acknowledgements

Much of the research included in this overview has been described based on reliance on the excellent research papers of colleagues within the Great Lakes community. We are indebted to those colleagues who have participated in both large-scale and smaller-scale investigations through historical studies which provided baseline characteristics and in more contemporary research with advanced instrumentation from which we drew heavily for our overview. We are also grateful to the Editor, Dr. Munawar, who encouraged us to complete this manuscript. We acknowledge the constructive comments from two peer reviewers.

References

- Allan, R.J., Dickman, M., Gray, C.B., Cromie, V., 1994. The Book of Canadian Lakes. The Canadian Association on Water Quality, Monograph Series, No. 3.
- Assel, R.A., 1986. Fall and winter thermal structure of Lake Superior. J. Great Lakes Res. 12(4), 251-262.

- Assel, R.A., 2005. Classification of annual Great Lakes ice cycles: Winters of 1973-2002. *J. Clim.* 18, 4895-4905.
- Assel, R.A., Quinn, F.H., Leshkevich, G.A., Bolsenga, S.J., 1983. *NOAA Great Lakes Atlas*. U.S. Dept. of Commerce, National Oceanic and Atmospheric Administration, Great Lakes Environmental Research Laboratory, Ann Arbor, MI.
- Assel, R.A., Cronk, K. and Norton, D., 2003. Recent trends in Laurentian Great Lakes ice cover. *Clim. Change* 57, 185-204.
- Aubert, E.J. and Richards, T.L. (Eds.), 1981. IFYGL-The International Field Year for the Great Lakes, NOAA-GLERL, U.S. Dept. of Commerce, Ann Arbor, MI.
- Auer, M.T., Gatzke, T.L., 2004. The spring runoff event, thermal bar formation and cross margin transport in Lake Superior. *J. Great Lakes Res.* 30(Supplement 1), 64-81.
- Austin, J.A., Colman, S.M., 2007. *Geophys. Res. Letters*, 34: doi:10.1029/2006GL029021.
- Beletsky, D., Saylor, J.H., Schwab, D.J., 1999a. Mean circulation in the Great Lakes. *J. Great Lakes Res.* 25(1), 78-93.
- Beletsky, D., Lee, K.K., Schwab, D.J., 1999b, *Large-scale Circulation*. In: D.C.L. Lam and W.M. Schertzer (Eds.), *Potential Climate Change Effects on Great Lakes Hydrodynamics and Water Quality, Chapter 4*, pp. 1-42. Amer. Soc. of Civil Engr., ASCE, Reston, VA.
- Bell, G.L., Eadie, B.J., 1983. Variations in the distribution of suspended particles during an upwelling event in Lake Michigan. *J. Great Lakes Res.* 9, 559-567.
- Bennett, E.B., 1978a. Characteristics of the thermal regime of Lake Superior. *J. Great Lakes Res.* 4(3-4), 310-319.
- Bennett, E.B., 1978b. Water budgets for Lake Superior and Whitefish Bay. *J. Great Lakes Res.* 4(3-4), 331-342.
- Boyce, F.M., Hamblin, P.F., Harvey, D., Schertzer, W.M., McCrimmon, R.C., 1993. Impact of heat load on Lake Ontario to deep cooling water withdrawals and to global warming. *J. Great Lakes Res.*, 19(3), 603-616.
- Brooks, A.S., Zastrow, J.C., 2002. The potential influence of climate change on offshore primary production in Lake Michigan. *J. Great Lakes Res.* 28, 597-607.
- Bukata, R.P., Jerome, J.H., Bruton, J.E., Bennett, E.B., 1978. Relationship among optical transmission, volume reflectance, suspended mineral concentration, and Chlorophyll *a* concentration in Lake Superior. *J. Great Lakes Res.* 4(3-4), 456-461.
- Bukata, R.P., Jerome, J.H., Bruton, J.E., 1988. Relationships among Secchi disc depth, beam attenuation coefficient, and irradiance attenuation coefficient for Great Lakes waters. *J. Great Lakes Res.* 14(3), 347-355.
- Budd, J., Kerfoot, W.C., Pilant, A., Jipping, L.M., 1999. The Keweenaw Current and ice rafting: Use of satellite imagery to investigate copper-rich particle dispersion. *J. Great Lakes Res.* 25(4), 642-662.
- Chai, Y., Urban, N.R., 2004. 210Po and 210Pb distributions and residence times in the nearshore region of Lake Superior. *J. Geophys. Res.* C10, D01, 10.1029/2003JC002081, C10507.
- Chen, C., Zhu, J., Ralph, E., Green, S., Budd, J.W., Zhang, F.Y., 2001. Prognostic modeling studies of the Keweenaw current in Lake Superior. Part I: Formation and evolution, *J. Phys. Oceanogr.*, 379-395.

- Chen, C., Xu, Q., Ralph, E., Budd, J.W., Lin, H., 2004a. Response of Lake Superior to mesoscale wind forcing: A comparison between currents driven by QuikSCAT and buoy winds, *J. Geophys. Res.*, 109: C10S02, doi:10.1029/2002JC001692, 2004.
- Chen, C., Zhu, J., Zheng, L., Ralph, E., Budd, J.W., 2004b. A non-orthogonal primitive equation coastal ocean circulation model: application to Lake Superior. *J. Great Lakes Res.* 30(Supplement 1), 31-54.
- Churchill, J.H., Kerfoot, W.C., Auer, M.T., 2004. Exchange of water between the Keweenaw waterway and Lake Superior: Characteristics and forcing mechanisms. *J. Great Lakes Res.* 30(Supplement 1), 55-63.
- Clites, A.H., Quinn, F.H., 2003. The history of Lake Superior regulation: Implications for the future. *J. Great Lakes Res.* 29(1), 157-171.
- Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data, 1994. Hydraulic discharge measurements and regimen changes on the Great Lakes connecting channels and the international section of the St. Lawrence River 1841-1993.
- Croley, T.E., II., 1989. Verifiable evaporation modeling on the Laurentian Great Lakes, *Water Resour. Res.* 23, 781-792.
- Croley, T.E., II., 1990. Laurentian Great Lakes double-CO2 climate change hydrology impacts. *Climate Change* 17, 27-47.
- Croley, T.E., II, Quinn, F.H., Kunkel, K.E., Changnon, S.A., 1996. Climate transposition effects on the Great Lakes hydrological cycle. NOAA Technical Memorandum ERL GLERL-89, Great Lakes Environmental Research Laboratory, Ann Arbor, MI.
- Derecki, J.A., 1976. Multiple estimates of Lake Erie evaporation. *J. Great Lakes Res.* 2(1), 124-149.
- Derecki, J.A., 1981. Operational estimates of Lake Superior evaporation based on IFYGL findings, *Wat. Resour. Res.*, 17(5), 1453-1462.
- Diehl, S., Maanum, W., Jordan, T., Sydor, M., 1977. Transports in Lake Superior. *J. Geophys. Res.*, 82, 977-978.
- Eadie, B.J., Chamber, R.L., Gardner, W.S., Bell, G.L., 1984. Sediment trap studies in Lake Michigan: resuspension and chemical fluxes in the southern basin. *J. Great Lakes Res.* 10, 307-321.
- Feit, D.M., Goldenberg, D.S., 1976. Climatology of surface temperatures of Lake Superior, Huron, Erie and Ontario. TDL Office Note 76-16, National Weather Service, Systems Development Office, Techniques Development Laboratory, Silver Spring, MD, 20233.
- Gray, C.B., Neilson, M., Johannsson, O., Fitzsimmons, J., Millard, S., Dermott, R., 1994. Lake Superior. In: R. J. Allan, M. Dickman, C.B. Gray and V. Cromie (Eds.), *The Book of Canadian Lakes*, pp. 75-98. The Canadian Assoc. on Water Quality, Monograph Series, No. 3.
- Green, S.A., Eadie, B.J., 2004. Introduction to Special Session: Transport and transformation of biogeochemically important materials in coastal waters. *J. Geophys. Res.* 109, C10D01 doi:10.1029/2004JC002697.
- Green, T., Yeske, L.A., 1974. Surface kinetic energy transfers in the Keweenaw Current. Contribution No. 24, Marine Studies Center, University of Wisconsin.
- Harbeck, G.E., 1962. A practical field technique for measuring evaporation utilizing mass transfer theory, Prof. Paper 272-E USGS, Dept of Interior, Washington, DC.

- Harrington, M.W., 1894. Currents of the Great Lakes as deduced from the movements of bottle papers during the seasons of 1892 and 1893. U.S. Weather Bureau, Washington, DC.
- Hartmann, H.C., 1988. Historical basis for limits on Lake Superior water level regulations. *J. Great Lakes Res.* 14, 316-324.
- Hill, D.K., Magnuson, J.J., 1990. Potential effects of global climate warming on the growth and prey consumption of Great Lakes fish. *Trans. Am. Fish. Soc.* 119, 265-275.
- Hoopes, J.A., Ragotskie, R.A., Lien, S.L., Smith, N.P., 1973. Circulation patterns in Lake Superior. Tech. Rep. WIS WRC 73-04, Water Resources Centre, University of Wisconsin, Madison, Wisconsin.
- Hubbard, D.W., Spain, J.D., 1973. The structure of the early spring thermal bar in Lake Superior, p. 735-742. In: Proc. 10th Conf. Great Lakes Res. 1973 April 16-18. Internat. Assoc. Great Lakes Res., Ohio State University Huron, Ohio.
- Hultquist, T.R., Dutter, M.R., Schwab, D.J., 2006. Reexamination of the 9-10 November 1975 *Edmund Fitzgerald* storm using today's technology. *Bull. Amer. Meteorol. Soc.*, May 2006, 1-16.
- Hunter, T.S., Croley, T.E., II. 1993. Great Lakes monthly hydrologic data. NOAA Data Report ERL-GLERL, National Technical Information Service, Springfield, Virginia, 22161. <http://www.glerl.noaa.gov/data/arc/hydro/mnth-hydro.html>; Accessed May, 2002.
- Hsu, S.A., 1986. Correction of the land-based wind data for offshore application: A further evaluation. *J. Phys. Oceanogr.* 16, 390-394.
- Irbe, G., 1972. Aerial surveys of the Great Lakes water temperatures, April 1968 to March 1970. Climatology Division, Atmospheric Environment Service, Downsview, ON.
- Jones, M.L., Shuter, B.J., Zhao, Y., and Stockwell, J.D., 2006. Forecasting effects of climate changes on Great Lakes fisheries: Models that link habitat supply to population dynamics can help. *Can. J. Fish. Aquat. Sci.* 63, 457-468.
- Lam, D.C.L., 1978. Simulation of water circulations and chloride transports in Lake Superior for summer 1973. *J. Great Lakes Res.* 4(3-4), 343-349.
- Lam, D.C.L., Schertzer, W.M. (Eds.), 1999. Potential Climate Change Effects on Great Lakes Hydrodynamics and Water Quality, Amer. Soc. of Civil Engr., ASCE, Reston, VA.
- Lehman, J.T., 2002. Mixing patterns and plankton biomass of the St. Lawrence Great Lakes under climate change scenarios, *J. Great Lakes Res.* 28, 583-596.
- Lenters, J.D., 2004. Trends in the Lake Superior water budget since 1948: A weakening seasonal cycle. *J. Great Lakes Res.* 30(Sup. 1), 20-40.
- Leshkevich, G.A., Schwab, D.J., Muhr, G.C., 1993. Satellite environmental monitoring of the Great Lakes: A review of NOAA's Great Lakes CoastWatch program. *Photogrammer. Eng. Rem. Sens.* 59(3), 371-380.
- Lofgren, B.M., Zhu, Y., 1999. Seasonal climatology of surface energy fluxes on the Great Lakes, *NOAA Tech. Mem.* ERL GLERL-112.
- Lofgren, B.M., Zhu, Y., 2000. Surface energy fluxes on the Great Lakes based on satellite-observed surface temperatures. *J. Great Lakes Res.* 26(3), 305-314.
- Lofgren, B.M., Quinn, F.M., Clites, A.H., Assel, R.A., Eberhardt, A.J., Luukkoenen, C.L., 2002. Evaluation of potential impacts on Great Lakes Water Resources

- based on climate scenarios of two GCM's. *J. Great Lakes Res.* 28(4), 537-554.
- Magnuson, J.J., Webster, K.E., Assel, R.A., Bowser, C.J., Dillon, P.J., Eaton, J.G., Evans, H.E., Fee, E.J., Hall, R.I., Mortsch, L.R., Schindler, D.W., and Quinn, F.H., 1997. Potential effects of climate changes on aquatic systems: Laurentian Great Lakes and Precambrian Shield Region, *Hydrol. Processes* 11, 825-871.
- Matheson, D.H., Munawar, M., 1978. Lake Superior basin and its development. *J. Great Lakes Res.* 4(3-4), 249-263.
- McCormick, M.J., 1990. Potential changes of thermal structure and cycle of Lake Michigan due to global warming, *Trans. Am. Fish. Soc.* 119, 183-194.
- Millar, F.G., 1952. Surface temperature of the Great Lakes, *J. Fish. Res. Board Can.*, 9, 329-371.
- Mortimer, C.H., 1971. Large-scale oscillatory motions and seasonal temperature changes in Lake Michigan and Lake Ontario, Part I, Text. Special Rep. No 12. Centre for Great Lakes Studies, University of Wisconsin-Milwaukee.
- Morton, F.I., 1967. Evaporation from large deep lakes, *Wat. Resour. Res.* 3, 181-200.
- Munawar, M., 1978. Preface: Limnology of Lake Superior, *J. Great Lakes Res.* 4(3-4).
- Neibauer, H.J., 1976. Wind driven coastal upwelling in Lake Superior, Ph.D. Thesis, University of Wisconsin-Madison.
- Neibauer, H.J., Green, T., Ragotskie, R.A., 1977. Coastal upwelling/downwelling cycles in southern Lake Superior, *J. Phys. Oceanogr.*, 7(6), 918-927.
- Phillips, D.W., 1978. Environmental climatology of Lake Superior. *J. Great Lakes Res.* 4(3-4), 288-309.
- Quinn, F.H., 1979. An improved aerodynamic evaporation technique for large lakes with application to the International Field Year for the Great Lakes, *Water Resour. Res.* 15(4), 935-940.
- Quinn, F.H., den Hartog, G., 1981. Evaporation Synthesis, In: E.J. Aubert and T.L. Richards (Eds.), *IFYGL- The International Field Year for the Great Lakes*, pp. 221-245. NOAA-GLERL, U.S. Dept. of Commerce, Ann Arbor, MI.
- Ragotskie, R.A., Neibauer, H.J., 1975. Temperature regime in Lake Superior at Silver Bay, Minnesota. *Verh. Internat. Verein. Limnol.* 19, 144-148.
- Ragotskie, R.A., Bratnick, M., 1965. Infrared temperature patterns on Lake Superior and inferred vertical motions. Tech. Rep. No. 27, Dept. of Meteorology, University of Wisconsin-Madison.
- Ralph, E.A., 2002. Scales and structures of large eddies, *Geophys. Res. Letters* 29(24.2177) 30-1-30-4.
- Rao, D.B. Murty, T.S., 1970. Calculation of the steady-state wind-driven circulation in Lake Ontario, *Arch. Meteor. Geophys., Bioklim.* A19, 195-210.
- Rao, Y.R., Murthy, C.R., 2001a. Coastal boundary layer characteristics during summer stratification in Lake Ontario, *J. Phys. Oceanogr.* 31, 1088-1104.
- Rao, Y.R., Skafel, M.G., Charlton M.N., 2004. Circulation and turbulent exchange characteristics during the thermal bar in Lake Ontario, *Limnol. Oceanogr.* 49, 2190-2200.
- Richards, T.L., Irbe, J.G., Massey, D.G., 1969. Aerial surveys of Great Lakes water temperatures, April 1966 to March 1968. Climatology Division, Atmospheric Environment Service, Downsview, ON.
- Richards, T.L., Irbe, J.G., 1969. Estimates of monthly evaporation losses from the Great Lakes 1950 to 1968. based on the mass transfer technique. p.469-487. In: *Proc. 12th Conf. Great Lakes Res., Internat. Assoc. Great Lakes Res.*

- Rodgers, G.K., 1965. The thermal bar in the Laurentian Great Lakes, In. Proc. 8th Conf. on Great Lakes Research, Internat. Assoc. for Great Lakes Res.
- Rondy, D.R., 1969. Great Lakes Ice Atlas, U.S. Lake Survey Research Report, 5-6.
- Ruschmeyer, O.R., Olson, T.A., Bosch, H.H., 1958. Lake Superior Studies, Minnesota Public School of Health, University of Minnesota, Duluth.
- Schertzer, W.M., 1978. Energy budget and monthly evaporation estimates for Lake Superior, 1973. *J. Great Lakes Res.* 4(3-4), 320-330.
- Schertzer, W.M., Elder, F.C., Jerome, J., 1978. Water transparency of Lake Superior in 1973. *J. Great Lakes Res.* 4(3-4), 350-358.
- Schertzer, W.M., Sawchuk, A.M., 1990. Thermal structure of the lower Great Lakes in a warm year: Implications for the Occurrence of hypolimnion anoxia, *Trans. Amer. Fish. Soc.* 119(2), 195-209.
- Schertzer, W.M., Taylor, B., 2008. Modelling evaporation from Lake Okanagan, mainstem lakes, upland lakes and reservoirs based on existing databases. In: Kellie Bunting (Ed.), *Proc. One Watershed – One Water Conference*, 2008 Oct. 21-23. Kelowna, BC,
- Schwab, D.J., 1978. Simulation and forecasting of Lake Erie storm surge. *Mon. Weather Rev.*, 106, 1476-1487.
- Schwab, D.J., Bedford, K., 1994. Initial Implementation of the Great Lakes Forecasting System: A real-time system for predicting lake circulation and thermal structure, *Water Poll. Res. J. Canada* 29, 203-220.
- Schwab, D.J., Leshkevich, G.A., Muhr, G.C., 1999. Automated mapping of surface water temperature in the Great Lakes, *J. Great Lakes Res.*, 25(3), 468-481.
- Shuter, V., Stortz, K., Oman, G., Sydor, 1978. Turbidity dispersion in Lake Superior through use of Landsat data., *J. Great Lakes Res.* 4(3-4), 359-360.
- Simons, T.J., 1973. Development of three-dimensional numerical models of the Great Lakes, CCIW Scientific Series No. 12.
- Simons, T.J., Schertzer, W.M., 1987. Stratification, currents and up-welling in Lake Ontario, summer 1982. *Can. J. Fish. Aquat. Sci.*, 44, 2047-2058.
- Sloss, P.W., Sayor, J.H., 1975. Large-scale current measurements in Lake Superior, NOAA Tech. Rep. ERL 363 GLERL 8.
- Sly, P.G., Thomas, R.L., 1974. Review of Geological Research as it Relates to an Understanding of Great Lakes Limnology. *J. Fish. Res. Board of Can.* 31, 795-825.
- Smith, N., 1972. Summertime temperature and circulation patterns in Lake Superior, Ph.D. dissertation, Dept. Oceanogr. University of Wisconsin, Madison, WI.
- Smith, N.P., Ragotskie, R.A., 1970. A comparison of computed and measured currents in Lake Superior, *Proc. 13th Conf. Great Lakes Res., Internat. Assoc. Great Lakes Res.*, 969-977.
- Smith, N.P., Ragotskie, R.A., 1971. Some characteristics of the temperature recorder time-series from Lake Superior. In. *Proc. 14th Conf. Great Lakes Res., Internat. Assoc. Great Lakes Res.*, 670-676.
- Snyder, F.F., 1960. Evaporation on the Great Lakes, *Internat. Assoc. of Science: Hydrology, General Assembly, Helsinki, Publ. No. 53.*
- Spain, J.D., Wernert, G.M., Hubbard, D.W., 1975. The structure of the spring thermal bar in Lake Superior, II., *J. Great Lakes Res.* 2, 296-306.
- Thomas, R.L., Dell, C.I., 1978. Sediments of Lake Erie, *J. Great Lakes Res.* 4(3-4), 264-275.

- Tikhomirov, A., 1963. The thermal bar of Lake Ladoga. *Proc. of the All-Union Geographic Soc.* 95(2), 134-142.
- Urban, N.R., Jeong, J., Chai, Y., 2004a. The benthic nepheloid layer (BNL) north of the Keweenaw Peninsula in Lake Superior: Composition, dynamics and role in sediment transport. *J. Great Lakes Res.* 30(Supplement 1), 133-146.
- Urban, N.R., Lu, X., Chai, Y., Apul, D.S., 2004b. Sediment trap studies in Lake Superior: Insights into resuspension, cross-margin transport, and carbon cycling. *J. Great Lakes Res.* 30 (Supplement 1), 147-161.
- Ullman, D., Brown, J., Cornillion, P., Mavor, T., 1998. Surface temperature fronts in the Great Lakes, *J. Great Lakes Res.* 24(4), 753-775.
- USCG (United States Coast Guard), 1977. Marine Board Casualty Report, SS Edmund Fitzgerald; Sinking in Lake Superior on 10 November, 1975 with loss of life. U.S. Coast Guard Marine Board of investigation Report and Commander's Action, Rep. USCG 156732/64216.
- Yeske, L.A., Green, T., 1975. Short-period variations in a great lake coastal current by aerial photogrammetry, *J. Phys. Oceanogr.* 5, 125-135.
- Zhu, J., Chen, C., Ralph, E., S. A. Green, S.A., Budd, J.W., and Zhang, F.Y., 2001: Prognostic modeling studies of the Keweenaw Current in Lake Superior. Part II: Simulation, *J. Phys. Oceanogr.*, 31, 396-410.