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CLIMATE TRANSPOSITION EFFECTS ON THE GREAT LAKES HYDROLOGICAL CYCLE

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NOTICE

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CLIMATE TRANSPOSITION EFFECTS ON THE GREAT LAKES HYDROLOGICAL CYCLE[1]

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ABSTRACT. Historical climate scenarios, based on 41-yr data periods from the southeastern and southwestern continental United States, were used in hydrological models of the Great Lakes to examine possible changes in variability associated with various hydrological conditions. The Great Lakes Environmental Research Laboratory (GLERL) used their conceptual models for simulating moisture storages in, and runoff from, the 121 watersheds draining into the Laurentian Great Lakes, over-lake precipitation into each lake, and the heat storages in, and evaporation from, each lake. GLERL combined these components as net water supplies for each lake and estimated lake levels and connecting channel flows to consider transposed climate scenarios. We transposed four climate zones, ranging from 6° south and 0° west to 10° south and 11° west of the Great Lakes, to the Great Lakes area. These represent analog climates that could occur over the Great Lakes basin under global warming. This transposition of actual climates was essential since it incorporates natural changes in variability within the existing climate; this is not true for GCM-generated corrections applied to existing historical data in many other hydrological impact assessment studies. Average air temperatures increased between 4 and 11°C, and precipitation ranged from 80% to +170% of the current climate, over various lakes under various scenarios. These resulted in Great Lakes whole-basin water supply changes from the current condition of -1% to -54%. The higher air temperatures under the transposed climate scenarios led to higher over-land evapotranspiration and lower runoff to the lakes with earlier runoff peaks, since snow-pack is reduced up to 100%, and the snow season is eliminated in some scenarios. This also resulted in a reduction in available soil moisture. Water temperatures increased and peaked earlier; heat resident in the deep lakes increased throughout the year. Mixing of the water column diminished, as most of the lakes become mostly monomictic, and lake evaporation increased. Water supplies decreased dramatically for the two driest scenarios with Lake Superior becoming a terminal lake. Also, lake level variability increased for all lakes for most of the scenarios. Maximum lake levels exceeded the recorded maximums for several scenarios on the lower lakes.

1. INTRODUCTION

Climatic change will impact many aspects of the hydrological cycle with consequences for mankind that are interrelated and often-times difficult to discern. Climate warming will have impacts on Great Lakes water supply components, and basin storages of water and heat, that must be understood before lake level impacts can be assessed. Because the Laurentian Great Lakes possess tremendous water and heat storage capacities, they respond slowly to changed meteorological inputs. This "memory" results in a filtering or dampening of most short-term meteorological fluctuations and a response to longer-period fluctuations characteristic of climate change. Thus, the large Great Lakes system is ideal for studying regional effects of climate changes.

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1.1 Early Great Lakes Climate Change Impact Studies

Considerations of future climate situations that may occur (scenarios) help to identify possible effects and can bound future conditions, if widely different scenarios are tested. Preliminary impact estimates considered simple constant changes in air temperature or precipitation. Quinn and Croley (1983) estimated net basin supply to Lakes Superior and Erie. Cohen (1986) estimated net basin supply to all Great Lakes. Quinn (1988) estimated lower water levels due to decreases in net basin supplies on Lakes Michigan-Huron, St. Clair, and Erie.

Researchers have run general circulation models (GCMs) of the earth's atmosphere to simulate climates for current conditions and for a doubling of global carbon dioxide levels ($2xCO_2$). They used a larger-than-regional scale for many internally-consistent daily meteorological variables. The U.S. Environmental Protection Agency (USEPA, 1984) and Rind (personal communication, 1988) used the hydrological components of general circulation models. They assessed changes in water availability in several regions throughout North America, but the regions were very large. Rind used only four regions for the entire continent and indicated the need for smaller region assessments. Regional hydrological models can link to GCM outputs to assess changes associated with climate change scenarios. Allsopp and Cohen (1986) used Goddard Institute of Space Sciences (GISS) $2xCO_2$ climate scenarios with net basin supply estimates.

Other efforts that linked hydrological models to GCM outputs originated in studies commissioned by the U.S. EPA. The EPA, at the direction of the U.S. Congress, coordinated several regional studies of the potential effects of a $2xCO_2$ atmosphere. The studies addressed various aspects of society, including agriculture, forestry, and water resources (USEPA, 1989). They directed others to consider alternate climate scenarios by changing historical meteorology similar to the changes observed in GCM simulations of $2xCO_2$, observing changed process model outputs, and comparing to model results from unchanged data. Cohen (1990a, 1991) discusses other studies that use this type of linkage methodology and also presents his concerns for comparability between studies using different types.

1.2 Recent Great Lakes Climate Change Impact Studies

1.2.1 GLERL–EPA $2xCO_2$ Impacts

As part of the EPA study, GLERL assessed steady-state and transient changes in Great Lakes hydrology consequent with simulated $2xCO_2$ atmospheric scenarios from three GCMs (Croley, 1990; Hartmann, 1990; USEPA, 1989). EPA required that GLERL first simulate 30 years of "present" Great Lakes hydrology by using historical daily data with present diversions and channel conditions. GLERL arbitrarily set initial conditions but used an initialization period to allow their models to converge to conditions initial to the simulation. GLERL repeated their simulation, with initial conditions set equal to the averages over the simulation period, until these averages were unchanging. This facilitated investigation of "steady-state" conditions. The next step was to conduct simulations with adjusted data sets.

EPA obtained output from atmospheric GCM simulations, representing both "present" and $2xCO_2$ steady-state conditions, from GISS, the Geophysical Fluid Dynamics Laboratory (GFDL), and the Oregon State University (OSU). They supplied monthly adjustments of "present" to $2xCO_2$ for each meteorological variable. GLERL applied them to daily historical data sets to estimate 33-year sequences of atmospheric conditions associated with the $2xCO_2$ scenarios. This method keeps spatial and temporal (interannual, seasonal, and daily) variability the same in the adjusted data sets as in the historical base period. GLERL then used the $2xCO_2$ scenarios in hydrology impact model simulations similar to those for the base case scenario. They interpreted differences between the $2xCO_2$ scenario and the base case scenario as resulting from the changed climate. They observed that the three scenarios changed precipitation little, but snowmelt and runoff were greatly decreased, evapotranspiration and lake evaporation were greatly increased, and net basin supplies to the lakes and lake levels were decreased. The scenario derived from the GFDL GCM was the most extreme, with evaporation 44% higher than the base case, and net basin supply less than 50% of the base case.

1.2.2 GLERL–IJC $2xCO_2$ Impacts

The EPA studies, in part, and the high water levels of the mid 1980s prompted the International Joint Commission (IJC) to reassess climate change impacts on Great Lakes hydrology and lake thermal structure. GLERL adapted the EPA study methodology for the IJC studies (Croley, 1992b) to consider $2xCO_2$ GCM scenarios supplied by the Canadian Climate Centre (CCC) for the period 1948–88. GLERL's procedure to estimate "steady-state" suggested, for a few subbasins, very different initial groundwater storages than were used in model calibrations. Since there is little confidence in estimates of very large groundwater half-lives on these subbasins with only 10–20 years in calibrations, the initial values used in calibrations were also used in the simulations for those subbasins.

Average monthly meteorological outputs were supplied for each month of the year over a 1º latitude by 1º longitude grid (Louie, 1991) by the CCC as resulting from their second-generation GCM; see McFarlane (1991). GLERL computed $2xCO_2$ monthly adjustments at each location, used them with historical data to estimate the $2xCO_2$ 41-year sequences (1948–88) for each Great Lake basin, and then used the $2xCO_2$ scenario in simulations similar to the base case as before. This scenario proved similar to the earlier GFDL-based scenario in that net basin supplies were reduced to almost 50% of the base case. However, the CCC-based scenario reduced runoff more and evaporation less than the GFDL-based scenario.

Other EPA studies included partial assessments of large-lake heat storage associated with climate change on Lakes Michigan (McCormick, 1989) and Erie (Blumberg and DiToro, 1989). The IJC study looked in less detail but more breadth at large-lake thermodynamics in that while only lake-wide effects were considered, all lakes were assessed.

1.3 GCM Linkage Problems

The hydrological study results from the EPA and IJC studies should be used with caution. They are, of course, dependent on GCM outputs with inherent large uncertainties in the GCM components, assumptions, and data. Transfer of information between the GCMs and GLERL's hydrological models in the manners described above involves several assumptions. Solar insolation at the top of and through the atmosphere on a clear day are assumed to be unchanged under the changed climate, modified only by cloud cover changes. Over-water corrections are made in the same way, albeit with changed meteorology, which presumes that over-water/over-land atmospheric relationships are unchanged.

Heat budget data from GCM simulations for Great Lakes grid points may not adequately describe conditions over the lakes due to the coarse resolution of the grids. GLERL's procedure for transferring information from the GCM grid is an objective approach but simple in concept. It ignores interdependencies in the various meteorological variables as all are averaged independently in the same manner. Of secondary importance, the spatial averaging of meteorological values over a box centered on the GCM grid point (implicit in the use of the nearest grid point to each square kilometer of interest) filters all variability that exists in the GCM output over that box. If GCM output were interpolated between these point values, then at least some of the spatial variability might be preserved. The interpolation performed by Louie (1991) from the original GCM grid to a finer grid reduced this problem, but it still exists in the use of the finer grid with the hydrology models. Of course, little is known about the validity of various spatial interpolation schemes and, for highly variable spatial data, they may be inappropriate. Furthermore, much of the variability at the smallest resolvable scale of GCMs is, unfortunately, spurious.

Spatial and temporal variabilities in meteorology of the $2xCO_2$ data sets are the same as the base case, in both the EPA and IJC studies. The methodology does not address changes in variabilities that would take place under a changed climate. The method of coupling does not reproduce seasonal timing differences under a changed climate from the GCMs but preserves seasonal meteorological patterns as they exist in the historical (base case) data. This is a result of applying simple ratios or differences to calculate $2xCO_2$ scenarios from base case scenarios. This implicitly ignores spatial and temporal phase and frequency changes consequent in the $2xCO_2$ GCM simulations. For example, a changed climate alters the movement (direction, speed, frequencies) of air masses over the lakes. This implies an alteration of the seasonal temporal structure for storms and cyclonic events as well as the intensities of storms. The above method only allows modification of the latter. Seasonal changes induced by the changed meteorology because of a time-lag storage effect are observable, however. Shifts in snowpack or in the growth and decay of water surface temperatures are examples. Changes in annual variability are less clear, again as a result of using the same historical time structure for both the base case and the changed climate scenarios.

Finally, the use of GCM outputs in the EPA and IJC studies, to drive GLERL's hydrological process models, forced the use of inappropriately large spatial and temporal scales for studying the Great Lakes impacts of climate change. While the hydrological process models were defined over daily intervals and subbasin areas averaging 4,300 km^2, the GCM adjustments were made over monthly time intervals and grids of 7.83° latitude by 10° longitude (GISS), 4.44° by 7.5° (GFDL), 4° by 5° (OSU), and 3.75° by 3.75° interpolated to 1° by 1°(CCC GCM).

1.4 Climate Transposition

While the EPA and IJC studies looked at changes in the mean values of hydrological variables, changes in variability were unaddressed. This variability is the singular key problem for shipping, power production, and resource managers. GLERL and the Midwest Climate Center (MCC) recognized the importance of investigating the effects of shifts in the daily, seasonal, inter annual, and multi-year climate variability on lake net supply behavior, as well as related changes in mean supplies. They considered studies that used climate change scenarios that were not drawn so directly from historical data that they preserved historical spatial and temporal patterns. Changnon and Quinn (1989) developed synthetic 12-year extreme wet and dry climate scenarios from historical records of the Great Lakes basin to examine effects on the basin's hydrological components. Such "instrumental analogues" are one empirical approach identified by Robinson and Finkelstein (1989) to develop realistic scenarios since the actual values of the past were used to form the wet and dry extremes. However, the climate changes represented by these 12-year time series were not as large as many GCMs predict could happen in the future over the basin, and the effect of weather fluctuations over time due to large climatic changes could not be assessed by that approach. Atmospheric modelers are developing nested mesoscale numerical models of the Great Lakes basin (Bates et al., 1994) but these are not yet capable of generating multi-decadal series of conditions essential for the sensitivity study. In summary, these approaches simply could not provide the spatial and long temporal climatic data needed for the hydrological model and cannot accomplish the desired sensitivity study of fluctuations in the hydrological system of the Great Lakes.

GLERL and MCC investigated these changes in variability by utilizing data for climates that actually exist to the south and west of the Great Lakes and that resemble some of the $2xCO_2$ GCM scenarios. Lengthy (at least 40 years) and detailed records of daily weather conditions at about 2000 sites are available to represent physically plausible and coherent scenarios of alternate climates. Such data sets incorporate reasonable values and frequencies of extreme events, ensuring that the desired temporal and spatial variabilities are represented, and are being transposed over the Great Lakes.

MCC supplied the data, and GLERL transposed them to the Great Lakes by relocating all meteorological station data and Thiessen-weighting to obtain areal averages over the 121 watersheds and 7 lake surfaces for all days of record (1948–1992). GLERL also reduced all historical data (base case) within the Great Lakes (1900–1990). This involved extensive error checking and data correction for thousands of stations, and regeneration of areal averages. Since the Great Lakes affect the climate near the shoreline and these effects are not present in the transposed data sets, MCC prepared maps of generalized seasonal lake effects on the area's meteorology, to be applied to the transposed climates.

The Great Lakes hydrology of each transposed climate is estimated, as before, by applying the system of hydrological models to these data sets (but this time, directly) and comparing outputs for each transposed climate to a base case derived with the models from historical meteorological data. This approach allows preservation of reasonable spatial and temporal variations in meteorology and preserves the interdependencies that exist between the various meteorological variables. It also allows the use of appropriate spatial and temporal scales, better matching the models than do the GCM output corrections.

The GLERL–MCC study of transposed climates in the Great Lakes basin is presented here. The following chapter describes the present Great Lakes climate, the physical characteristics of the Great Lakes and the dynamics of these large water bodies. Chapter 3 outlines the methodology of climate transposition, including descriptions of transposed data sets and the climates they represent. Next, the hydrological models for basin runoff, over-lake precipitation, and lake thermodynamics are described in Chapter 4, and results from these models are presented in Chapter 5. The models for channel routing, lake regulation, diversions, and consumptions are presented with their results in Chapter 6. Chapter 7 recapitulates the major points of this research.

2. GREAT LAKES DYNAMICS AND CLIMATE

There is a major tendency to think of Great Lakes water levels in terms of extremes rather than of normal conditions. Within recent memory we had the record low lake levels of 1964. This resulted in docks sitting out of the water, insufficient depths for navigation in many harbors and channels, and greatly reduced recreational opportunities. These low levels were followed in 1973 by record high lake levels with resultant flooding and shore damage and erosion. The lake levels remained high until 1986, when they returned to near-average conditions, and new record highs were once again set on Lakes Superior, Michigan-Huron, St. Clair, and Erie.

This section presents an overview of the physical characteristics of the Great Lakes from a water quantity perspective, outlines the basin and lake physical processes, summarizes the climatology of the Great Lakes, examines the types of natural lake level fluctuations and their causes, compares the natural fluctuations with existing diversions and regulation effects, describes current conditions, and concludes with a long-term perspective on lake levels.

2.1 Great Lakes Overview

The Great Lakes basin, shown in Figure 1, contains an area of approximately 770,000 km^2 (300,000 mi^2), about one-third of which is water surface. Cursory descriptions are given by Freeman and Haras (1978), the U. S. Army Corps of Engineers (1985), and the Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data (1977). The basin extends some 3,200 km (2,000 mi) from the western edge of Lake Superior to the Moses-Saunders Power Dam on the St. Lawrence River. The water surface drops in a cascade over this distance some 180 m (600 ft) to sea level. The most upstream, largest, and deepest lake, is Lake Superior. The lake has two interbasin diversions of water into the system from the Hudson Bay Basin: the Long Lac and Ogoki Diversions. Lake Superior waters flow through the lock and compensating works at Sault Ste. Marie and down the St. Marys River into Lake Huron where it is joined by water flowing from Lake Michigan. Lake Superior is completely regulated, to balance Lakes Superior, Michigan, and Huron water levels, according to Regulation Plan 1977, under the auspices of the International Joint Commission (International Lake Superior Board of Control 1981, 1982).

Lakes Michigan and Huron are considered to be one lake hydraulically because of their connection through the deep Straits of Mackinac. An interbasin diversion takes place from Lake Michigan at Chicago. Here water is diverted from the Great Lakes to the Mississippi River Basin. The water flows from Lake Huron through the St. Clair River, Lake St. Clair, and Detroit River system into Lake Erie. The

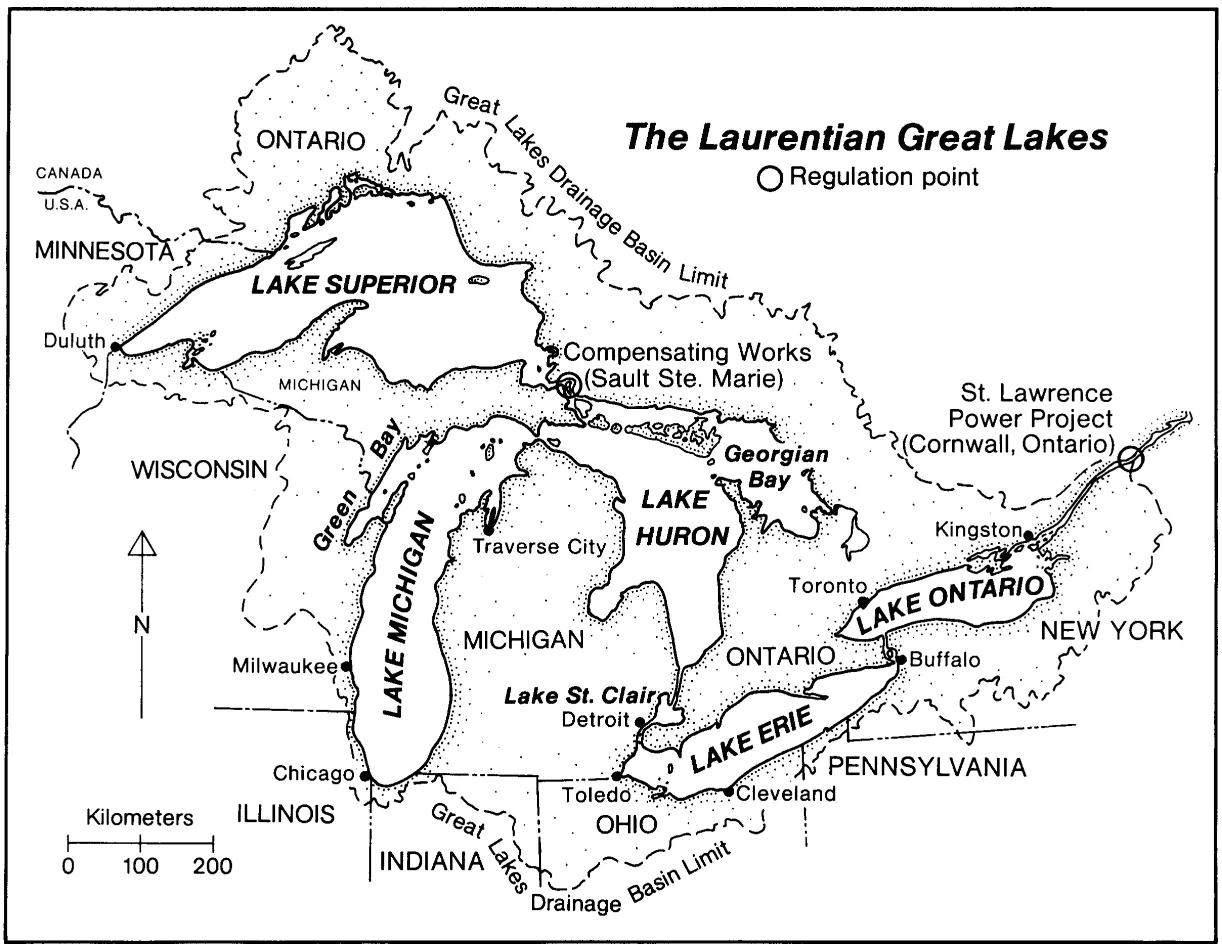


Figure 1.--Great Lakes Basin.

drop in water surface between Lakes Michigan-Huron and Lake Erie is only about 2 m (8 ft). This results in a large backwater effect between Lakes Erie, St. Clair, and Michigan-Huron; changes in Lakes St. Clair and Erie levels are transmitted upstream to Lakes Michigan and Huron. From Lake Erie, the flow is through the Niagara River and Welland Diversion into Lake Ontario. The major drop over Niagara Falls precludes changes on Lake Ontario from being transmitted to the upstream lakes. The Welland Diversion is an intrabasin diversion bypassing Niagara Falls and is used for navigation and hydropower. There is also a small diversion into the New York State Barge Canal System which is ultimately discharged into Lake Ontario. Lake Ontario is completely regulated in accordance with Regulation Plan 1958D to balance damages upstream on Lake Ontario with those downstream on the St. Lawrence Seaway [estimated to have lowered Lake Ontario 0.75 m (2.5 ft) during the record high water levels of 1986]. The outflows are controlled by the Moses-Saunders Power Dam between Massena, New York and Cornwall, Ontario. From Lake Ontario, the water flows through the St. Lawrence River to the Gulf of St. Lawrence and to the ocean.

Lakes Superior, Michigan, Huron, and Ontario are very deep, while Lakes Erie and St. Clair are very shallow. Table 1 contains pertinent gross statistics on the sizes of the Great Lakes, Lake St. Clair, and their basins.

2.2 Physical Processes

The behavior of the Laurentian Great Lakes system is governed by its huge storages of water and energy. There are three main conservation laws to consider relative to these huge storages: 1) mass balances in the basins, 2) mass balances in the lakes, and 3) energy balances in the lakes. There are also mass and energy balances to consider for the lakes' ice cover. The first conservation law (mass balance on the basins) comprises the primary process determining lake levels: the hydrological cycle of the Great Lakes Basin (Croley 1983a). As shown in Figure 2, precipitation enters the snowpack, if present, and is then available as snow melt, depending mainly on air temperature and solar radiation. Snow melt and rainfall partly infiltrate into the soil and partly run off directly to rivers, depending upon the moisture content of the soil. Infiltration is high if the soil is dry, and surface runoff is high if the soil is saturated. Soil moisture evaporates or is transpired by vegetation depending upon the types of vegetation, the

Table 1.--Laurentian Great Lake Size Statistics[a].

Characteristic		Superior	Michigan	Huron	St. Clair	Erie	Ontario
Basin Area[b]	km^2	128,000	118,000	131,000	12,400	58,800	60,600
	mi^2	49,300	45,600	50,700	4,800	22,700	23,400
Surface Area	km^2	82,100	57,800	59,600	1,114	25,700	18,960
	mi^2	31,700	22,316	23,000	430	9,920	7,320
Volume	km^3	12,100	4,920	3,540	3	484	1,640
	mi^3	2,900	1,180	850	1	116	393
Average Depth	m	147	85	59	3	19	86
	ft	482	280	190	10	62	280
Maximum Depth	m	405	281	229	6	64	244
	ft	1,330	923	750	21	210	802

[a]Reference: Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data (1977).
[b]This does not include the surface area of the lake.

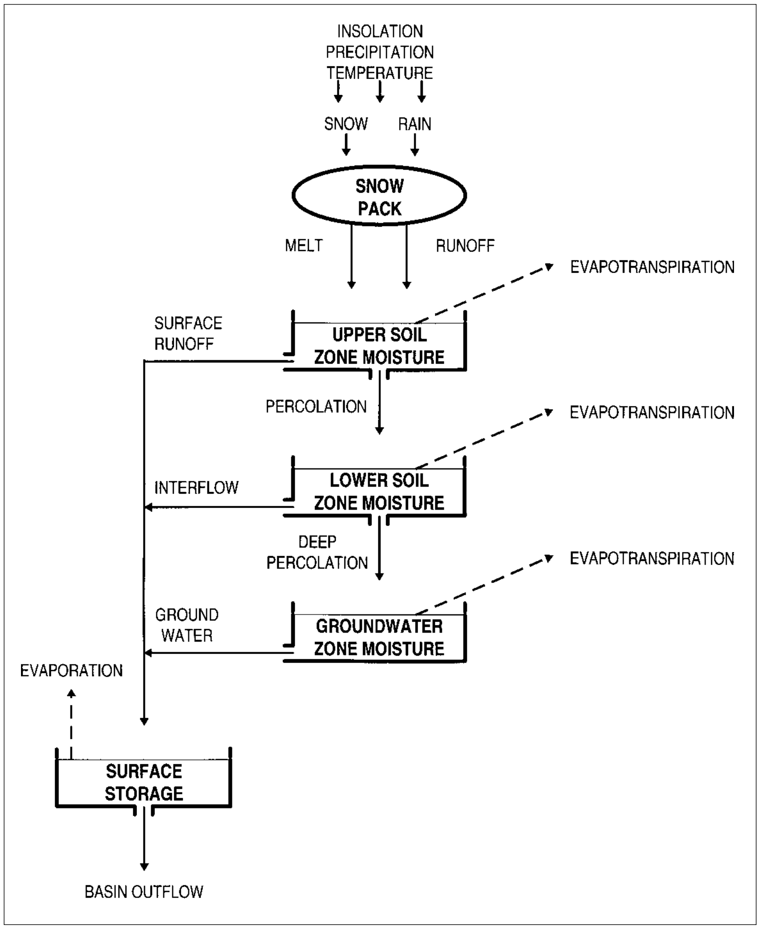


Figure 2.--Great Lakes Subbasin Mass Balance.

season, solar radiation, air temperature, humidity, and wind speed. The remainder percolates into deeper basin storages, which feed the rivers and lakes through interflows and groundwater flows. Generally, these river supplies are high if the soil and groundwater storages are large. Because of this buffering effect of the large snowpack and the large soil, groundwater, and surface storages, runoff from rivers into a lake can remain high for many months or years after high precipitation has stopped.

Mass conservation in the lake is the next major determinant of lake levels. Major sources of water into a lake include precipitation on the land basin, which results in runoff into the lake, precipitation over the lake surface, inflow from upstream lakes, and diversions into the lake. Net groundwater flows directly to each of the Great Lakes are generally neglected (DeCooke and Witherspoon, 1981). The outflows consist of evaporation from the lake surface, flow to downstream lakes, and diversions. The imbalance between the inflow and outflow results in lake levels either rising if there is more inflow than outflow, represented by a positive change in storage, or falling if there is more outflow than inflow, represented by a negative change in storage. The large lake water storages provide a buffering of the input fluctuations with regard to output variations. The large surface areas of the lakes enable large storage changes with very small water level changes; hence, outputs (which are a function of water levels) change slowly.

Energy conservation in a lake actually must be considered together with a lake's mass balance. Lake heat storage is a function of the lake's size and shape and of its surface inputs of solar insolation and reflection (short wave exchanges), thermal and atmospheric emission (net long wave exchange), conduction to the atmosphere (sensible heat transfer), heat loss through evaporation (latent and some advection), other advection terms (precipitation, inflows, and outflows), and ice growth and melt. Evaporation is a function of surface temperature (heat storage), air temperature (atmospheric stability), humidity, and wind speed. Water surface temperatures generally peak in August (September for Superior) at 15–25°C resulting in a stable summertime temperature stratification in the water column (high-density cool water at depth, and low-density warm water at the surface). Surface temperatures drop during the fall and winter, and the water column in each lake "turns over" as temperatures drop through 4°C where water density is maximum (deep now-lighter waters rise and mix with now-heavier surface layers). Turn over occurs again in the spring as surface temperatures rise to that of maximum density.

There is also extensive ice cover on most of the lakes during most winters. Lake Superior averages about 75% ice-covered, Michigan is 45%, Huron is 68%, Erie is 45%, and Ontario is 24%. Ice formation and breakup is governed by additional mass and energy balances that take place simultaneously with those of the lakes' water bodies. The Great Lakes do not ordinarily freeze-over completely (Assel et al. 1983) because of the combination of their large heat storage capacity, large surface area, and their location in the mid-latitude winter storm track. Alternating periods of mild and cold air temperatures combine with episodic high and low wind stresses at the water surface to produce transitory ice conditions during the winter. Ice cover in mid-lake regions is often in motion. Lake Erie ice speeds have been observed to average 8 cm/s with a maximum speed of 46 cm/s (Campbell et al. 1987). Ice can form, melt, or be advected toward or from most mid-lake areas throughout the winter (Rondy, 1976). When ice is advected into areas with existing ice cover, it can under- or over-ride the ice cover, forming rafted rubble 5–10 m thick. The normal seasonal progression of ice formation begins in the shallow shore areas of the Great Lakes in December and January. The deeper mid-lake areas normally do not form extensive ice cover until February and March. Ice is lost over all lake areas during the last half of March and during April.

Ice formation alters the surface thermodynamics of the lakes, changing subsequent ice formation, surface heating or cooling, lake evaporation, and lake responses to atmospheric changes. The large heat storages of the lakes provide a buffering; they forestall and reduce ice formation and shift the large evaporation response. Water temperatures lag air temperatures, and evaporation lags surface heating (insolation). Evaporation peaks in October–November on Lake Erie and in November–December on Lake Superior.

The large basin and lake storages of water and ice and the large lake and ice storages of energy represent an "intrinsic memory" that allow scientists to forecast basin moisture storage and runoff (basin storage buffering) in the face of uncertain meteorology. It also allows prediction of evaporation (heat storage buffering) and lake levels (lake storage buffering) of up to about 6 months of low-frequency changes. It further enables estimation of ice formation amounts and timing as well as all secondary hydrological variables.

2.3 Climatology

Precipitation causes the major long-term variations in lake levels (Quinn and Croley, 1981; Quinn, 1985). Table 2 shows that annual precipitation ranges from about 82 cm (32 in) for Superior to 93 cm (37 in) for Ontario. Figure 3 depicts total annual precipitation over Lakes Michigan-Huron, St. Clair, and Erie for the 1900–90 period (Quinn 1981; Quinn and Norton 1982). From 1900 through 1939, a low precipitation regime predominated with the majority of the years falling below the mean. From about 1940 until recently, a high precipitation regime has existed. Of particular interest is the high precipitation in the early 1950s, the low precipitation in the early 1960s that led to the record lows, and a consistently very high precipitation regime from the late 1960s through the late 1980s. Table 3 summarizes Great Lakes annual precipitation totals by basin for several periods. Of particular interest are the progressions

Table 2.--Partial Great Lakes Annual Water Balance (1951-1988).

Component	Superior		Michigan		Huron		Erie		Ontario	
	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)
Lake Precipitation[a]	82	32	83	32	87	34	81	36	93	37
Lake Runoff[a]	62	24	64	25	84	33	80	32	169	67
Lake Evaporation[a]	56	22	65	25	63	25	90	35	67	26

[a]Equivalent depth over the lake area.

Table 3.--Great Lakes Annual Precipitation Summary.

Period	Superior		Michigan		Huron		Erie		Ontario	
	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)
1900-39	72	29	78	31	77	31	85	34	86	34
1940-90	81	32	82	33	86	34	89	35	93	37
1970-90	84	33	86	34	89	35	94	37	98	39
1985	98[a]	39[a]	102[a]	40[a]	105[a]	41[a]	106	42	100	40
1900-69[b]	75	30	79	31	80	32	87	34	87	34
1900-90[b]	79	31	84	33	84	33	89	35	88	35

[a]Record high for 1900-90.
[b]Long-term period averages are supplied for comparison.

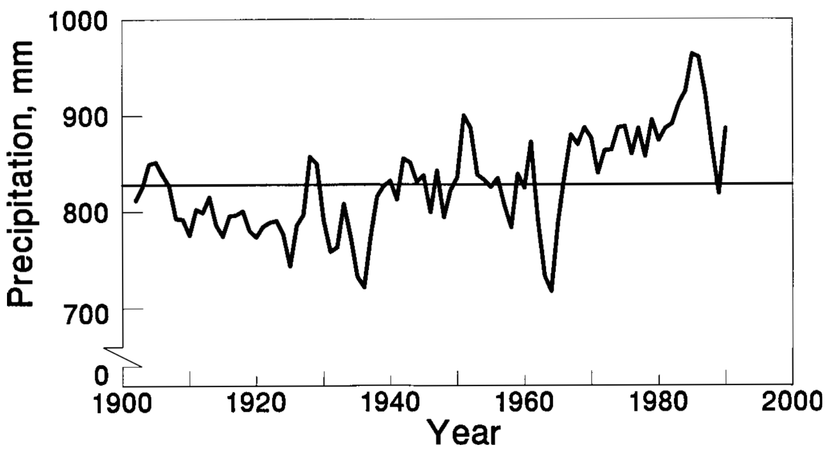


Figure 3.--Lakes Michigan, Huron, St. Clair, and Erie 3-year Mean Precipitation (1900-90).

of increasing precipitation for each basin. While the 1940–90 period is generally above normal (2–8% higher than the 1900–69 average and 2–6% higher than the 1900–90 average), the last 20 of those years are higher still (8–13% higher than the 1900–69 average and 2–11% higher than the 1900–90 average); 1985 set many new records with the highest precipitation to that date (8–40% higher than the 1900–69 average and 7–33% higher than the 1900–90 average).

Variations in air temperature also influence lake level fluctuations. At higher air temperatures, plants tend to use more water, resulting in more transpiration, and there are higher rates of evaporation from both the ground surface and the lake. This yields less runoff for the same amount of precipitation than would exist during a low temperature period when there is less evaporation and transpiration. Coupled with the higher lake evaporation, lake levels drop with increasing air temperature, all other things being equal. The annual mean air temperature around the perimeter of the Great Lakes since 1900, summarized in Figure 4, indicate three distinct temperature regimes: a low temperature regime from 1900–1929, a

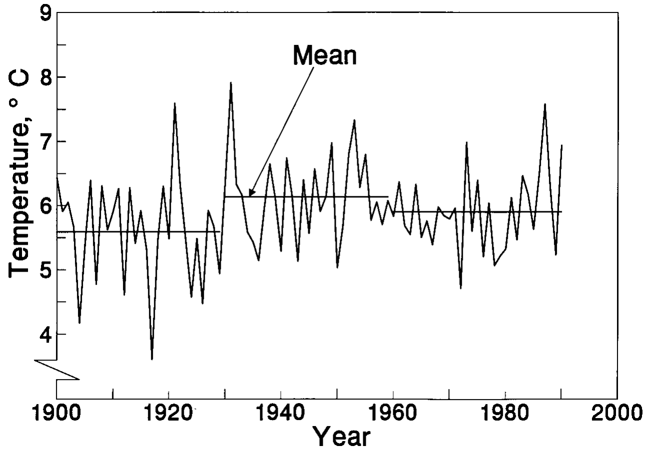


Figure 4.--Great Lakes Annual Air Temperature (1900-29, 1930-59, 1960-90).

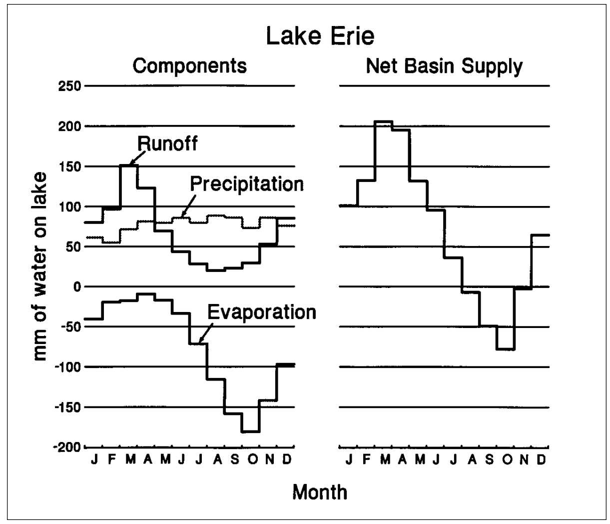


Figure 5.--Lake Erie Seasonal Net Basin Supplies.

higher temperature regime from about 1930–1959, and an additional low regime from 1960-present period. The difference between the previous and current regime is a drop of about 1°F.

The magnitude of the hydrological variables vary with season, as shown in Figure 5 for Lake Erie (Quinn, 1982; Quinn and Kelley, 1983). The monthly precipitation is fairly uniformly distributed throughout the year, while the runoff has a peak during the spring that results primarily from the spring snow melt. The runoff is at a minimum in the late summer and early fall due to large evapotranspiration from the land basin. The lake evaporation reaches a minimum during the spring and gradually increases until it reaches a maximum in the late fall or early winter. The high evaporation period is due to very cold dry air passing over warm lake surfaces. The integration of these components is depicted in the net basin supply, which consists of the precipitation plus the runoff minus the evaporation. As seen from Table 2, these three components of net basin supply are all of the same order of magnitude for each lake. Annual runoff to the lake ranges from about 62 cm (24 in) for Superior to 169 cm (67 in) for Ontario, and annual lake evaporation ranges from about 56 cm (22 in) for Superior to 90 cm (35 in) for Erie. The net basin supply is seen in Figure 5 to reach a maximum in April and a minimum in the late fall. The negative values indicate that more water is leaving the lake through evaporation than is being provided by precipitation and runoff.

2.4 Lake Level Fluctuation and Trends

There are three primary types of lake level fluctuations: long-term lake levels (represented on an annual basis), seasonal lake levels, and short-period lake level changes due to wind setup and storm surge. Annual fluctuations result in most of the variability leading to the record high and low lake levels. The annual lake levels are shown in Figure 6 from 1860 through the present to illustrate the long-term variability of the system. The record highs in 1952, 1973, and 1986 and record lows in 1935 and 1964 are readily apparent. There is an overall range of about 2 m (6 ft) in the annual levels. Of particular interest is the fall in the levels of Lakes Michigan and Huron occurring in the mid-1880’s from which the lakes never recovered. This probably results from dredging for deeper draft navigation in the St. Clair River. Other changes in the St. Clair River include sand and gravel dredging between about 1908 and 1924, a 7.6 m (25 ft) navigational project in the mid-1930’s, and an 8.2 m (27 ft) navigation project in the late 1950’s and early 1960’s. Without these changes, Lake Michigan-Huron would be approximately 0.5 m (1.5 ft) higher than it is today.

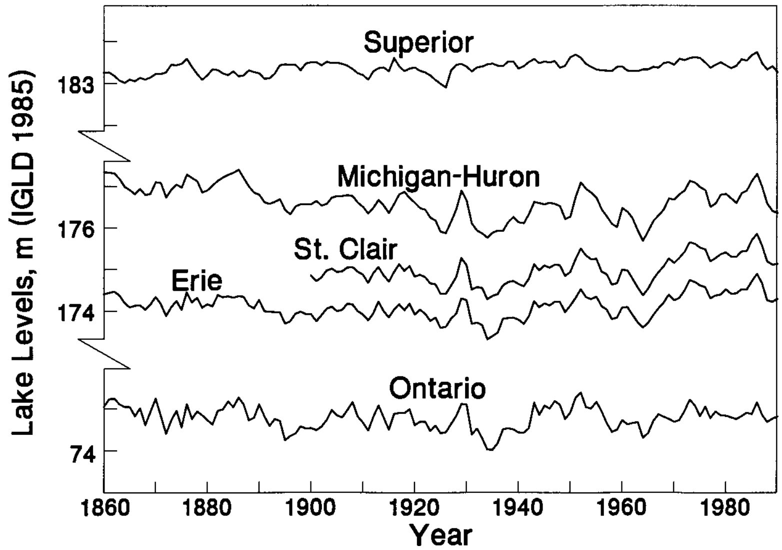


Figure 6.--Great Lakes Annual Water Levels (1900-90).

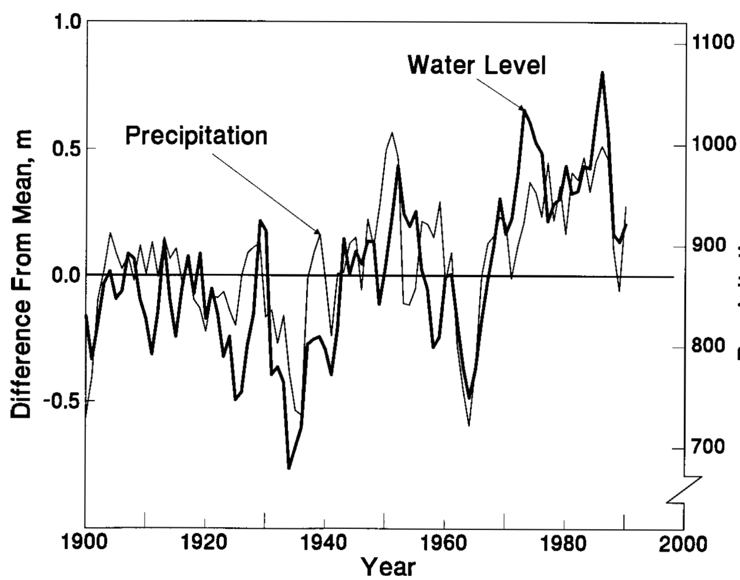


Figure 7.--Lake Erie Annual Water Levels and Precipitation.

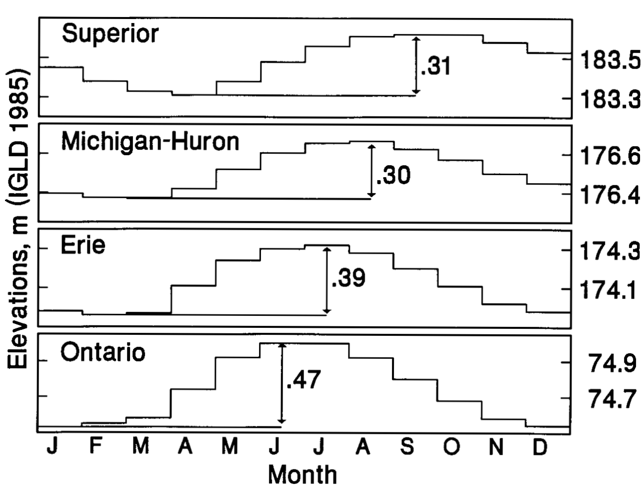


Figure 8.--Average Seasonal Great Lakes Levels (1900-90).

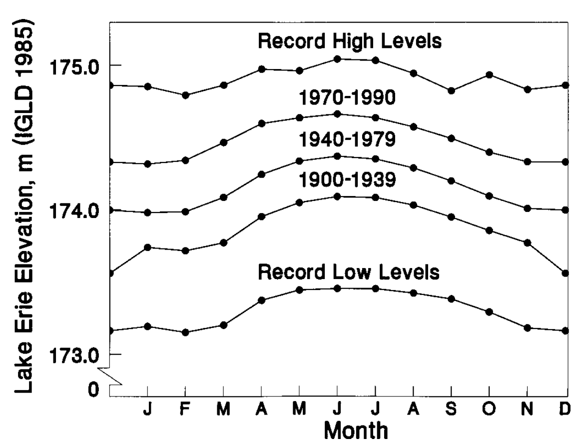


Figure 9.--Lake Erie Seasonal Water Level Comparisons.

The 3-year precipitation mean in Figure 3 correlates very well with annual lake levels as observed by superimposing the annual precipitation on the annual Lake Erie water levels in Figure 7. The precipitation tends to lead the water levels by approximately 1 year, as shown here by the 1929 highs, the 1935 lows, the 1952 highs, and the 1963 lows. In particular, the last 15 years of high precipitation resulted in very high water levels. Thus, the continuing high levels are the result of the increased precipitation regime since 1940 coupled with the lower temperature regime since 1960.

Superimposed on the annual levels are the seasonal cycles shown in Figure 8; each lake undergoes a seasonal cycle every year. The magnitude depends upon the individual water supplies. The range varies from about 30 cm (1 ft) on the upper lakes to about 40 cm (1.3 ft) or more on the lower lakes. In general, the seasonal cycles have a minimum in the winter, usually January or February. The levels then rise due to increasing water supplies from snow melt and spring precipitation until they reach a maximum in June for the smaller lakes, Erie and Ontario, and in September for Lake Superior. When the net water supplies diminish in the summer and fall, the lakes begin their seasonal decline.

The final type of fluctuation, which is common along the shallower areas of the Great Lakes, particularly Lake Erie, Saginaw Bay, and in some cases on Green Bay, is storm surges and wind set-up. Under these conditions, when the wind is blowing along the long axis of a shallow lake or bay, a rapid difference in levels can build up between one end of the lake and the other. This difference can be as large as 5 m (16 ft) for Lake Erie (storm of 2 December 1985). These storm conditions, when superimposed on high lake levels, cause most of the Great Lakes shoreline damage.

Looking in more detail at past trends in lake levels, along with more recent conditions for Lake Erie, we see a steady progression of changes in lake levels with time in Figure 9. These changes reflect the changes in precipitation, illustrated in Figure 3 and summarized in Table 3. At the bottom of Figure 9 are the record low lake levels for each month, which were set primarily in 1964. Proceeding upwards we have the 40-year average from 1900–1939. From 1940–1979, the lake is at a still higher average level. Taking the 21-year period from 1970–1990, we see that the lake level average is higher yet, followed by the record highs set in 1985. Record levels for the month were set in April and May 1985 on Lakes Michigan-Huron, St. Clair, and Erie; they were set for November 1985 through April 1986 on Lakes Erie and St. Clair. Since that time, a record drought brought water levels back to their long-term normal values in the late 1980s and early 1990s.

2.5 Diversions

It is interesting to compare the impacts of the existing diversions on lake levels in Table 4 with natural lake-level fluctuations (International Great Lakes Diversions and Consumptive Uses Study Board 1985). This enables a comparison of man's impacts with natural fluctuations. The Long Lac and Ogoki Diversions average about 160 m^3s^{-1} (5,600 ft^3s^{-1}) and raise lake levels between 6 cm (0.21 ft) and 11 cm (0.37 ft). The Chicago Diversion averages about 90 m^3s^{-1} (3,200 ft^3s^{-1}) and lowers lake levels between 2 cm (0.07 ft) and 6 cm (0.21 ft). The Welland Canal, which bypasses Niagara Falls, averages about 270 m^3s^{-1} (9,400 ft^3s^{-1}) and lowers lake levels between 2 cm (0.06 ft) and 13 cm (0.44 ft) with no effect on Lake Ontario. The combined effect on the lakes ranges from a 2 cm (0.07 ft) rise for Lake Superior to a 10 cm (0.33 ft) drop for Lake Erie. The diversion effects are therefore small in comparison with the one or more meter (several foot) variation associated with short-term storm movements, the 30–40 cm (1–1.3 ft) seasonal cycle, and the 2 m (6 ft) range of annual variations.

The small effects of the diversions along with the long response time of the system illustrate why diversions are not suitable for lake regulation. Due to the large size of the Great Lakes system, it responds very slowly to man-induced changes. This is illustrated in Figure 10 by the length of time it takes

Table 4.-- Impact of Existing Diversions on Lake Levels.

Diversion	Amount		Superior		Mich-Huron		Erie		Ontario	
	(m^3s^{-1})	(f^3s^{-1})	(cm)	(f)	(cm)	(f)	(cm)	(f)	(cm)	(f)
Ogoki-Long Lac	160	5600	+6	+0.21	+11	+0.37	+8	+0.25	+7	+0.22
Chicago	90	3200	-2	-0.07	-6	-0.21	-4	-0.14	-3	-0.10
Welland	270	9400	-2	-0.06	-5	-0.18	-13	-0.44	0	0
COMBINED	+2	+0.07	-1	-0.02	-10	-0.33	+2	+2		

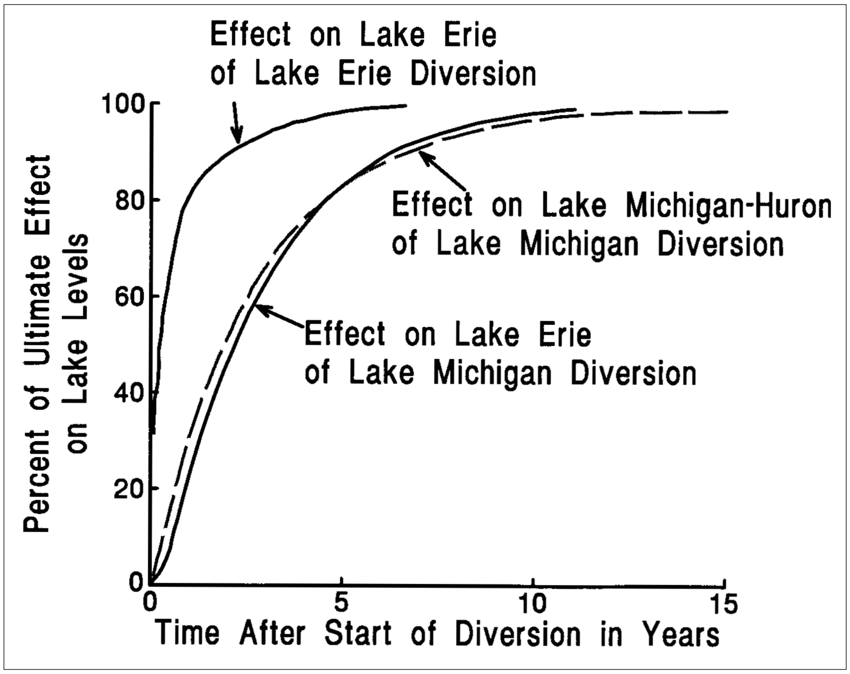


Figure 10.--Selected Great Lake Responses to Diversions.

from the start of a hypothetical diversion on Lakes Michigan and Huron (of the magnitude of the Chicago diversion) until the ultimate effect of that diversion is reached on Lakes Michigan-Huron, and Erie. It takes approximately 3–3.5 years to achieve 50% of the ultimate effect and 12–15 years to get 99% of the effect. (These results depend somewhat on the lake levels at the beginning of the diversion.) Thus, regulation by diversion would not produce changes responsive to natural fluctuations. Recent studies at GLERL indicate that an increase of 10% in the Niagara River discharge from Lake Erie (and consequent increases in Lake Erie inflow) would lower it 27 cm (10.5 in) in about 11–12 years and lower Lakes Michigan and Huron 14 cm (5–6 in) in this same period. If Lake Erie inflows were held constant (not possible at the present time), then it would take 6 months to 1 year to achieve this lowering.

Additional interbasin diversions are a highly controversial issue at the present time around the Great Lakes. Possible uses of Great Lakes water outside the basin are flow augmentation for navigation, energy uses such as synthetic fuels or pipelines, agriculture and aquifer recharge, and municipal water supplies. A small pipeline project such as the Powder River coal slurry pipeline would require 0.2 m^3s^{-1} (5.8 ft^3s^{-1}) of water and would have no measurable impact on lake levels. A synthetic fuels project, highly unlikely at this time, could require approximately 23 m^3s^{-1} (800 ft^3s^{-1}) and result in a lake level lowering of 1–2 cm (0.04–0.06 ft). A major agricultural or aquifer recharge project could require 300 m^3s^{-1} (10,000 ft^3s^{-1}) and would result in lake level decreases ranging from 12 cm (0.4 ft) on Lake Erie to 21 cm (0.7 ft) on Lake Michigan-Huron. It should be emphasized that these are hypothetical projections for illustration only.

2.6 Future

Water levels ordinarily do not change fast, as shown by the above consideration of diversions. Other studies at GLERL indicate that if normal meteorological conditions were realized ("normal" being the average conditions over 1900–69) instead of the record drought of the late 1980s, it would have taken about 6 years for Lake Michigan-Huron to return from its January 1986 level to its normal (1900–69) level. About 7 years would have been required for Lakes St. Clair and Erie to return to within 10 cm (4 in) of normal, and about 9 years would have been required for them to return to within 5 cm (2 in) of normal. Even supposing that we encountered a drought similar to the 1960–64 conditions, about 3.5 years would have been required for Lake Michigan-Huron and about 4 years would have been required for Lakes St. Clair and Erie.

A long-term perspective on Lake Michigan levels for 7,000 years was reconstructed through geologic and archaeologic evidence (Larsen, 1985) under work sponsored by the Illinois State Geological Survey. Conditions several thousand years ago were not necessarily the same as today due to isostatic rebound and uplift during the intervening time. But, in general, this provides additional perspective on possible conditions we may experience in the future. Looking at just the last 2,500 years, during which time the Great Lakes were in their current state, there were major lake level fluctuations. During most of this time the levels were much higher and more variable than they have been during the last 120 years of record. If the past is any indication, lake levels in the future could go through a considerably larger range than we have experienced lately. Indeed, the period of record, which makes up what many consider to be normal, the early 1900's through the 1960's, may represent abnormal conditions.

2.7 Summary Comments on Great Lakes Dynamics

Huge storages of water in the basins and the lakes and of energy in the lakes give the Laurentian Great Lakes their characteristic behavior. They filter the variability of the meteorological inputs and enable hydrological predictions in the face of uncertain meteorology, if the storage amounts are known. Historically, lake levels are most affected by temporal patterns of precipitation; air temperature patterns play a lesser but important role also. It is important to keep in perspective that while we have ranges in annual lake levels of 1–2 m (4–6 ft), and additional short term effects on the order of 2–3 m (7–8 ft), the effects of man on the system are relatively small, on the order of about 5 cm (0.2 ft). While the lakes are slow changing over the long term in the face of normal meteorology, past fluctuations have been very large. Future changes will depend mostly on future climate.

3. METHODOLOGY

3.1 Climate Data

The Great Lakes hydrological models used in this study (described subsequently) require daily values of precipitation, air temperature, wind speed, humidity, and cloud cover or insolation at many surface locations. The choice of climate data to use in the models is dictated by the primary goal of this work: to ascertain the sensitivity of Great Lakes net supplies to future climate changes. The focus of this study is not on the development of new techniques for creating scenarios. Nevertheless, the data must be a representation of other possible climate conditions, especially with regard to daily, seasonal, and inter-annual variability on a time scale of decades and spatial variability on the regional scale of the Great Lakes Basin (Cohen, 1990b).

In past determinations of water supply effects from climate change scenarios (Croley, 1990, 1992b, 1993a; Croley and Hartmann, 1989; Hartmann, 1990), GLERL used about 1,800 meteorological stations for over-land precipitation and air temperature and about 40 meteorological stations for over-lake air temperature, humidity, wind speed, and cloud cover (for determining insolation). Recent experience (Croley and Hartmann, 1986, 1987) suggests that 200–300 stations per lake basin for over-land meteorology and about 5–8 stations per lake for over-lake meteorology would be sufficient for operation of the large-area runoff and evaporation models at daily time intervals for studies of the type considered here. The climate scenarios, needed to achieve the objectives of this study, had to possess the following characteristics:

- Daily precipitation and daily maximum and minimum air temperature data must be at a high spatial density of approximately one station per 1,000 km^2 over the Great Lakes region of 770,000 km^2.

- Hourly observations of wind, humidity, cloud cover, and temperature must be at a spatial density of approximately one station per 20,000 km^2.

- The hourly and daily data time series must be of considerable length in order to study variability. At least 30 years of data, and ideally 40 years or more, are considered essential.

- The data must possess spatial and temporal coherence that is realistic and consistent with the physical laws governing atmospheric behavior.

- At least four widely different climate scenarios are required. These must bracket the range of values estimated under doubled CO_2 conditions derived from widely used GCMs.

3.2 Climate Transposition

It was not considered scientifically valid to attempt to create 40 years of synthetic hourly and daily weather data for 1,000 locations and across four different climatic zones the size of the Great Lakes basin. Another approach, using direct output of current GCMs, also was rejected because a single grid box represents an area of 20,000 km^2 or more, which is far too crude of a spacing for the purposes of this study. At this time, the results of GCMs are not considered sufficiently reliable for determining regional scale change (Dickinson, 1987), short-term variability, and extremes. A third possibility is the use of regional climate models that are embedded in GCMs. However, the massive computer requirements of this approach and their development have thus far limited the length of time series to 1–2 years in length (Bates et al., 1994). Other approaches were needed.

The technique, selected to define weather conditions for this study, is climate transposition. GCMs predict that continuing increases in atmospheric trace gas concentrations will result in warmer conditions, comparable to climates south of the Great Lakes. Some GCMs also predict drier conditions, comparable to climates to the west of the Great Lakes. Therefore, the future climate of the Great Lakes may be similar (at least in terms of annual means and other very general features) to the present climate of regions to the south and west of the Great Lakes. We relocated the Great Lakes basin to four other climatic zones in the western and southern United States to sample climatic differences in fluctuations over time. Robinson and Finkelstein (1989) assessed means of developing climate scenarios for impact studies like this, and they suggested climate analogues as one useful empirical method for developing climate scenarios. Climate analogues use historical data to represent a changed and often extreme climate condition perceived to exist at some future time, and one form of climate analogue is labeled as “spatial transfers”

(Changnon, 1991). The underlying rationale for this approach is the expectation that future changes in basin climate conditions may approximate latitudinal and/or longitudinal shifts. For example, current temperatures in Dayton, Ohio, may become the norm in Toronto at some future time. Thus, we could use the current values for Dayton to perform impacts assessments for Toronto.

The major advantage of this approach is that the transposed data represent an actual climate time series (Robinson and Finkelstein, 1989). The data exist and, thus, conceivably could happen again somewhere else in the future. All key features of the climate are realistic, including the temporal variability and the frequency and magnitude of extremes. Further, the spatial relationships are obviously realistic. One could question whether the exact climate conditions currently existing in the southwestern U.S. could ever exist in the Great Lakes region, but for the sake of achieving the desired sensitivity analysis from drastically different climates, we have assumed this possibility exists.

This technique takes advantage of the existing detailed climatic data record in the U.S. and Canada. In this study we used data for the period 1949–1990. Data in digital form for both hourly and daily weather elements are available for a dense network of observing sites; see Figure 11. By using the existing long-term observations, we are able to create a wide range of surface climate conditions. Also, by carefully choosing our latitudinal and longitudinal shifts, we are able to realize climates that match closely with temperature and precipitation outcomes predicted by GCMs or that are more extreme. Finally, our past experience with those attempting to assess the impact of climate change reveals that use of “actual” historical data leads to improved credence in understanding and accepting outcomes.

There are approximately 2,000 climate (temperature and precipitation) stations in the Great Lakes basin; see Figure 11. For a 40-year period of daily measurements, this corresponds to about 30 million values for each climate element. The development of such detailed scenarios, by means other than transposing climates by relocating stations, faces a monumental problem in ensuring that such a large 40-year data set properly incorporates physically plausible characteristics for temporal and spatial variability.

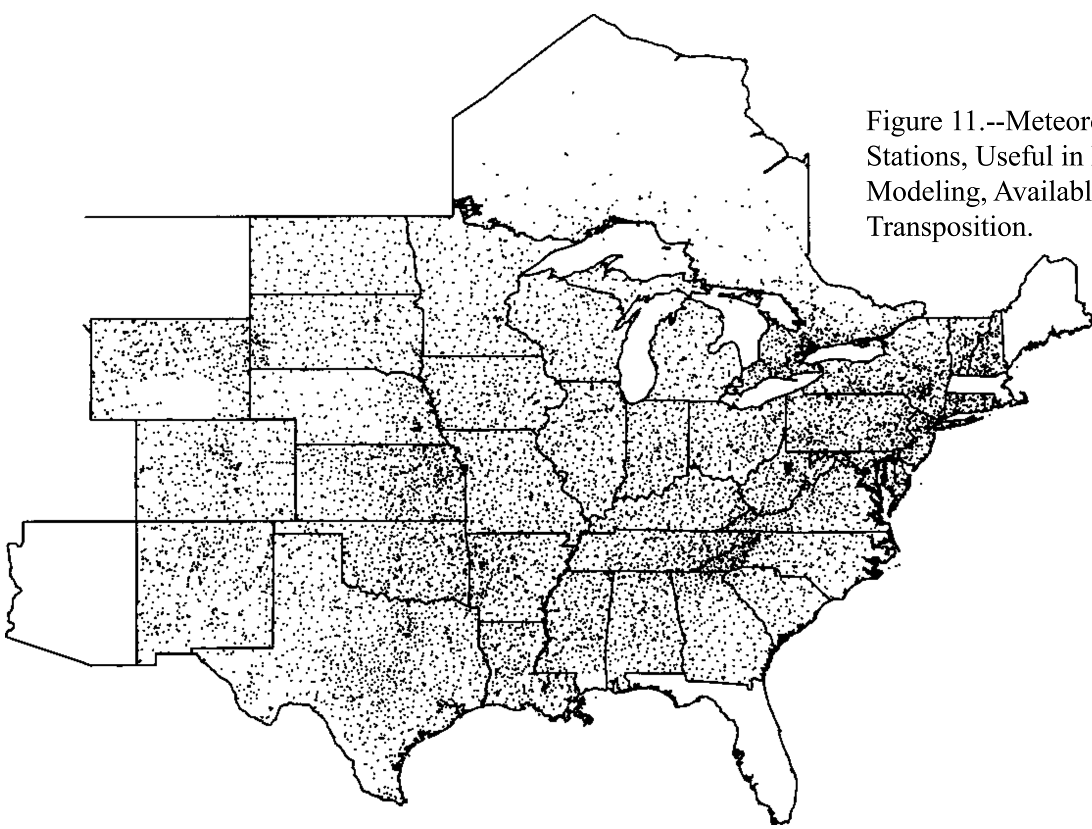


Figure 11.--Meteorological Stations, Useful in Runoff Modeling, Available for Transposition.

Daily maximum and minimum temperatures, precipitation, and snowfall were obtained for the 42-year period of 1949–1990 from the dense array of stations in the National Weather Service's cooperative observer network for each region; see Figure 11. Out of this dense array of stations, a subset have daily records of wind speed, humidity, and cloud cover, and are generally located at the National Weather Service offices and other airport observing stations; see Figure 12. There are about 40 of these stations for each climate scenario.

The published results of GCMs under doubled CO_2 conditions were used to determine a likely range for changes in mean annual temperature and precipitation in the Great Lakes basin for a doubling of atmospheric trace gas concentrations. We considered four separate climate regimes by moving the Great Lakes basin to the south and west of its current position. In all of these, the relative spatial relationships of the geography of the Great Lakes were preserved with the outline of the basin laid over the existing climate network. Figure 13 shows the location of the Great Lakes basin for each scenario. The first two climate regimes correspond roughly to the upper range of GCM predictions for temperature for the Great Lakes basin (IPCC, 1992). Scenario 1 (warm and dry) corresponds to warmer temperatures and mixed precipitation changes. It represents a movement of the Great Lakes basin 6ºS and 10ºW. In this scenario, the climate of the Great Lakes is similar to the present climate of the central Great Plains, middle Mississippi, and lower Ohio valleys. Scenario 2 (warm and wet) is a simple shift 6ºS and corresponds to warmer temperatures and increases in precipitation amounts over the entire basin.

The next two climate regimes went beyond the range of current GCM predictions for a doubling of atmospheric trace gas concentrations. They were chosen to determine how the Great Lakes would respond to a major climatic shock. Scenario 3 (very warm and dry) corresponds to very high temperatures and mixed precipitation changes. It is a shift 10ºS and 11ºW and, while generally wetter over much of the basin, is drier in the western part of the basin. Scenario 4 (very warm and wet) corresponds to very high temperatures and large increases in precipitation over the entire basin. It is a shift 10ºS and 5ºW.

The stations from earlier studies (Croley, 1990, 1992b, 1993a; Croley and Hartmann, 1986, 1987, 1989; Hartmann, 1990) were augmented here to extend their data period through 1990; their locations are depicted in Figures 14 (precipitation and air temperature) and 15 (air temperature, humidity, wind speed, and cloud cover). The earlier data reduction to determine areal Thiessen-averaged meteorological time series over each of the 121 sub-basins and the 7 lake surfaces was enormous (Croley and Hartmann, 1985b), but the software for this was developed at that time. Now, improved computers are available that allow the re-reduction of all data in a timely fashion with this software.

By taking the stations depicted in Figures 11 and 12 and translating their locations in accordance with Figure 13, station networks were created for each of the scenarios already identified. These are depicted in Figure 16 for stations reporting air temperature and precipitation an in Figure 17 for stations reporting air temperature, humidity, wind speed, and cloud cover. We repeated data preparation and data reduction computations for the Thiessen-weighting over all 121 subbasins for daily minimum and maximum air temperature and precipitation networks (Figures 14 and 16) and over the 7 lake surfaces for daily air temperature, humidity, wind speed, and cloud cover networks (Figures 15 and 17). These data sets are available at GLERL.

3.3 Scenario Climatology

The use of existing climate data from nearby regions to create different climatic regions puts certain constraints on the scenario climatology. A brief description of the spatial variations of climate conditions across the four regions is offered to highlight these constraints.

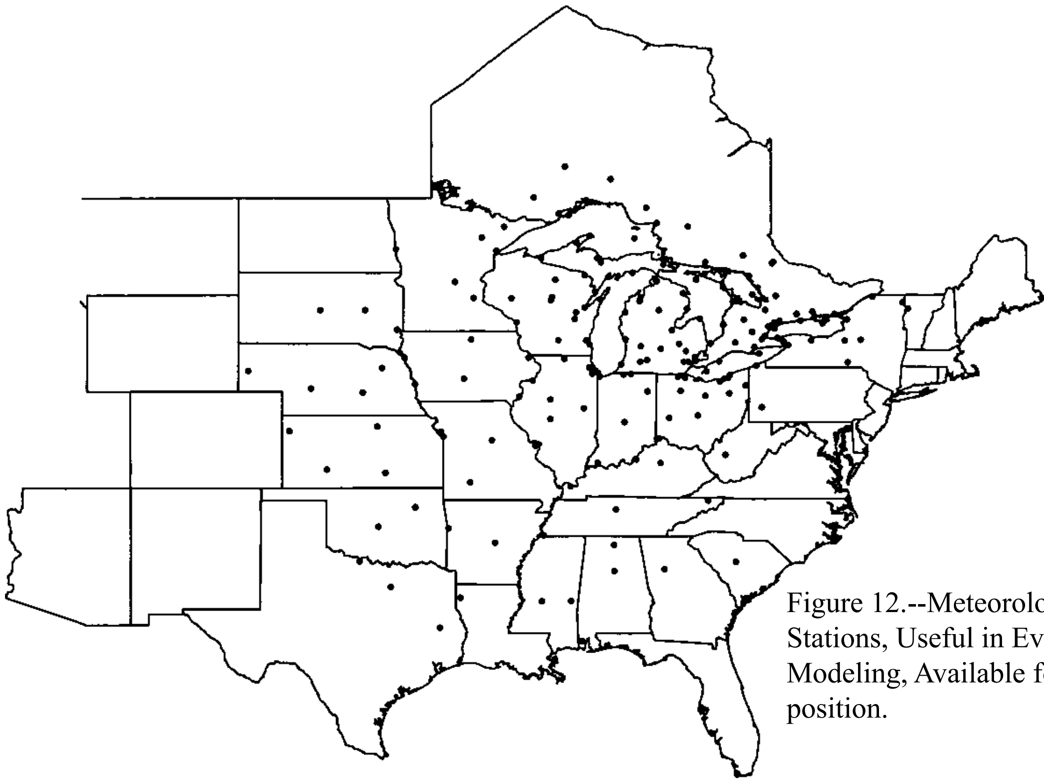


Figure 12.--Meteorological Stations, Useful in Evaporation Modeling, Available for Transposition.

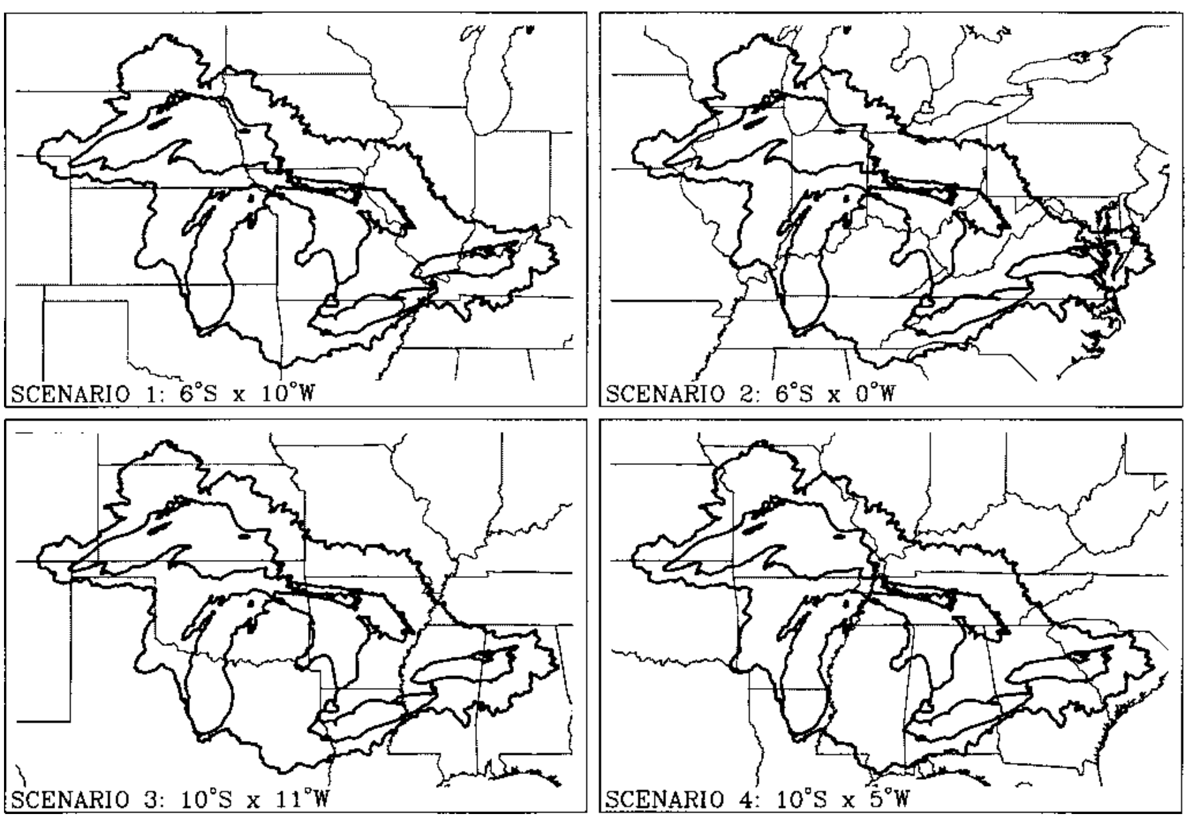


Figure 13.--Transposed Climate Shifts to the Great Lakes Basin.

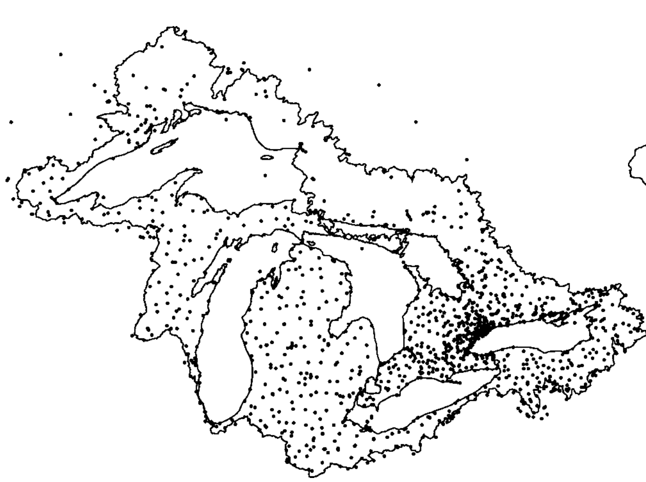


Figure 14.--Base Case Temperature and Precipitation Stations.

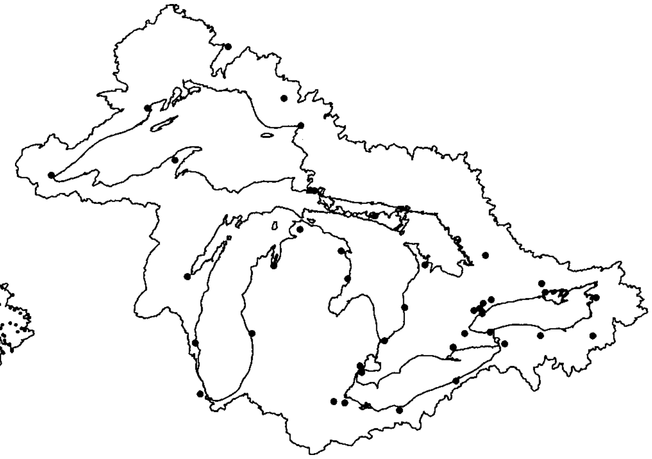


Figure 15.--Base Case Temperature, Humidity, Wind Speed, and Cloud Cover Stations.

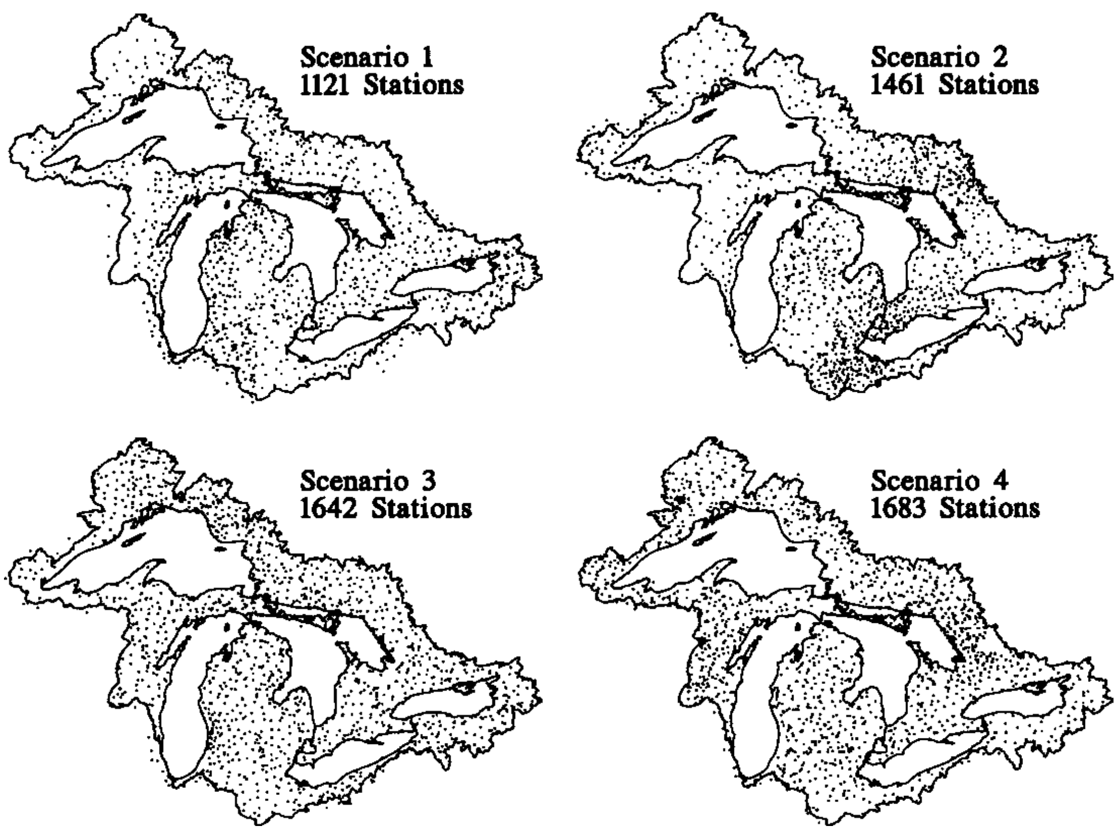


Figure 16.--Temperature and Precipitation Stations.

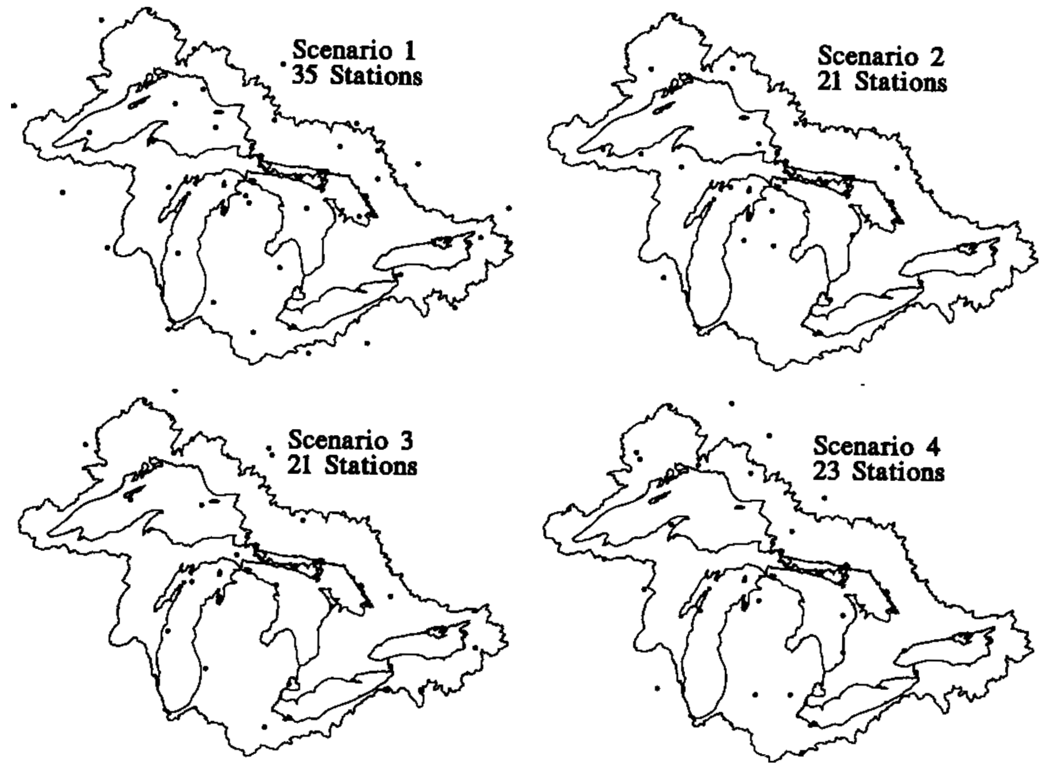


Figure 17.--Temperature, Humidity, Wind Speed, and Cloud Cover Stations.

The spatial pattern of temperature everywhere east of the Rocky Mountains is characterized by a predominantly north-south gradient. As a result, temperature changes produced by transposition of the Great Lakes maintain this north-south gradient. The summertime temperature gradients are somewhat weaker than the gradients during the rest of the year. Therefore, the scenario temperature changes are somewhat smaller in the summer than in the other seasons. Basically all four scenarios yield north-south gradients similar to what exists now in the Great Lakes Basin. Scenarios 1 and 2 are warmer than current conditions by roughly 4 to 7ºC. Scenarios 3 and 4 are warmer than current conditions by 9 to 10ºC.

Precipitation patterns for the four scenarios are more complex. Annual total precipitation exhibits a north-south gradient in the eastern part of the U.S., similar to the gradient in temperature. However, the direction of the gradient gradually shifts westward becoming nearly east-west in the High Plains. There are also pronounced seasonal differences in the gradients. The gradients are large during the colder half-year (November–April). During the warmer half-year, the gradients are smaller; in fact, there is little spatial variation during mid-summer. As a result of the above patterns, and the "movement" of the lakes southward in all four scenarios, the amount of precipitation is increased in most scenarios for most basins. Scenario 4 has the largest precipitation increase of around 50%. However, in scenarios 1 and 3, the westward shift of the Great Lakes is sufficient such that the western part of the Great Lakes basin is positioned in the region of the east-west precipitation gradient. Thus, these scenarios produce slight decreases in precipitation over the western lakes relative to current values. Scenarios 1 and 3 for Lake Superior show a roughly 20% decrease in annual precipitation. These precipitation patterns place a constraint on the range of precipitation in the scenarios considered in this study. Because the movement of the basin in these scenarios is generally toward regions of greater precipitation, none of the scenarios are warmer and drier than current conditions over the whole basin.

Total annual snowfall exhibits a predominantly north-south gradient except for the snowbelt areas on the lee sides of the Great Lakes. Thus, the transposition of the basins to the south and west results in a significant decrease in snowfall for all scenarios because of the warmer temperatures to the south.

Evaporation is a function of several climatic variables including temperature, atmospheric water vapor content, radiative energy, and wind speed. The spatial pattern of dewpoint temperature (one measure of atmospheric water vapor content) is similar to that of precipitation. In the eastern U.S., the gradient is north-south. In the Great Plains this gradient changes to predominantly east-west. When combined with the gradient of temperature, the pattern of relative humidity exhibits weak gradients over the eastern U.S. However, a westward decrease in relative humidity occurs throughout the Great Plains. Thus, for scenarios 2 and 4, with little or no westward shift, there is little change in relative humidity in the scenarios. However, for scenarios 1 and 3, there are decreases in relative humidity over the western part of the basin. However, even in scenarios 2 and 4, the average water vapor pressure deficit increases everywhere in all basins, even where relative humidity does not increase, due to the non-linear (Clausius-Capeyron) relationship between temperature and saturation water vapor pressure.

The Great Lakes are located in a region of relatively high cloud cover and low solar radiation. Cloud cover generally decreases to the south and west of the Great Lakes. Thus, there is decreased cloud cover in all scenarios over all lake basins.

Wind speed gradients are weak in much of the eastern U.S. The wind speed does increase somewhat toward the west within the Great Plains. Thus, in most scenarios for most basins, the wind speed does not change significantly. The exception is the western part of the basin in scenarios 1 and 3 where slight increases in wind speed are experienced.

When the patterns for each of the evaporation-related variables are combined, the spatial pattern of potential evaporation is characterized by a gradient with increasing values from northeast to southwest. Particularly high values are seen in the southern Great Plains.

Although higher dewpoint temperatures are experienced in all scenarios, the Clausius-Clapeyron relationship provides a powerful constraint on the direction of evaporation changes. It appears highly likely that in any warmer scenario, the water vapor pressure deficit is likely to increase because of this relationship. Likewise, the current climate of the Great Lakes is characterized by high cloudiness and low radiation. Although cloudiness could increase, it would appear to be more likely that cloudiness would either remain the same or decrease. The third factor, wind speed, is an unknown in future climates. There are no fundamental physical factors that might provide a preference for increases or decreases.

3.4 Water Impact Assessment

GLERL constructed a master computer procedure to integrate their Large Basin Runoff Model and Lake Evaporation and Thermodynamics Model (described subsequently) with over-lake precipitation estimates, for all lakes, to provide a net water supply model for the entire Great Lakes system. GLERL estimated the Great Lakes hydrology of each transposed climate as in earlier studies, by applying the system of hydrological models to the transposed climate 42-year daily time series directly and comparing outputs for each transposed climate to a base case derived with the models from (the untransposed) Great Lakes historical meteorological data. This approach allows preservation of reasonable spatial and temporal variations in meteorology and preserves the interdependencies that exist between the various meteorological variables. It also allows the use of appropriate spatial and temporal scales, better matching the models than do GCM outputs.

3.5 Consideration of Lake Effect

3.5.1 Introduction

One aspect of the study included the assessment of the lake effects on various weather conditions that occur over the Great Lakes region. The existing amount of change in the climatic conditions over adjacent land areas around the lakes was determined and used to estimate the extent of altered weather conditions in other climatic zones. Our initial test involved use of scenario 3 and its weather conditions, tested with and without lake effects, as input to the hydrologic model to ascertain the degree of difference in basin hydrologic components the changes made. If the estimated lake effects were found to create major changes in the hydrologic components of the lakes, we planned to calculate and use lake effects with the three other scenarios.

Past Water Survey studies of the Lake Michigan basin used a climatological technique for defining the extent of the lake effect on both sides of the lake on monthly, seasonal, and annual precipitation, temperatures, and other weather conditions (Changnon, 1968). A similar technique was used in this study to derive measures of lake effects around each lake for the four major seasons (winter, spring, summer, and fall), and for seven weather parameters in each season including daily precipitation, maximum daily air temperature, minimum daily air temperature, average daily air temperature, daily cloud cover, daily wind speed, and daily water vapor pressure. A major investigation was conducted to measure average and extreme lake effects in all seasons, and the findings are the subject of a separate report (Scott and Huff, 1995).

Numerous investigations have concerned lake effects on climatic conditions in the Great Lakes basin. For example, Day (1926) investigated precipitation in the drainage area of the Great Lakes, and Horton and Grunsky (1927) made a study of the hydrology of the basin including estimated effects on precipitation and temperature. Peterssen and Calabrese (1959) evaluated weather influences related to the warming of air by the Great Lakes. Blust and DeCooke (1960) made comparisons of precipitation on islands in Lake Michigan with precipitation on the perimeter of the lake. Changnon (1968) made an intensive study of the precipitation climatology of Lake Michigan, and Lyons (1966) assessed lake effects on storms and convective activity. Jones and Meredith (1972) published information on the Great Lakes hydrology by months. Phillips and McCulloch (1972) and Saulesleja (1986) provided additional information on the climate of the Great Lakes basin. Braham and Dungey (1995) studied lake effects on winter precipitation over Lake Michigan. The use of atmospheric models to simulate and calculate the effects of the Great Lakes on regional climate conditions has begun, but as yet is limited to examining short periods of time. Hence, modeling of effects in the four climate zones was discarded as a feasible approach. An empirical three-step climatological technique was developed to define lake effects.

3.5.2 Method of Analysis

It was assumed that any significant lake effect would occur within 80 km of the lake shore. This is considered a conservative assumption based on previous studies, available long-term climate data, and consideration of regional climatic conditions.

In the warm season when storm movements across the Great Lakes region are larger from the northwest and southwest quadrants, the 80-km band may be an over-compensation in the predominantly upwind directions. For example, significant precipitation-producing storms rarely move from east to west across Lake Michigan. Any lake effect on precipitation generated by lake breeze activity would be limited to 30–35 km from the western shore of the lake. However, the 80-km lake effect band was used in all lake effect analyses to ensure that all areas of potential lake effect were included.

For each meteorological element evaluated, three sets of maps were generated. The first map included all observations. This map was used primarily to establish the spatial distribution pattern with the Great Lakes basin and the surrounding areas. The spatial pattern of any condition in the basin incorporates both lake-induced changes and those produced by the broad-scale climatology of the region.

A relatively large region surrounding the basin was used in the analyses to provide an adequate measure of the non-lake pattern, the second map. Seasonal average patterns of each condition were constructed based on data for the period 1951–1980. This second map was generated by eliminating data from all stations in the 80-km lake effect band. The pattern existing in the no-effect region surrounding the basin was used as the primary guide in establishing the climatological pattern assumed to exist if no lake effects were present. Digital files were created for both patterns and for the seven daily parameters: precipitation, maximum temperature, minimum temperature, average temperature, cloud cover, wind speed, and water vapor pressure.

A data plotting routine was developed to map and display the data. An interpretative analysis was then performed on the plotted data to create the final patterns. Hand-analyzed charts offered better presentations, compared to machine-analyzed maps, due to the small spatial differences in some of the parameters between "lake" and "no lake" analyses. In addition, this approach allowed for physical and climatological insights in the placement of isopleths which available objective techniques could not be programmed to define.

The no-effect and all-data maps were compared, and the amount and placement of the lake effect was derived by determining the differences between the two sets of values (map 1 and map 2). These differences were used to digitally generate a third map of lake effects based on the computed differences across the basin. After these lake-effect patterns had been determined for each condition, the analyzed data were digitized on grid square as input to the Great Lakes basin hydrology models.

3.5.3 Examples of Lake Effect Patterns

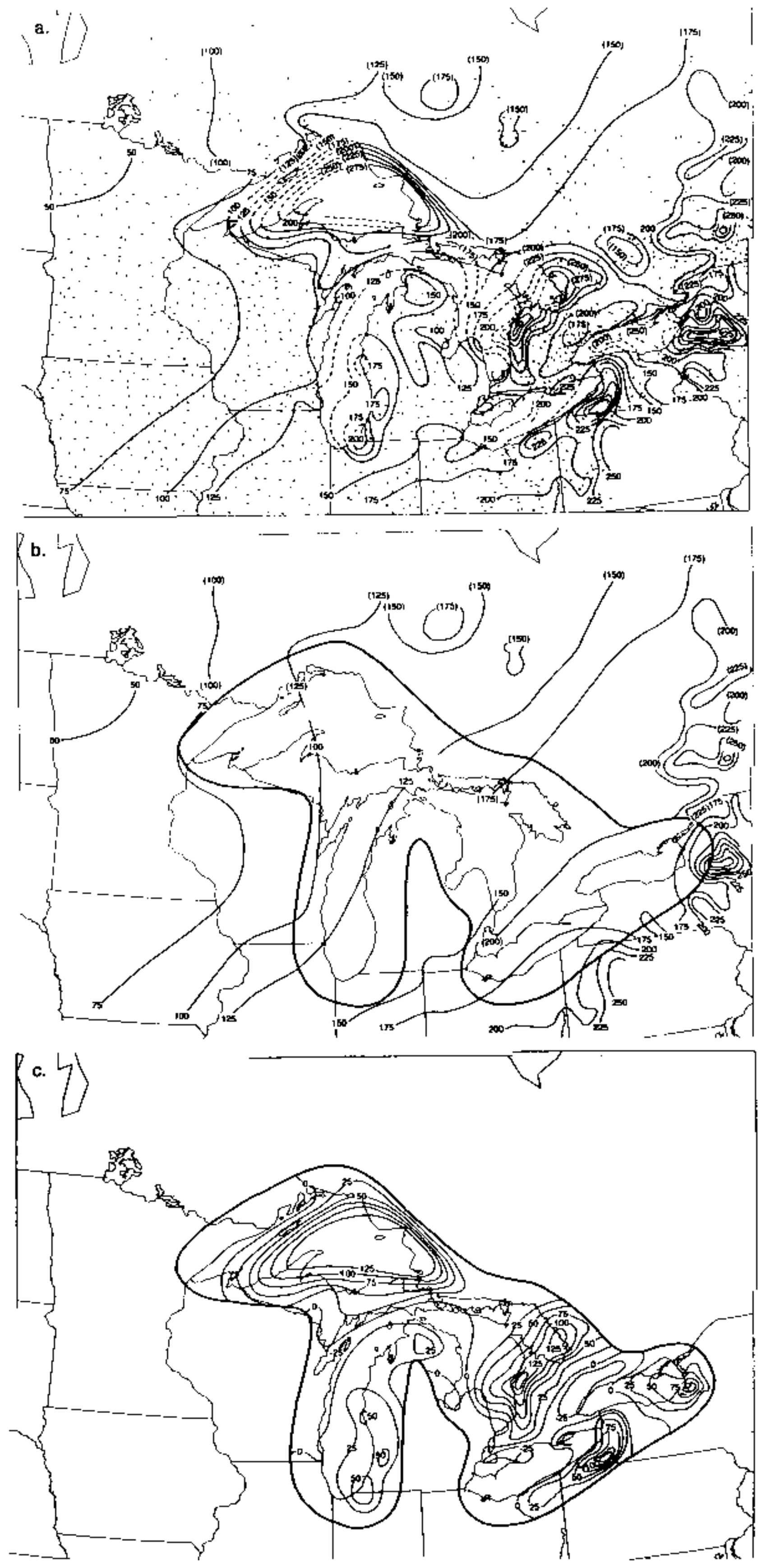
Figure 18 shows the three maps generated for winter precipitation across the Great Lakes basin and surrounding region. As noted, a large area surrounding the Great Lakes basin was included since it was essential for estimating the no-lake effect patterns over and beyond the basin.

Figure 18a shows the climatic pattern for winter that incorporates both climatic variations and lake-induced effects. Major highs in the pattern are indicated over the eastern part of Lake Superior, eastern Lake Huron, south of Lake Erie, and east of Lake Ontario. A less-pronounced high is located over eastern Lake Michigan.

Figure 18b shows the spatial pattern developed when the lake-induced effect values were eliminated. All of the highs (Figure 18a) have become less pronounced over and downwind of the lakes, except for a small area just east of Lake Ontario where topographic effects are important.

Figure 18c is the pattern of differences between the maps of Figures 18a and 18b. A major increase in precipitation induced by the lake effect is centered in the southern portion of the Lake Superior basin and extends east of the lake. The lake-induced increase, which exceeds 75 mm over the southern part of the lake, corresponds to an increase of approximately 50%. The high centers of 75 mm in the Lake Huron area correspond to an increase of approximately 35%. The highs at the eastern ends of Erie and Ontario indicate increases of approximately 30% and 20%, respectively. The 25–30 mm highs along the eastern shore of Lake Michigan correspond to increases of 20%–30%.

Figure 18.--Great Lakes Basin Climatic Lake Effect Winter Precipitation (mm) Patterns (**a**: both climatic variations and lake induced effects; **b**: climatic variations without lake induced effects; **c**: lake induced effects as the difference between a and b).



Except for Lake Superior, the lake-induced increases in winter are in the 20%–35% range, similar to those obtained in the Lake Michigan basin by Gatz and Changnon (1968). The greater increases in Lake Superior may be related to a more frequent exposure to arctic and polar front passages in the cold season, combined with a larger surface area, resulting in a longer fetch of air flow over the lakes. Braham and Dungey (1995) made quantitative estimates of land-induced snowfall over Lake Michigan during 72 winters, 1910–1918. By using snowfall measurements far to the west and far to the east of the lake, they interpolated snowfall directly adjacent to the lake under the assumption that no lake was present. When compared to actual observations around the lake, this procedure yielded an estimated 10% increase in snowfall along the Wisconsin shore and a greater than 60% increase along the Michigan shore that they attributed to lake effects.

Figure 19 presents the three maps for the average winter minimum temperatures. A sizable influence of the lakes on temperatures is noted (Figure 19c) with changes of up to 8°C over Lake Superior. These values were those used to modify the over-land winter minimum air temperatures existing within 80 km of the lakes, as positioned in scenario 3. Results from the hydrologic models for scenario 3, both with and without the calculated lake effects, showed little significant difference in basin hydrologic conditions (described later). Hence lake effects were not applied to the three other climate scenarios investigated here.

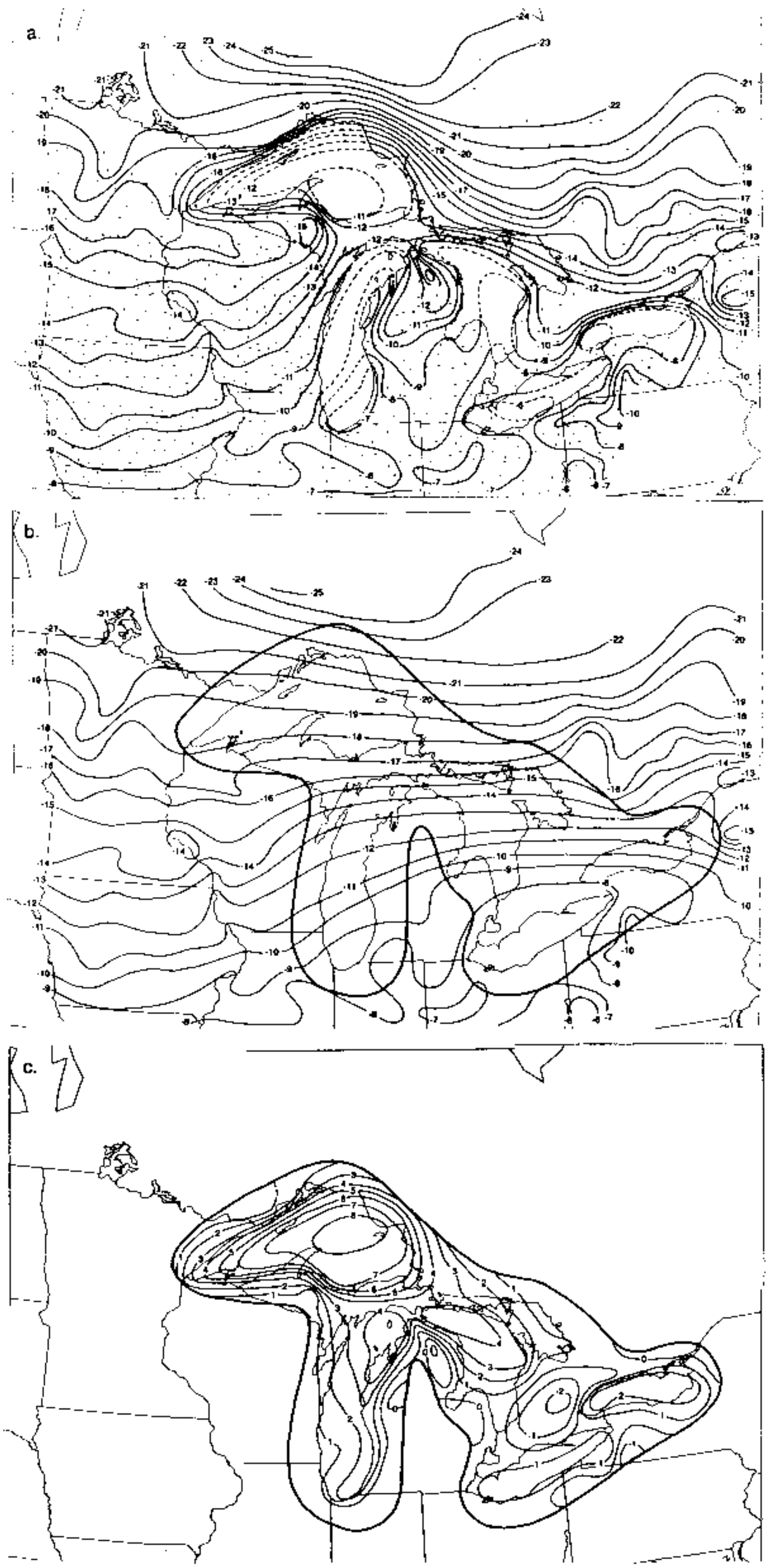
4. GREAT LAKES PHYSICAL PROCESS MODELS

The GLERL developed, calibrated, and verified conceptual model-based techniques for simulating hydrological processes in the Laurentian Great Lakes (including Georgian Bay and Lake St. Clair, both as separate entities). GLERL integrated the models into a system to estimate lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1993a,b; Croley and Hartmann, 1987, 1989; Croley and Lee, 1993; Hartmann, 1990). These include models for rainfall-runoff [121 daily watershed models (Croley, 1982, 1983a,b; Croley and Hartmann, 1984)], over-lake precipitation (a daily estimation model), one-dimensional (depth) lake thermodynamics [seven daily models for lake surface flux, thermal structure, and heat storage (Croley, 1989a,b, 1992a; Croley and Assel, 1993)], channel routing [four daily models for connecting channel flow and level, outlet works, and lake levels (Hartmann, 1987, 1988; Quinn, 1978)], lake regulation [a monthly plan balancing Lakes Superior, Michigan, and Huron (International Lake Superior Board of Control, 1981, 1982) and a quarter-monthly plan balancing Lake Ontario and the St. Lawrence Seaway (International St. Lawrence River Board of Control, 1963)], and diversions and consumptions (International Great Lakes Diversions and Consumptive Uses Study Board, 1981). The hydrological models for basin runoff, over-lake precipitation, and lake thermodynamics are described in this chapter; results from these models are presented in Chapter 5. The models for channel routing, lake regulation, and diversions and consumptions are presented with their results in Chapter 6.

4.1 Runoff Modeling

The GLERL Large Basin Runoff Model (LBRM) is an interdependent tank-cascade model that employs analytic solutions of climatologic considerations relevant for large watersheds (Croley, 1983a,b). It consists of moisture storages arranged as a serial and parallel cascade of “tanks” to coincide with the perceived basin storage structure of Figure 2. Water enters the upper soil zone tank and flows from the upper to the lower soil zone and surface storage tanks, from the lower to the groundwater and surface tanks, from the groundwater to the surface tank, and from the surface tank out of the watershed.

Figure 19.--Great Lakes Basin Climatic Lake Effect Average Winter Minimum Air Temperature (°C) Patterns (**a**: both climatic variations and lake induced effects; **b**: climatic variations without lake induced effects; **c**: lake induced effects as the difference between a and b).



4.1.1 Snowmelt and Infiltration

Water enters the snowpack, if present, and some then infiltrates into the upper soil zone based on degree-day determinations of snowmelt and net supply:

$$
\begin{aligned}
m_p &= 0, & T_a &\le 0, \\
&= a_s\, DD, & T_a &> 0
\end{aligned}
\tag{1}
$$

where mp = daily potential snowmelt rate ($m^3\ d^{-1}$); as = proportionality constant for snowmelt per degree-day ($m^3\ °C^{-1}\ d^{-1}$]); T_a = air temperature, estimated as the average of the daily maximum and minimum air temperatures (°C); and *DD* = degree-days per day ($°C\ d\ d^{-1}$), computed as the integral of air temperature with time over those portions of the day when it is positive. Since the fluctuation of air temperature during the diurnal cycle is unknown, a triangular distribution is assumed (to approximate an expected sinusoidal variation) for ease of computation. The resulting expression for degree-days is:

$$
\begin{aligned}
DD &= 0, & T_{\max} &\le 0, \\
&= T_{\max}^2 \big/ \left(T_{\max} - T_{\min}\right) \big/ 2, & T_{\min} &< 0 < T_{\max}, \\
&= T_a, & 0 &\le T_{\min}
\end{aligned}
\tag{2}
$$

where T_{max} = maximum daily air temperature (°C), and T_{min} = minimum daily air temperature (°C). Actual snowmelt depends upon the snowpack:

$$
\begin{aligned}
m &= m_p, & m_p d &\le SNW_0, \\
&= SNW_0 \big/ d, & m_p d &> SNW_0
\end{aligned}
\tag{3}
$$

where *m* = daily snowmelt rate ($m^3\ d^{-1}$), and SNW_0 = water content of the snowpack at the beginning of the day (m^3). Snowpack mass balance and water supply to the watershed surface can now be determined:

$$
\begin{aligned}
\frac{\partial}{\partial} SNW &= p, & T_a &\le 0, \\
&= -m, & T_a &\le 0
\end{aligned}
\tag{4}
$$

$$
\begin{aligned}
ns &= 0, & T_a &\le 0, \\
&= p + m, & T_a &> 0
\end{aligned}
\tag{5}
$$

where *t* = time, *p* = precipitation rate ($m^3\ d^{-1}$), and *ns* = daily net supply rate to watershed surface ($m^3\ d^{-1}$).

4.1.2 Heat Available for Evapotranspiration

The heat available for evapotranspiration is estimated empirically from the average air temperature as follows:

$$\Psi = K \exp\left(T_a / T_b\right) \tag{6}$$

where Ψ = total heat available for evapotranspiration during the day (cal), K = units and proportionality constant (cal), and T_b = a base scaling temperature (°C). The constant, K, is determinable from the following boundary constraint on the long-term heat balance:

$$\sum \Psi_i = \sum \left(rr_i - m_i \, \rho_w \, \gamma_f \right) d \tag{7}$$

where rr = daily solar insolation at the watershed surface (cal d^{-1}), ρ_w = density of water (= 106 gm m^{-3}), γ_f = latent heat of fusion (= 79.7 cal gm^{-1}), and the subscript, i, refers to daily values. Equation 7 conserves energy in that all absorbed insolation not used for snowmelt appears sooner or later as other components of the heat balance that determine Ψ. Daily insolation is taken as:

$$rr = 10000 \, A_b \, \tau \left(b_1 + b_2 \, X\right) \tag{8}$$

where A_b = area of the watershed (m^2), τ = daily extra-terrestrial solar radiation (langleys d^{-1}), b_1 and b_2 = empirical constants, and X = daily ratio of hours of bright sunshine to maximum possible hours of bright sunshine, estimated from daily air temperatures:

$$X = MIN\left[\left(T_{\max} - T_{\min}\right)/15, 1.0\right] \tag{9}$$

While calculations for ns and Ψ are performed on a daily basis, the mass balance computations (following) are performed on an n-day basis (n = 1, 7, and 28–31 are typical). The net supply and energy available for evapotranspiration are summed over the n-day periods prior to the mass balance:

$$ns_a = \frac{1}{n} \sum_{i=1}^{n} ns_i \tag{10}$$

$$\Psi_a = \sum_{i=1}^{n} \Psi_i \tag{11}$$

where ns_a = average net supply rate for n days (m^3 d^{-1}), Ψ_a = accumulated energy available for evapotranspiration over n days (cal), and n = number of days in the mass balance computation periods. The subscripts refer to daily values within the computation period.

4.1.3 Infiltration

Infiltration is taken as instantaneously proportional to the supply rate and to the areal extent of the unsaturated portion of the upper soil zone (partial-area infiltration concept).

$$f = ns_a \left(USZC - USZM\right) / USZC \tag{12}$$

where f = infiltration rate (m^3 d^{-1}), *USZC* = capacity of the upper soil zone (m^3), and *USZM* = content of upper soil zone (m^3). The difference between the net supply rate and infiltration is surface runoff in Figure 2.

4.1.4 Evapotranspiration

All incoming heat is considered here to be released by the watershed surface by ignoring heat storage and the energy advected by evaporation. The release consists of short-wave reflection, atmospheric heating (composed of net long wave exchange, sensible heat exchange, net atmospheric advection, and net hydrospheric advection), and evaporation-evapotranspiration (referred to herein jointly as evapotranspiration). The total heat available for evapotranspiration over a day is composed of the heat actually used for evapotranspiration and that used for atmospheric heating. At any instant, the rate of evaporation or evapotranspiration, e, is proportional to the amount of water available, Z (reflecting both areal coverage and extent of supply), and to the rate of nonlatent heat released to the atmosphere, $\partial H_s/\partial t$ (atmospheric heating):

$$e = \beta Z e_p, \qquad e_p = \frac{\partial H_s}{\partial t} \Big/ (\rho_w \gamma_v) \tag{13}$$

where e = evaporation or evapotranspiration rate (m^3 d^{-1}), β = partial linear reservoir coefficient (m^{-3}), Z = volume of water in storage (m^3), e_p = rate of evaporation or evapotranspiration, respectively (m^3 d^{-1}), still possible and γ_v = latent heat of vaporization (596–0.52 T_a cal gm^{-1}). This agrees with existing climatologic and hydrological concepts for evapotranspiration opportunity.

Over large areas, climatic observations suggest that actual evapotranspiration affects temperatures, wind speeds, humidities, and so forth, and hence it affects the potential evapotranspiration (evapotranspiration opportunity or capacity); the heat used for evapotranspiration reduces the opportunity for additional evapotranspiration (complementary evapotranspiration and evapotranspiration opportunity concept). This concept is modified here by considering that, for short time periods, the total amount of energy available for evapotranspiration, Ψ, during the time period is split into that used for evapotranspiration and that used for atmospheric heating. From (13), for the daily time period,

$$\Psi = H_s + \rho_w \gamma_v \left(E_u + E_l + E_g + E_s \right) \tag{14}$$

where H_s = nonlatent heat released to the atmosphere during the day (cal) and E_u, E_l, E_g, and E_s = evaporation or evapotranspiration from the upper soil zone, lower soil zone, groundwater, and surface storages (m^3), respectively. The evaporation from stream channels and other water surfaces (surface zone) in a large basin is very small compared to the basin evapotranspiration; groundwater evapotranspiration is also taken here as being relatively small.

4.1.5 Mass Conservation

Percolation from the upper zone enters the lower soil zone, and deep percolation from the lower zone enters the groundwater zone; see Figure 2. Lateral flows from these zones of surface runoff, interflow, and groundwater flow, respectively, enter the surface storage zone, which represents surface water that ultimately flows from the basin. These flow rates are taken as instantaneously proportional to their respective storages (linear-reservoir flow concept). The mass balances for snowpack, upper and lower soil zones, groundwater, and surface water use these physically-based concepts, in the cascade of Figure 2, to form a set of simultaneous ordinary linear differential equations whose joint solution depends upon the relative magnitude of all parameters, inputs, and system states (storages) pictured in Figure 2.

$$\frac{\partial}{\partial t}USZM = ns_a\left(1-\frac{USZM}{USZC}\right)-\alpha_{per}USZM-\beta_{eu}e_pUSZM \tag{15}$$

$$\frac{\partial}{\partial}USZM = \alpha_{per}USZM-\alpha_{\text{int}}LSZM-\alpha_{dp}LSZM-\beta_{el}e_pLSZM \tag{16}$$

$$\frac{\partial}{\partial}GZM = \alpha_{dp}LSZM-\alpha_{gw}GZM-\beta_{eg}GZM \tag{17}$$

$$\frac{\partial}{\partial t}SS = ns_a\frac{USZM}{USZC}+\alpha_{\text{int}}LSZM+\alpha_{gw}GZM-\alpha_{sf}SS-\beta_{es}e_pSS \tag{18}$$

$$Q=\alpha_{sf}\int_0^{\Delta}SS\,\partial t \tag{19}$$

where α_{per} = percolation coefficient (d^{-1}), β_{eu} = upper zone evapotranspiration coefficient (m^{-3}), α_{int} = interflow coefficient (d^{-1}), *LSZM* = content of lower soil zone (m^3), α_{dp} = deep percolation coefficient (d^{-1}), β_{el} = lower zone evapotranspiration coefficient (m^{-3}), α_{gw} = groundwater coefficient (d^{-1}), *GZM* = content of groundwater zone (m^3), β_{eg} = groundwater zone evapotranspiration coefficient (m^{-3}), α_{sf} = surface outflow coefficient (d^{-1}), *SS* = content of surface storage zone (m^3), β_{es} = surface zone evapotranspiration coefficient (m^{-3}), *Q* = basin outflow volume for *n* days (m^3), and Δ = *n* times *d*. The value of e_p is determined by simultaneous solution of (15)–(19) and the following complementary relationship between actual evapotranspiration and that still possible from atmospheric heat, derived from (13) and (14):

$$\int_0^{\Delta}\left[e_p+(\beta_{eu}USZM+\beta_{el}LSZM+\beta_{eg}GZM+\beta_{es}SS)e_p\right]\theta t=\Psi_a/\left(p_w\gamma_v\right) \tag{20}$$

4.1.6 Analytical Solution

In the analytical solution, results from one storage zone are used in other zones where their outputs appear as inputs. There are 30 different analytic results, depending upon the relative magnitudes of the inputs (*ns*), the initial conditions ($USZM_0$, $LSZM_0$, GZM_0, SS_0, SNW_0), and the model parameters (T_b, a_s, α_{per}, β_{eu}, α_{int}, α_{dp}, b_{el}, α_{gw}, and α_{sf}) in (15)–(20) (note that β_{eg} and β_{es} are taken as zeroes). Complete analytic solutions for all possible ranges of values are available (Croley, 1982). Since the daily inputs and initial storages change from day to day, the appropriate analytic result, as well as its solution, varies with time; mathematical continuity between solutions is preserved however. Small parameter values for a tank outflow imply small releases and large storage volumes; large values imply small storages and outflows nearly equal to inflows. The differential equations for the mass balances can be applied over any time increment by assuming that the input (precipitation and snowmelt) and heat available for evapotranspiration are uniform over the time increment. Thus, the resolution of the equations is limited only by the intervals over which precipitation and temperature data are available; numerical solutions are unnecessary so that approximation errors are avoided. Furthermore, solutions may proceed for either flow rates or storage volumes directly. The mass-balance computation interval may be any length greater than or equal to the interval length for which meteorological data are available.

4.1.7 Application

The model is applied to daily data with either a fixed 1-d or a fixed 7-d mass-balance computation interval. Input and heat available for evapotranspiration are combined on a daily basis and summed over the interval as input to the mass-balance computations. The model is applied to monthly data with a variable mass-balance computation interval. The interval may represent 28–31 d, depending on the month and year. Input and heat available for evapotranspiration are computed over the same monthly interval. Data requirements include initial storage values, daily maximum and minimum air temperatures, daily precipitation, and for comparison purposes, daily basin outflow. Other data requirements are easily met. The mid-monthly extra-terrestrial solar radiation (from which daily values are interpolated) and the empirical constants, b_1 and b_2, are available in standard climatologic summaries. The area of the watershed is also required.

For application of the LBRM to a Great Lakes drainage basin, the basin is first divided into subbasins draining directly to the lake (there are 121 subbasins in the entire Great Lakes basin). The meteorological data from typically 150–300 stations about and in the subbasins are combined through Thiessen weighting to produce areally-averaged daily time series of precipitation and minimum and maximum air temperatures for each subbasin. Weights are determined for each day of record, if necessary, since the data collection network changes frequently as stations are added, dropped, and moved or fail to report from time to time. This is feasible through the use of an algorithm for determining a Thiessen area-of-influence about a station by its edge [Croley and Hartmann, 1985]. Records for all "most-downstream" flow stations are combined by aggregating and extrapolating for ungaged areas to estimate the daily runoff to the lake from each subbasin. Thus, the LBRM is applied in a "distributed-parameter" application by combining model outflows from each of the subbasins to produce the entire basin runoff.

By combining the meteorological and hydrological data for all subbasins to represent the entire basin, the LBRM may be calibrated in a lumped-parameter application to the entire basin at one time. Although the application of lumped-parameter models to very large areas necessarily fails to represent areal distributions of watershed and meteorological characteristics, spatial filtering effects tend to cancel data errors for small areas as the areas are added together. Distributed-parameter applications, in which the LBRM is calibrated for each subbasin and model outflows are combined to represent the entire basin, make use of information that is lost in the lumped-parameter approach; the integration then filters individual subbasin model errors.

There are five variables to be initialized prior to modeling: *SNW, USZM, LSZM, GZM*, and *SS*. While the initial snowpack, SNW_0, is easy to determine as zero during major portions of the year, these variables are generally difficult to estimate. If the model is to be used in forecasting or for short simulations, then it is important to determine these variables accurately prior to use of the model. If the model is to be used for calibration or for long simulations, then the initial values are generally unimportant. The effect of the initial values diminishes with the length of the simulation and after 1 year of simulation, the effects are nil from a practical point of view. Calibrations are repeated with initial conditions equal to observed long-term averages until there is no change in the averages to avoid arbitrary initial conditions when their effects do not diminish rapidly.

4.1.8 Calibration

We calibrate the LBRM for each subbasin with 30 years of daily weighted subbasin climatologic data. The nine parameters are determined (Croley and Hartmann, 1984) by searching the parameter space systematically, minimizing the root mean square error between model and actual outflows for each parameter, selected in rotation, until all parameters converge within two significant digits. Comparisons with other runoff models (Croley, 1983a) and climatology (Croley and Hartmann, 1984) show the LBRM to be far superior for estimates of runoff volumes from large basins.

The LBRM captures a “realism” in its structure that has several advantages over other models. Basin storages, modeled as “tanks”, are automatically removed as respective parameters approach their limits. Thus, the structure of the model changes within a calibration. This is achieved without the use of “threshold” parameters in the model since physical concepts are used which avoid discontinuities in the goodness-of-fit as a function of the parameters; these concepts appear especially relevant for large-basin modeling. Because the “tanks” relate directly to actual basin storages, initialization of the model corresponds to identifying storages from field conditions which may be measured; interpretations of a basin’s hydrology then can aid in setting both initial and boundary conditions. The tanks in Figure 2 may be initialized to correspond to measurements of snow and soil moisture water equivalents available from aerial or satellite monitoring. Snow water equivalents are used in Lake Superior applications (Gauthier et al., 1984).

The LBRM calibration periods generally cover 1965–1982 depending upon flow data availability. Tables 5–11 present the LBRM calibrated parameters. Table 12 presents overall calibration results for the distributed-parameter applications. The LBRM was also used in forecasts of Lake Superior water levels (Croley and Hartmann, 1987), and comparisons with climatic outlooks showed the runoff model was very close to actual runoff (monthly correlations of water supply were on the order of 0.99) for the period August 1982–December 1984, which is outside of and wetter than the calibration period (Croley and Hartmann, 1986). The model also was used to simulate flows for the time period 1956–63, outside of the period of calibration. The correlation of monthly flow volumes between the model and observed during this verification period are also contained in Table 12. They are a little lower than the calibration correlations but quite good except for Lakes Superior and Huron (there were less than two thirds as many flow gages available for 1956–63 as for the calibration period for these basins).

Studies on the Lake Ontario basin (Croley, 1982, 1983b) show that the simple search algorithm described herein does not give unique optimums for calibrated parameter sets because of synergistic relationships between parameters. However, the calibration procedure does show a high degree of repeatability for recalibrations with different starting values, and consistent parameter values are obtained for subbasins with similar hydrological characteristics. On the other hand, the nonuniqueness of calibrated parameters was demonstrated by recalibrating for a synthetic data set. The model was calibrated for the entire Lake Superior basin and then used to simulate outflows to create a new data set for calibration. Subsequent calibration started with a very different initial parameter set and yielded an “optimum” parameter set different from the original with a relatively poor goodness-of-fit. If the original parameter set had been unique, the parameter values produced from the recalibration to the synthetic data set should have been the same as the parameters used to create that data set. This illustrates the nonuniqueness of the parameters, the importance of the starting values used in the search, and the problems inherent in searching the parameter space. Additionally, some components of the LBRM (such as linear reservoirs) are more likely to adequately represent their processes in the real world than others (such as degree-day melting or complementary evapotranspiration). Parameter estimation techniques that properly weight a model’s more accurate parts could improve parameter estimates.

Table 5.--Large Basin Runoff Model Parameters for the Lake Superior Subbasins.

No.	T_b ºC	a_s $m^3\ ºC^{-1}\ d^{-1}$	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	3.0	$.22\times10^{+8}$	$.30\times10^{+0}$	$.59\times10^{+2}$	$.21\times10^{-9}$	$.60\times10^{-2}$	$.10\times10^{-9}$	$.35\times10^{-1}$	$.86\times10^{-1}$	$2.28\times10^{+15}$
2	7.2	$.94\times10^{+7}$	$.12\times10^{+3}$	$.10\times10^{-5}$	$.10\times10^{+0}$	$.76\times10^{-1}$	$.79\times10^{-8}$	$.33\times10^{-1}$	$.23\times10^{+2}$	$4.06\times10^{+16}$
3	4.7	$.11\times10^{+8}$	$.23\times10^{+1}$	$.19\times10^{-7}$	$.39\times10^{-1}$	$.30\times10^{-1}$	$.32\times10^{+4}$	$.30\times10^{-1}$	$.22\times10^{+0}$	$6.88\times10^{+15}$
4	6.0	$.70\times10^{+7}$	$.71\times10^{+0}$	$.85\times10^{-7}$	$.11\times10^{-9}$	$.18\times10^{-1}$	$.30\times10^{-8}$	$.67\times10^{+0}$	$.22\times10^{+0}$	$1.38\times10^{+16}$
5	6.3	$.92\times10^{+7}$	$.23\times10^{+2}$	$.12\times10^{-9}$	$.40\times10^{+1}$	$.63\times10^{+1}$	$.26\times10^{+1}$	$.14\times10^{-1}$	$.26\times10^{+0}$	$2.24\times10^{+16}$
6	4.5	$.85\times10^{+7}$	$.54\times10^{+0}$	$.48\times10^{-7}$	$.21\times10^{-9}$	$.27\times10^{+0}$	$.95\times10^{-7}$	$.58\times10^{-1}$	$.43\times10^{+0}$	$5.49\times10^{+15}$
7	5.2	$.38\times10^{+7}$	$.10\times10^{+1}$	$.70\times10^{-7}$	$.11\times10^{-1}$	$.24\times10^{-2}$	$.35\times10^{-8}$	$.85\times10^{-2}$	$.17\times10^{+0}$	$5.95\times10^{+15}$
8	$9.2\times10^{+9}$	$.10\times10^{+10}$	$.14\times10^{+2}$	$.45\times10^{-7}$	$.10\times10^{-9}$	$.99\times10^{-2}$	$.78\times10^{-9}$	$.14\times10^{-1}$	$.13\times10^{-1}$	$1.05\times10^{+17}$
9	2.3	$.70\times10^{+7}$	$.46\times10^{+0}$	$.35\times10^{+4}$	$.14\times10^{-2}$	$.82\times10^{-2}$	$.40\times10^{-5}$	$.89\times10^{-3}$	$.75\times10^{-1}$	$8.39\times10^{+13}$
10	2.3	$.48\times10^{+7}$	$.46\times10^{+0}$	$.35\times10^{+4}$	$.14\times10^{-2}$	$.82\times10^{-2}$	$.40\times10^{-5}$	$.89\times10^{-3}$	$.75\times10^{-1}$	$6.70\times10^{+13}$
11	2.3	$.19\times10^{+7}$	$.46\times10^{+0}$	$.35\times10^{+4}$	$.14\times10^{-2}$	$.82\times10^{-2}$	$.40\times10^{-5}$	$.89\times10^{-3}$	$.75\times10^{-1}$	$2.90\times10^{+13}$
12	5.1	$.13\times10^{+8}$	$.68\times10^{+0}$	$.22\times10^{+4}$	$.11\times10^{+4}$	$.79\times10^{+4}$	$.90\times10^{+0}$	$.14\times10^{-1}$	$.19\times10^{+0}$	$2.01\times10^{+16}$
13	1.1	$.82\times10^{+7}$	$.93\times10^{+1}$	$.91\times10^{+4}$	$.11\times10^{-2}$	$.92\times10^{-2}$	$.69\times10^{-4}$	$.10\times10^{-2}$	$.95\times10^{-1}$	$2.00\times10^{+10}$
14	3.9	$.24\times10^{+8}$	$.47\times10^{+1}$	$.91\times10^{-8}$	$.27\times10^{-2}$	$.43\times10^{-2}$	$.20\times10^{-9}$	$.17\times10^{-2}$	$.84\times10^{-1}$	$1.87\times10^{+16}$
15	9.7	$.20\times10^{+8}$	$.47\times10^{+2}$	$.53\times10^{-6}$	$.20\times10^{-1}$	$.79\times10^{-2}$	$.39\times10^{-9}$	$.53\times10^{+0}$	$.11\times10^{+0}$	$1.59\times10^{+17}$
16	5.8	$.16\times10^{+8}$	$.54\times10^{+0}$	$.87\times10^{-8}$	$.10\times10^{-9}$	$.27\times10^{-1}$	$.33\times10^{-8}$	$.60\times10^{-1}$	$.10\times10^{+0}$	$4.91\times10^{+16}$
17	1.8	$.11\times10^{+8}$	$.55\times10^{+0}$	$.44\times10^{-1}$	$.32\times10^{-2}$	$.20\times10^{-2}$	$.98\times10^{-2}$	$.59\times10^{-2}$	$.71\times10^{-1}$	$1.02\times10^{+14}$
18	1.4	$.37\times10^{+7}$	$.34\times10^{+0}$	$.64\times10^{-3}$	$.99\times10^{-3}$	$.10\times10^{-2}$	$.96\times10^{-1}$	$.93\times10^{+1}$	$.81\times10^{-1}$	$7.02\times10^{+11}$
19	2.0	$.18\times10^{+8}$	$.91\times10^{+9}$	$.20\times10^{-9}$	$.79\times10^{-2}$	$.81\times10^{-1}$	$.27\times10^{-8}$	$.53\times10^{-2}$	$.52\times10^{-2}$	$4.23\times10^{+14}$
20	2.1	$.11\times10^{+8}$	$.61\times10^{+0}$	$.13\times10^{-6}$	$.22\times10^{-2}$	$.83\times10^{-3}$	$.57\times10^{-7}$	$.99\times10^{+9}$	$.42\times10^{-1}$	$1.74\times10^{+14}$
21	2.0	$.27\times10^{+8}$	$.57\times10^{+1}$	$.76\times10^{+1}$	$.42\times10^{-2}$	$.68\times10^{-2}$	$.46\times10^{-1}$	$.67\times10^{-2}$	$.11\times10^{+0}$	$1.18\times10^{+14}$
22	3.1	$.13\times10^{+8}$	$.54\times10^{+0}$	$.69\times10^{-1}$	$.11\times10^{-9}$	$.18\times10^{+0}$	$.63\times10^{+0}$	$.26\times10^{-1}$	$.18\times10^{+0}$	$4.17\times10^{+15}$

Table 6.-- Large Basin Runoff Model Parameters for the Lake Michigan Subbasins.

No.	T_b ºC	a_s $m^3\ ºC^{-1}\ d^{-1}$	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	4.5	$.36\times10^{+7}$	$.25\times10^{+0}$	$.16\times10^{-6}$	$.11\times10^{+0}$	$.25\times10^{+0}$	$.10\times10^{-9}$	$.50\times10^{-2}$	$.58\times10^{-1}$	$2.83\times10^{+15}$
2	4.5	$.94\times10^{+7}$	$.25\times10^{+0}$	$.16\times10^{-6}$	$.11\times10^{+0}$	$.25\times10^{+0}$	$.10\times10^{-9}$	$.50\times10^{-2}$	$.58\times10^{-1}$	$6.97\times10^{+15}$
3	7.1	$.30\times10^{+7}$	$.10\times10^{+2}$	$.99\times10^{-6}$	$.47\times10^{-1}$	$.25\times10^{-1}$	$.86\times10^{-8}$	$.12\times10^{-1}$	$.69\times10^{+0}$	$1.08\times10^{+16}$
4	7.1	$.33\times10^{+7}$	$.10\times10^{+2}$	$.99\times10^{-6}$	$.47\times10^{-1}$	$.25\times10^{-1}$	$.86\times10^{-8}$	$.12\times10^{-1}$	$.69\times10^{+0}$	$1.20\times10^{+16}$
5	5.2	$.50\times10^{+7}$	$.72\times10^{+0}$	$.12\times10^{-6}$	$.92\times10^{-2}$	$.90\times10^{-7}$	$.26\times10^{-8}$	$.33\times10^{-1}$	$.11\times10^{+0}$	$7.48\times10^{+15}$
6	5.1	$.52\times10^{+7}$	$.29\times10^{+0}$	$.11\times10^{-6}$	$.10\times10^{-4}$	$.68\times10^{-1}$	$.83\times10^{-1}$	$.53\times10^{-1}$	$.96\times10^{-1}$	$7.66\times10^{+15}$
7	3.9	$.24\times10^{+8}$	$.32\times10^{+1}$	$.11\times10^{-7}$	$.74\times10^{-2}$	$.46\times10^{-2}$	$.15\times10^{-8}$	$.98\times10^{-3}$	$.96\times10^{-1}$	$9.74\times10^{+15}$
8	4.7	$.77\times10^{+7}$	$.17\times10^{+1}$	$.12\times10^{-6}$	$.80\times10^{-2}$	$.62\times10^{-2}$	$.63\times10^{-8}$	$.52\times10^{-2}$	$.12\times10^{+0}$	$5.53\times10^{+15}$
9	5.7	$.71\times10^{+7}$	$.35\times10^{+1}$	$.29\times10^{-6}$	$.89\times10^{-2}$	$.10\times10^{-1}$	$.51\times10^{-8}$	$.34\times10^{-2}$	$.12\times10^{+0}$	$9.42\times10^{+15}$
10	6.0	$.33\times10^{+7}$	$.11\times10^{+0}$	$.18\times10^{+2}$	$.38\times10^{-5}$	$.40\times10^{-5}$	$.10\times10^{+2}$	$.64\times10^{-1}$	$.22\times10^{+0}$	$4.43\times10^{+15}$
11	9.4	$.12\times10^{+10}$	$.27\times10^{+3}$	$.43\times10^{-6}$	$.10\times10^{-1}$	$.91\times10^{-7}$	$.25\times10^{-9}$	$.40\times10^{+0}$	$.21\times10^{+0}$	$1.78\times10^{+17}$
12	5.6	$.78\times10^{+7}$	$.13\times10^{+1}$	$.20\times10^{+2}$	$.18\times10^{-1}$	$.95\times10^{-6}$	$.75\times10^{+0}$	$.25\times10^{-5}$	$.45\times10^{+0}$	$6.40\times10^{+15}$
13	5.9	$.17\times10^{+8}$	$.16\times10^{+1}$	$.19\times10^{+0}$	$.19\times10^{-1}$	$.82\times10^{-5}$	$.99\times10^{-2}$	$.98\times10^{-5}$	$.15\times10^{+0}$	$1.36\times10^{+16}$
14	5.7	$.70\times10^{+7}$	$.40\times10^{+1}$	$.32\times10^{-5}$	$.28\times10^{-1}$	$.34\times10^{-1}$	$.21\times10^{-6}$	$.10\times10^{-1}$	$.38\times10^{+0}$	$5.78\times10^{+15}$
15	8.1	$.92\times10^{+7}$	$.48\times10^{+0}$	$.39\times10^{-6}$	$.80\times10^{-5}$	$.68\times10^{+6}$	$.38\times10^{+0}$	$.22\times10^{-1}$	$.30\times10^{+0}$	$1.72\times10^{+16}$
16	8.3	$.53\times10^{+8}$	$.17\times10^{+2}$	$.47\times10^{-7}$	$.86\times10^{-1}$	$.24\times10^{+0}$	$.39\times10^{-7}$	$.14\times10^{-1}$	$.13\times10^{+0}$	$8.19\times10^{+16}$
17	5.6	$.10\times10^{+9}$	$.84\times10^{+6}$	$.17\times10^{-6}$	$.42\times10^{-1}$	$.50\times10^{-1}$	$.13\times10^{-6}$	$.17\times10^{-1}$	$.26\times10^{+2}$	$2.03\times10^{+15}$
18	8.5	$.26\times10^{+8}$	$.89\times10^{+1}$	$.34\times10^{-6}$	$.11\times10^{-1}$	$.11\times10^{-1}$	$.13\times10^{-8}$	$.43\times10^{-2}$	$.18\times10^{+0}$	$3.88\times10^{+16}$
19	5.9	$.23\times10^{+7}$	$.11\times10^{-4}$	$.57\times10^{-2}$	$.93\times10^{+1}$	$.66\times10^{+0}$	$.11\times10^{-3}$	$.12\times10^{-4}$	$.58\times10^{+0}$	$1.68\times10^{+15}$
20	5.9	$.43\times10^{+8}$	$.20\times10^{+0}$	$.25\times10^{-6}$	$.10\times10^{-2}$	$.19\times10^{-1}$	$.90\times10^{-8}$	$.21\times10^{-1}$	$.57\times10^{-1}$	$4.01\times10^{+16}$
21	4.8	$.76\times10^{+6}$	$.50\times10^{+1}$	$.49\times10^{-6}$	$.44\times10^{-2}$	$.64\times10^{-2}$	$.51\times10^{-9}$	$.17\times10^{-3}$	$.15\times10^{+0}$	$3.36\times10^{+14}$
22	6.0	$.23\times10^{+8}$	$.38\times10^{+1}$	$.16\times10^{-6}$	$.57\times10^{-2}$	$.41\times10^{-2}$	$.39\times10^{-9}$	$.12\times10^{-2}$	$.13\times10^{+0}$	$2.76\times10^{+16}$
23	4.8	$.15\times10^{+8}$	$.50\times10^{+1}$	$.49\times10^{-6}$	$.44\times10^{-2}$	$.64\times10^{-2}$	$.51\times10^{-9}$	$.17\times10^{-3}$	$.15\times10^{+0}$	$8.22\times10^{+15}$
24	7.5	$.20\times10^{+10}$	$.18\times10^{+4}$	$.62\times10^{-4}$	$.60\times10^{-2}$	$.75\times10^{-2}$	$.24\times10^{-9}$	$.26\times10^{-3}$	$.76\times10^{+1}$	$3.82\times10^{+16}$
25	4.8	$.63\times10^{+7}$	$.50\times10^{+1}$	$.49\times10^{-6}$	$.44\times10^{-2}$	$.64\times10^{-2}$	$.51\times10^{-9}$	$.17\times10^{-3}$	$.15\times10^{+0}$	$3.45\times10^{+15}$
26	6.4	$.56\times10^{+7}$	$.60\times10^{+2}$	$.26\times10^{-5}$	$.22\times10^{-1}$	$.12\times10^{+0}$	$.32\times10^{-8}$	$.23\times10^{-3}$	$.51\times10^{+2}$	$1.82\times10^{+16}$
27	6.4	$.11\times10^{+7}$	$.60\times10^{+2}$	$.26\times10^{-5}$	$.22\times10^{-1}$	$.12\times10^{+0}$	$.32\times10^{-8}$	$.23\times10^{-3}$	$.51\times10^{+2}$	$4.01\times10^{+15}$

Table 7.--Large Basin Runoff Model Parameters for the Lake Huron Subbasins.

No.	T_b °C	a_s m^3 °C^{-1} d^{-1}	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	4.9	$.41\times10^{+7}$	$.16\times10^{+0}$	$.29\times10^{-6}$	$.31\times10^{-1}$	$.88\times10^{-5}$	$.55\times10^{-9}$	$.75\times10^{-1}$	$.23\times10^{+0}$	$4.76\times10^{+15}$
2	6.6	$.93\times10^{+6}$	$.14\times10^{+2}$	$.23\times10^{-6}$	$.65\times10^{-2}$	$.82\times10^{-2}$	$.10\times10^{-8}$	$.74\times10^{-3}$	$.14\times10^{+0}$	$1.59\times10^{+15}$
3	6.6	$.13\times10^{+8}$	$.14\times10^{+2}$	$.23\times10^{-6}$	$.65\times10^{-2}$	$.82\times10^{-2}$	$.10\times10^{-8}$	$.74\times10^{-3}$	$.14\times10^{+0}$	$2.44\times10^{+16}$
4	6.6	$.47\times10^{+7}$	$.14\times10^{+2}$	$.23\times10^{-6}$	$.65\times10^{-2}$	$.82\times10^{-2}$	$.10\times10^{-8}$	$.74\times10^{-3}$	$.14\times10^{+0}$	$8.08\times10^{+15}$
5	5.3	$.98\times10^{+7}$	$.47\times10^{+1}$	$.32\times10^{-6}$	$.89\times10^{-2}$	$.82\times10^{-2}$	$.91\times10^{-8}$	$.42\times10^{-2}$	$.23\times10^{+0}$	$1.01\times10^{+16}$
6	5.3	$.14\times10^{+7}$	$.47\times10^{+1}$	$.32\times10^{-6}$	$.89\times10^{-2}$	$.82\times10^{-2}$	$.91\times10^{-8}$	$.42\times10^{-2}$	$.23\times10^{+0}$	$1.32\times10^{+15}$
7	6.2	$.11\times10^{+8}$	$.10\times10^{+2}$	$.20\times10^{-6}$	$.46\times10^{-2}$	$.10\times10^{-1}$	$.49\times10^{-9}$	$.96\times10^{-3}$	$.27\times10^{+0}$	$2.60\times10^{+16}$
8	5.6	$.79\times10^{+7}$	$.30\times10^{+1}$	$.41\times10^{-6}$	$.10\times10^{-1}$	$.55\times10^{-6}$	$.83\times10^{-9}$	$.52\times10^{-3}$	$.36\times10^{+0}$	$9.46\times10^{+15}$
9	5.7	$.34\times10^{+7}$	$.91\times10^{-6}$	$.66\times10^{-3}$	$.71\times10^{+0}$	$.62\times10^{-2}$	$.70\times10^{-9}$	$.30\times10^{-1}$	$.97\times10^{-1}$	$3.68\times10^{+15}$
10	5.4	$.45\times10^{+8}$	$.20\times10^{+0}$	$.32\times10^{-6}$	$.97\times10^{-3}$	$.20\times10^{-1}$	$.55\times10^{-7}$	$.26\times10^{-1}$	$.11\times10^{+0}$	$3.63\times10^{+16}$
11	5.7	$.65\times10^{+7}$	$.80\times10^{-6}$	$.27\times10^{-4}$	$.30\times10^{+0}$	$.60\times10^{-2}$	$.70\times10^{-9}$	$.41\times10^{-1}$	$.21\times10^{+0}$	$6.25\times10^{+15}$
12	6.3	$.81\times10^{+7}$	$.10\times10^{+0}$	$.47\times10^{-2}$	$.17\times10^{-1}$	$.99\times10^{-6}$	$.26\times10^{-8}$	$.25\times10^{-2}$	$.30\times10^{+0}$	$5.69\times10^{+15}$
13	6.3	$.17\times10^{+8}$	$.10\times10^{+0}$	$.47\times10^{-2}$	$.17\times10^{-1}$	$.99\times10^{-6}$	$.26\times10^{-8}$	$.25\times10^{-2}$	$.30\times10^{+0}$	$1.24\times10^{+16}$
14	4.8	$.19\times10^{+8}$	$.42\times10^{+0}$	$.93\times10^{-5}$	$.15\times10^{-1}$	$.86\times10^{-6}$	$.15\times10^{-5}$	$.35\times10^{-3}$	$.23\times10^{+0}$	$4.97\times10^{+15}$
15	6.6	$.12\times10^{+8}$	$.16\times10^{+1}$	$.13\times10^{-1}$	$.19\times10^{-1}$	$.95\times10^{-6}$	$.76\times10^{-2}$	$.80\times10^{-1}$	$.73\times10^{+0}$	$6.83\times10^{+15}$
16	5.3	$.30\times10^{+8}$	$.16\times10^{+1}$	$.22\times10^{-6}$	$.12\times10^{-1}$	$.53\times10^{-2}$	$.50\times10^{-8}$	$.11\times10^{-1}$	$.17\times10^{+0}$	$1.21\times10^{+16}$

Table 8.--Large Basin Runoff Model Parameters for the Georgian Bay Subbasins.

No.	T_b °C	a_s m^3 °C^{-1} d^{-1}	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	4.3	$.16\times10^{+8}$	$.11\times10^{+0}$	$.18\times10^{-3}$	$.21\times10^{-1}$	$.83\times10^{-6}$	$.11\times10^{-4}$	$.87\times10^{-3}$	$.81\times10^{-1}$	$2.97\times10^{+15}$
2	6.1	$.14\times10^{+8}$	$.37\times10^{+1}$	$.20\times10^{-6}$	$.15\times10^{-1}$	$.89\times10^{-6}$	$.43\times10^{-8}$	$.20\times10^{-2}$	$.24\times10^{+0}$	$7.98\times10^{+15}$
3	6.0	$.21\times10^{+8}$	$.33\times10^{+1}$	$.41\times10^{-6}$	$.10\times10^{-1}$	$.77\times10^{-2}$	$.94\times10^{-8}$	$.70\times10^{-2}$	$.35\times10^{+0}$	$1.85\times10^{+16}$
4	4.7	$.23\times10^{+8}$	$.84\times10^{+1}$	$.46\times10^{-6}$	$.71\times10^{-2}$	$.57\times10^{-2}$	$.14\times10^{-7}$	$.13\times10^{-3}$	$.13\times10^{+0}$	$1.15\times10^{+16}$
5	4.8	$.41\times10^{+8}$	$.43\times10^{+4}$	$.80\times10^{-5}$	$.16\times10^{-1}$	$.74\times10^{-2}$	$.27\times10^{-8}$	$.31\times10^{-3}$	$.34\times10^{+0}$	$1.43\times10^{+16}$
6	3.9	$.25\times10^{+8}$	$.52\times10^{+2}$	$.59\times10^{-4}$	$.12\times10^{+0}$	$.53\times10^{-1}$	$.59\times10^{-6}$	$.38\times10^{-5}$	$.79\times10^{-1}$	$6.14\times10^{+15}$
7	2.6	$.16\times10^{+9}$	$.57\times10^{+6}$	$.82\times10^{-9}$	$.75\times10^{-2}$	$.58\times10^{-2}$	$.59\times10^{-9}$	$.50\times10^{-5}$	$.79\times10^{-1}$	$1.30\times10^{+15}$
8	3.7	$.45\times10^{+8}$	$.21\times10^{+2}$	$.22\times10^{-7}$	$.46\times10^{-2}$	$.18\times10^{-2}$	$.81\times10^{-9}$	$.85\times10^{-3}$	$.17\times10^{+0}$	$2.99\times10^{+15}$
9	4.4	$.46\times10^{+8}$	$.12\times10^{+1}$	$.75\times10^{-8}$	$.62\times10^{-2}$	$.35\times10^{-3}$	$.24\times10^{-9}$	$.19\times10^{-1}$	$.75\times10^{-1}$	$3.98\times10^{+16}$
10	4.4	$.99\times10^{+7}$	$.74\times10^{+0}$	$.34\times10^{-6}$	$.26\times10^{-1}$	$.27\times10^{-5}$	$.22\times10^{-8}$	$.46\times10^{-1}$	$.54\times10^{-1}$	$5.01\times10^{+15}$
11	2.2	$.22\times10^{+8}$	$.30\times10^{+1}$	$.70\times10^{-7}$	$.48\times10^{-2}$	$.22\times10^{-2}$	$.50\times10^{-9}$	$.26\times10^{-3}$	$.99\times10^{-1}$	$2.34\times10^{+14}$
12	4.9	$.99\times10^{+7}$	$.16\times10^{+0}$	$.29\times10^{-6}$	$.31\times10^{-1}$	$.88\times10^{-5}$	$.55\times10^{-9}$	$.75\times10^{-1}$	$.23\times10^{+0}$	$1.19\times10^{+16}$
13	4.9	$.29\times10^{+7}$	$.16\times10^{+0}$	$.29\times10^{-6}$	$.31\times10^{-1}$	$.88\times10^{-5}$	$.55\times10^{-9}$	$.75\times10^{-1}$	$.23\times10^{+0}$	$3.41\times10^{+15}$

Table 9.--Large Basin Runoff Model Parameters for the Lake St. Clair Subbasins.

No.	T_b °C	a_s m^3 °C^{-1} d^{-1}	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	6.4	$.11\times10^{+8}$	$.97\times10^{-1}$	$.11\times10^{+1}$	$.90\times10^{-6}$	$.10\times10^{-1}$	$.98\times10^{+0}$	$.42\times10^{-1}$	$.25\times10^{+0}$	$1.17\times10^{+16}$
2	8.2	$.72\times10^{+6}$	$.16\times10^{+2}$	$.42\times10^{-5}$	$.10\times10^{+0}$	$.16\times10^{+0}$	$.11\times10^{-6}$	$.15\times10^{-1}$	$.31\times10^{+1}$	$2.85\times10^{+15}$
3	8.2	$.53\times10^{+7}$	$.16\times10^{+2}$	$.42\times10^{-5}$	$.10\times10^{+0}$	$.16\times10^{+0}$	$.11\times10^{-6}$	$.15\times10^{-1}$	$.31\times10^{+1}$	$1.30\times10^{+16}$
4	8.2	$.37\times10^{+6}$	$.16\times10^{+2}$	$.42\times10^{-5}$	$.10\times10^{+0}$	$.16\times10^{+0}$	$.11\times10^{-6}$	$.15\times10^{-1}$	$.31\times10^{+1}$	$1.31\times10^{+15}$
5	8.5	$.53\times10^{+7}$	$.11\times10^{+0}$	$.53\times10^{-5}$	$.51\times10^{-1}$	$.60\times10^{-5}$	$.86\times10^{-7}$	$.51\times10^{-1}$	$.65\times10^{+0}$	$5.02\times10^{+15}$
6	7.8	$.24\times10^{+8}$	$.41\times10^{+5}$	$.72\times10^{+2}$	$.31\times10^{+0}$	$.11\times10^{+0}$	$.34\times10^{-7}$	$.21\times10^{-1}$	$.28\times10^{+0}$	$3.96\times10^{+16}$
7	6.4	$.18\times10^{+8}$	$.45\times10^{-1}$	$.34\times10^{-4}$	$.97\times10^{-2}$	$.11\times10^{-5}$	$.51\times10^{-8}$	$.59\times10^{-1}$	$.18\times10^{+0}$	$1.33\times10^{+16}$

Table 10.--Large Basin Runoff Model Parameters for the Lake Erie Subbasins.

No.	T_b ºC	a_s $m^3\,{}^oC^{-1}\,d^{-1}$	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	8.0	$.47\times10^{+7}$	$.13\times10^{+1}$	$.67\times10^{-6}$	$.24\times10^{-1}$	$.53\times10^{-1}$	$.14\times10^{-6}$	$.14\times10^{-1}$	$.11\times10^{+1}$	$1.02\times10^{+16}$
2	10.	$.85\times10^{+7}$	$.30\times10^{+2}$	$.57\times10^{-5}$	$.45\times10^{-1}$	$.12\times10^{+0}$	$.59\times10^{-7}$	$.17\times10^{-1}$	$.40\times10^{+0}$	$2.30\times10^{+16}$
3	6.2	$.23\times10^{+7}$	$.73\times10^{+0}$	$.37\times10^{-2}$	$.10\times10^{-4}$	$.17\times10^{-1}$	$.65\times10^{-3}$	$.27\times10^{-1}$	$.36\times10^{+0}$	$2.29\times10^{+15}$
4	7.9	$.11\times10^{+8}$	$.88\times10^{-5}$	$.95\times10^{-2}$	$.90\times10^{-1}$	$.79\times10^{-1}$	$.88\times10^{-4}$	$.10\times10^{+0}$	$.10\times10^{+0}$	$1.79\times10^{+16}$
5	8.1	$.42\times10^{+7}$	$.58\times10^{-4}$	$.11\times10^{-5}$	$.60\times10^{+5}$	$.99\times10^{-1}$	$.10\times10^{-3}$	$.20\times10^{+0}$	$.33\times10^{+0}$	$5.92\times10^{+15}$
6	6.6	$.74\times10^{+8}$	$.39\times10^{-1}$	$.45\times10^{-7}$	$.97\times10^{-5}$	$.32\times10^{-5}$	$.64\times10^{-7}$	$.49\times10^{-1}$	$.19\times10^{+0}$	$6.07\times10^{+16}$
7	6.4	$.90\times10^{+7}$	$.43\times10^{-1}$	$.47\times10^{-6}$	$.86\times10^{-5}$	$.61\times10^{-5}$	$.12\times10^{-5}$	$.64\times10^{-1}$	$.41\times10^{+0}$	$7.50\times10^{+15}$
8	5.9	$.41\times10^{+8}$	$.49\times10^{-1}$	$.12\times10^{-5}$	$.60\times10^{-2}$	$.25\times10^{-5}$	$.16\times10^{-5}$	$.60\times10^{-1}$	$.28\times10^{+0}$	$1.10\times10^{+16}$
9	7.1	$.12\times10^{+8}$	$.92\times10^{-6}$	$.49\times10^{-5}$	$.10\times10^{-4}$	$.30\times10^{-5}$	$.59\times10^{-7}$	$.50\times10^{-1}$	$.12\times10^{+1}$	$9.28\times10^{+15}$
10	5.2	$.73\times10^{+7}$	$.11\times10^{+0}$	$.29\times10^{-6}$	$.12\times10^{-1}$	$.63\times10^{-5}$	$.23\times10^{-6}$	$.59\times10^{-1}$	$.66\times10^{+0}$	$3.77\times10^{+15}$
11	7.6	$.57\times10^{+7}$	$.70\times10^{+1}$	$.84\times10^{-6}$	$.11\times10^{+0}$	$.57\times10^{-1}$	$.16\times10^{-7}$	$.22\times10^{-1}$	$.32\times10^{+1}$	$1.12\times10^{+16}$
12	5.8	$.38\times10^{+7}$	$.11\times10^{+1}$	$.12\times10^{-5}$	$.65\times10^{-1}$	$.58\times10^{-5}$	$.19\times10^{-7}$	$.29\times10^{-1}$	$.34\times10^{+1}$	$2.44\times10^{+15}$
13	5.1	$.60\times10^{+7}$	$.64\times10^{-6}$	$.13\times10^{-5}$	$.10\times10^{-4}$	$.30\times10^{-5}$	$.59\times10^{-7}$	$.50\times10^{-1}$	$.33\times10^{+0}$	$3.16\times10^{+15}$
14	4.5	$.50\times10^{+8}$	$.64\times10^{-1}$	$.23\times10^{-5}$	$.85\times10^{-1}$	$.44\times10^{-5}$	$.19\times10^{-3}$	$.61\times10^{-1}$	$.68\times10^{+0}$	$6.33\times10^{+14}$
15	4.4	$.79\times10^{+7}$	$.32\times10^{-1}$	$.49\times10^{-6}$	$.39\times10^{-1}$	$.54\times10^{-5}$	$.10\times10^{-9}$	$.58\times10^{-1}$	$.66\times10^{+0}$	$1.99\times10^{+15}$
16	4.3	$.35\times10^{+7}$	$.17\times10^{+1}$	$.11\times10^{-5}$	$.15\times10^{+0}$	$.14\times10^{+0}$	$.53\times10^{-7}$	$.24\times10^{-1}$	$.68\times10^{+1}$	$1.57\times10^{+15}$
17	4.6	$.80\times10^{+7}$	$.16\times10^{+1}$	$.14\times10^{-4}$	$.17\times10^{+0}$	$.20\times10^{+0}$	$.15\times10^{-5}$	$.27\times10^{-1}$	$.66\times10^{+1}$	$2.55\times10^{+15}$
18	4.6	$.51\times10^{+6}$	$.16\times10^{+1}$	$.14\times10^{-4}$	$.17\times10^{+0}$	$.20\times10^{+0}$	$.15\times10^{-5}$	$.27\times10^{-1}$	$.66\times10^{+1}$	$1.33\times10^{+14}$
19	9.4	$.31\times10^{+8}$	$.27\times10^{+2}$	$.80\times10^{-2}$	$.33\times10^{+0}$	$.40\times10^{+0}$	$.38\times10^{-7}$	$.18\times10^{-1}$	$.35\times10^{+0}$	$7.44\times10^{+16}$
20	6.9	$.24\times10^{+8}$	$.26\times10^{+1}$	$.28\times10^{-5}$	$.57\times10^{-1}$	$.72\times10^{-1}$	$.16\times10^{-6}$	$.13\times10^{-1}$	$.51\times10^{+0}$	$2.08\times10^{+16}$
21	14.	$.14\times10^{+8}$	$.91\times10^{+2}$	$.11\times10^{-3}$	$.35\times10^{+0}$	$.44\times10^{-5}$	$.54\times10^{-7}$	$.75\times10^{-1}$	$.10\times10^{+1}$	$3.09\times10^{+16}$

Table 11.--Large Basin Runoff Model Parameters for the Lake Ontario Subbasins.

No.	T_b ºC	a_s $m^3\,{}^oC^{-1}\,d^{-1}$	α_{per} d^{-1}	β_{eu} m^{-3}	α_{int} d^{-1}	α_{dp} d^{-1}	β_{el} m-3	α_{gw} d^{-1}	α_{sf} d^{-1}	K cal
1	6.0	$.17\times10^{+8}$	$.51\times10^{-1}$	$.44\times10^{-4}$	$.61\times10^{-2}$	$.33\times10^{-5}$	$.15\times10^{-3}$	$.24\times10^{+0}$	$.33\times10^{+0}$	$1.39\times10^{+16}$
2	6.0	$.97\times10^{+7}$	$.51\times10^{-1}$	$.44\times10^{-4}$	$.61\times10^{-2}$	$.33\times10^{-5}$	$.15\times10^{-3}$	$.24\times10^{+0}$	$.33\times10^{+0}$	$7.70\times10^{+15}$
3	4.2	$.16\times10^{+8}$	$.90\times10^{-4}$	$.56\times10^{-7}$	$.20\times10^{-1}$	$.73\times10^{-5}$	$.14\times10^{-9}$	$.67\times10^{-1}$	$.63\times10^{-1}$	$5.70\times10^{+15}$
4	5.5	$.43\times10^{+7}$	$.49\times10^{+0}$	$.33\times10^{-5}$	$.17\times10^{-5}$	$.19\times10^{+0}$	$.36\times10^{-6}$	$.22\times10^{-1}$	$.40\times10^{+0}$	$3.66\times10^{+15}$
5	4.3	$.49\times10^{+8}$	$.12\times10^{+0}$	$.19\times10^{-7}$	$.41\times10^{-1}$	$.74\times10^{-5}$	$.96\times10^{-7}$	$.29\times10^{+0}$	$.45\times10^{-1}$	$1.12\times10^{+16}$
6	4.3	$.70\times10^{+7}$	$.18\times10^{+1}$	$.33\times10^{-5}$	$.12\times10^{+0}$	$.87\times10^{-1}$	$.39\times10^{-7}$	$.42\times10^{-1}$	$.36\times10^{+1}$	$2.28\times10^{+15}$
7	4.4	$.22\times10^{+8}$	$.10\times10^{-5}$	$.11\times10^{-7}$	$.20\times10^{-4}$	$.30\times10^{-5}$	$.81\times10^{-7}$	$.50\times10^{-1}$	$.59\times10^{-1}$	$8.62\times10^{+15}$
8	3.1	$.18\times10^{+8}$	$.11\times10^{+0}$	$.10\times10^{-6}$	$.21\times10^{-1}$	$.14\times10^{-5}$	$.10\times10^{-9}$	$.36\times10^{-1}$	$.94\times10^{-1}$	$9.13\times10^{+14}$
9	4.4	$.12\times10^{+8}$	$.83\times10^{+0}$	$.55\times10^{-1}$	$.89\times10^{-6}$	$.15\times10^{-1}$	$.20\times10^{-2}$	$.79\times10^{+2}$	$.10\times10^{+0}$	$2.11\times10^{+15}$
10	4.4	$.17\times10^{+8}$	$.83\times10^{+0}$	$.55\times10^{-1}$	$.89\times10^{-6}$	$.15\times10^{-1}$	$.20\times10^{-2}$	$.79\times10^{+2}$	$.10\times10^{+0}$	$3.39\times10^{+15}$
11	6.9	$.14\times10^{+8}$	$.49\times10^{+1}$	$.59\times10^{-1}$	$.15\times10^{+0}$	$.12\times10^{+0}$	$.27\times10^{-6}$	$.22\times10^{-1}$	$.10\times10^{+0}$	$1.68\times10^{+16}$
12	5.6	$.77\times10^{+8}$	$.57\times10^{+6}$	$.11\times10^{-9}$	$.17\times10^{-1}$	$.49\times10^{-2}$	$.19\times10^{-8}$	$.95\times10^{-2}$	$.53\times10^{+0}$	$4.46\times10^{+16}$
13	5.5	$.95\times10^{+7}$	$.19\times10^{+1}$	$.23\times10^{-4}$	$.95\times10^{-2}$	$.24\times10^{-1}$	$.25\times10^{-7}$	$.71\times10^{-2}$	$.15\times10^{+1}$	$7.43\times10^{+15}$
14	5.5	$.84\times10^{+7}$	$.19\times10^{+1}$	$.31\times10^{-5}$	$.20\times10^{-1}$	$.28\times10^{-1}$	$.23\times10^{-6}$	$.12\times10^{-1}$	$.48\times10^{+0}$	$7.22\times10^{+15}$
15	5.5	$.72\times10^{+7}$	$.19\times10^{+1}$	$.31\times10^{-5}$	$.20\times10^{-1}$	$.28\times10^{-1}$	$.23\times10^{-6}$	$.12\times10^{-1}$	$.48\times10^{+0}$	$6.14\times10^{+15}$

Table 12.--Large Basin Runoff Model Calibration Statistics[a].

Lake	Number of Sub-basins	Mean 1-day Flow (mm)[b]	Flow Standard Deviation (mm)[b]	Root Mean Square Error (mm)[b]	Correlation Calibration	Independent Verification
Superior	22	1.12	0.67	0.25	0.93	0.77
Michigan	29	0.89	0.47	0.18	0.93	0.86
Huron	27	1.06	0.69	0.26	0.92	0.69
St. Clair	7	0.90	1.36	0.62	0.89	0.87
Erie	21	1.01	1.28	0.54	0.91	0.90
Ontario	15	1.41	1.13	0.43	0.93	0.89

[a]Statistics and calibrations generally are 1966-83; verification generally is 1956-63.
[b]Equivalent depth over the land portion of the basin.

4.2 Over-Lake Precipitation

The lack of over-lake precipitation measurements means that estimates typically depend on land-based measurements and that there may be differences between land and lake meteorology. Although gage exposures may significantly influence the results of lake-land precipitation studies (Bolsenga, 1977, 1979), Wilson (1977) found that Lake Ontario precipitation estimates based on only nearshore stations averaged 5.6% more during the warm season and 2.1% less during the cold season than estimates based on stations situated in the lake. By using a network that also included stations somewhat removed from the Lake Ontario shoreline, Bolsenga and Hagman (1975) found that eliminating several gages not immediately in the vicinity of the shoreline increased over-lake precipitation estimates during the warm season and decreased them during the cold season. Thus, for the Great Lakes, where lake effects on nearshore meteorology are significant and the drainage basins have relatively low relief, the use here of all available meteorological stations throughout the basin is probably less biased than the use of only nearshore stations. Over-lake precipitation is taken equal to over-land precipitation (on the basis of depth) without further corrections.

4.3 Over-Lake Evaporation

Great Lakes hydrological research mandates the use of continuous-simulation models of daily lake evaporation over long time periods. Such models must be usable in the absence of water surface temperature and ice cover observations. They also must be physically based to have application under environmental conditions different than those under which they were derived. GLERL developed a lumped-parameter model of evaporation and thermodynamic fluxes for the Great Lakes based on an energy balance at the lake’s surface (Croley, 1989a,b) and on one-dimensional (vertical) lake heat storage (Croley, 1992a). Ice formation and loss is coupled also to lake thermodynamics and heat storage (Croley and Assel, 1993).

4.3.1 Thermodynamic Fluxes

The thermodynamic fluxes to and from a lake include incident short-wave radiation, q_i, reflected short-wave radiation, q_r and q_r (over water and over ice, respectively), evaporative (latent and advected) heat transfer, q_e and q_e', sensible heat transfer, q_{h} and q_h', precipitation heat advection, q_p and q_p', net long-wave radiation exchange, Q_l, and surface flow advection, Q_l; see Croley (1989a,b) for details:

$$q_i = \left[0.355 + 0.68(1 - N)\right]q_0 \tag{21}$$

where q_i = daily average unit (per unit area) rate of short-wave radiation incident to the earth's surface, N = fraction of the sky covered by clouds, and q_0 = daily average unit rate of short-wave radiation received on a horizontal unit area of the Earth's surface under cloudless skies;

$$q_r = 0.1 q_i \tag{22}$$

where q_r = average unit reflected short-wave radiation rate from the water surface;

$$e_w = \rho_a \, C_E (q_w - q) U / \rho_w \tag{23}$$

where e_w = over-water evaporation rate, ρ_a = density of air, C_E = bulk evaporation coefficient over water, q_w = specific humidity of saturated air at the temperature of the water surface, q = specific humidity of the atmosphere over water, and U = wind speed over water;

$$q_e = (\gamma_v + C_w T) e_w \, \rho_w \tag{24}$$

where q_e = average unit evaporative (latent and advected) heat transfer rate from the water surface, C_w = specific heat of water, and T = water surface temperature;

$$q_h = \rho_a \, C_p C_H (T_a - T) U \tag{25}$$

where q_h = average unit sensible heat transfer rate to the water surface, C_p = specific heat of air at constant temperature, and C_H = sensible heat coefficient over water;

$$\begin{aligned} q_p &= (C_w T_a - \gamma_f)\rho_w \; p, && T_a < 0^\circ \mathrm{C} \\ &= C_w T_a \, \rho_w \; p, && T_a \geq 0^\circ \mathrm{C} \end{aligned} \tag{26}$$

where q_p = average unit precipitation heat advection rate to the water surface;

$$q_\uparrow = \varepsilon_w \, \sigma (T + 273.16^\circ \mathrm{C})^4 \tag{27}$$

where $q_\uparrow$ = average unit long-wave radiation emitted by the water body, σ = Stephan-Bolzman constant (5.67×10^{-8} W m^{-2} °K^{-1}), and ε_w = emissivity of the water surface;

$$q_\downarrow = (1 - r_a)\varepsilon_a \, \sigma (T + 273.16^\circ \mathrm{C})^4 \tag{28}$$

where $q_\downarrow$ = average unit long-wave radiation from the atmosphere absorbed by the water surface, r_a = reflectivity of the water surface, and ε_a = emissivity of the atmosphere;

$$Q_l = \{q_\downarrow[\eta + (1-\eta)(1-N)] - q_\uparrow\}(A_w + A) \tag{29}$$

where Q_l = average net long-wave radiation exchange rate between the entire water body and the atmosphere (effects of ice cover on the net long-wave exchange are ignored here), η = empirical coefficient relating cloudiness to atmospheric long-wave radiation, A_w = area of the open-water (ice-free) surface, and A = area of the ice surface;

$$Q_I = \rho_w \, C_w \, T(\Theta_i - \Theta_o) \tag{30}$$

where Q_I = daily net heat advection to the lake from over-land flow and channel inputs and outputs, Θ_i = sum of all surface inflows to the lake, and Θ_o = sum of all outflows from a lake;

$$q_r' = (0.85 f_n + 0.70 f_o + 0.50 f_m + 0.45 f_b) q_i \tag{31}$$

where q_r' = average unit reflected short-wave radiation rate from the ice pack, f_n = fraction of ice covered with new snow, f_o = fraction of ice covered with old snow, f_m = fraction of ice covered with melting snow, and f_b = fraction of ice that is bare of snow;

$$e_w' = \rho_a \, C_E'(q_w' - q')U' / \rho_w \tag{32}$$

where e_w' = over-ice evaporation rate, C_E' = bulk evaporation coefficient over ice, q_w' = specific humidity of saturated air at temperature of ice, q' = specific humidity of the atmosphere over ice, and U' = wind speed over ice;

$$q_e' = (\gamma_v + \gamma_f + C_w \, T')e_w' \rho_w \tag{33}$$

where q_e' = average unit evaporative (latent and advected) heat transfer rate from the ice pack and T' = ice surface temperature;

$$q_h' = \rho_a \, C_p \, C_H'(T_a' - T')U' \tag{34}$$

where q_h' = average unit sensible heat transfer rate to the ice pack, C_H' = sensible heat coefficient over ice, and T_a' = temperature of the air over ice; and

$$q_p' = C_w \, T_a' \rho_w \, p \tag{35}$$

where q_p' = average unit precipitation heat advection rate to the ice pack.

Gray et al. (1973) provided (21), generalized maps of mid-monthly values from which q_0 may be interpolated by date, and the short-wave reflection of (22) and (31). Because data are unavailable and because subsequent heat budgets are insensitive to their values, f_n, f_o, and f_m are set to zero here, and f_b is set to unity. Values of over-water and over-ice meteorology (q, U, T_a, N, q', U', and T_a') are determined from over-land values by adjusting for over-water conditions. Phillips' and Irbe's (1978) regressions for over-water corrections are used directly by replacing the fetch (and derived quantities) with averages. The bulk evaporation coefficients over water and over ice (C_E and C_E') are determined similar to Quinn (1979) from over-water or over-ice (respectively) wind speed, air temperature, and surface temperature. The over-water and over-ice sensible heat coefficients (C_H and C_H') are taken equal to the bulk evaporation coefficients, respectively (Quinn, 1979). The emissivities of water and air in (27) and (28) [note the reflectivity of the water surface in (28) is $r_a = 1 - e_w$] are taken, respectively, as 0.97, and 0.53 + 0.065 $e_a^{1/2}$ where e_a is the vapor pressure of the air (mb) after Keijman (1974).

4.3.2 Heat Storage

The heat added to a lake and the heat added to the ice pack, from the surface fluxes, are governed by simple energy and mass balances, energy-storage relationships, and boundary conditions on ice growth, water temperature, and ice temperature. The rate of change of heat storage in a lake with time is:

$$\frac{\partial H}{\partial t} = A_w\left(q_i - q_r - q_e + q_h + q_p\right) + Q_l + Q_I - Q_w \tag{36}$$

where $\partial H/\partial t$ = time rate of change of heat storage H in the lake, and Q_w = total heat flux between the water body and the ice pack. The rate of change of heat storage in the ice pack with time is defined here as:

$$\frac{\partial H'}{\partial t} = A\left(q_i - q_r' - q_e' + q_h' + q_p'\right) + Q_w \tag{37}$$

where $\partial H'/\partial t$ = time rate of change of heat storage H' in the ice pack.

Kraus and Turner's (1967) mixed-layer thermal structure concept is extended to allow the determination of simple heat storage. Effects of past additions or losses are superimposed to determine surface temperature on any day as a function of heat in storage; each past addition or loss is parameterized by age. Turnovers (convective mixing of deep lower-density waters with surface waters as surface temperature passes through that at maximum density) can occur as a fundamental behavior of this superposition model and hysteresis between heat in storage and surface temperature, observed during the heating and cooling cycles on the lakes, is preserved. Water surface temperature becomes (Croley, 1992a):

$$T_k = 3.98°\mathrm{C} + \sum_{m=1}^{k} f_{k,m}\left(\underset{m\le n\le k}{\mathrm{MIN}} H_n - \underset{m-1\le n\le k}{\mathrm{MIN}} H_n\right) \tag{38}$$

where T_k' = water surface temperature, H_k' = heat storage in the lake k days after the last turnover, and $f_{k,m}$ is a "wind-aging" function, defined subsequently, relating surface temperature rise on day k to heat added on day m. Ice surface temperature relates to ice pack heat storage here as:

$$H_k' = \rho\, C_i\, V_k\, T_k'/2 - \rho\,\gamma_f\, V_k' \tag{39}$$

where T_k' = ice surface temperature on day k, H_k' = heat storage in the ice pack on day k, ρ = density of ice, C_i = specific heat of ice, V_k = volume of the ice pack on day *k*, and V_k' = volume of ice formed by freezing or melting on day *k*. The boundary conditions on water surface temperature and volume of the ice pack for every day (dropping the daily subscript) are:

$$V = 0, \qquad T \geq 0° \text{ C} \tag{40}$$

$$T = 0° \text{ C}, \qquad V \geq 0 \tag{41}$$

These equations are satisfied by selecting the heat flux between the water and ice, Q_w, appropriately. Q_w, if negative, is yielded as ice forms (to keep water surface temperature from going below freezing) and, if positive, is used in melting ice (to keep water surface temperature at freezing as long as there is ice present). The boundary conditions on ice surface temperature and volume of the ice pack for every day (dropping the daily subscript) are:

$$T' = T_a, \qquad V > 0 \text{ and } T_a \leq 0° \text{C} \tag{42}$$

$$T' = 0° \text{ C}, \qquad V = 0 \text{ or } T_a > 0° \text{C} \tag{43}$$

where T_a = over-ice air temperature. The volume of the ice pack, *V*, and the volume of ice formed by freezing or melting, *V*', are related:

$$\frac{\partial V}{\partial t} = \frac{\partial V'}{\partial t} + S - E \tag{44}$$

where *S* = volumetric rate of snow falling on the ice, and *E* = volumetric rate of evaporation from the ice. The "wind-aging" function, $f_{k,m}$, is:

$$\begin{aligned} f_{k,m} &= \frac{2 - M_{k,m}/F}{\rho_w C_w M_{k,m}}, && M_{k,m} < MIN\left(F, \frac{2V_C}{1+V_C/F}\right) \\ &= \frac{1}{\rho_w C_w M_{k,m}}, && MIN\left(F, \frac{2V_C}{1+V_C/F}\right) \leq M_{k.m} < MAX\left(V_c, \frac{2V_C}{1+V_C/F}\right) \\ &= \frac{1}{\rho_w C_w V_c}, && MAX\left(V_c, \frac{2V_C}{1+V_C/F}\right) \leq M_{k,m} \end{aligned} \tag{45}$$

where V_c = volume (capacity) of the lake and $M_{k,m}$ = mixing volume size in the lake, on day *k*, of the heat added on day *m* (a function of accumulated wind movement, W_j, from day *m* through day *k*),

$$M_{k,m} = V_e\left[1 + a\exp\left(-b\sum_{j=m}^{k} W_j\right)\right]^{-1} \tag{46}$$

Also, *a*, *b*, *F*, and V_e = empirical parameters to be determined in a calibration to observed data. V_e is interpreted as the "equilibrium" volume approached as a limit (in a sufficiently deep lake) since the effects of wind mixing at the surface diminish with distance from the surface. *F* is interpreted as the mixing volume at which a heat addition is fully mixed throughout. Parameters *a*, *b*, and *F* are defined for water temperatures above 3.98°C ("turnover" temperature of water at maximum density) and are replaced by *a*', *b*', and *F*', respectively, for water temperatures below 3.98°C. Details for the flux terms in (36) and (37) are presented by Croley (1989a,b). Derivation details of (38), (45), and (46) are available elsewhere (Croley, 1992a).

4.3.3 Ice Pack Growth

In (39), linear vertical temperatures are used through the ice pack from *T*' on the surface to 0°C on the bottom, similar to Green and Outcalt (1985). Differentiating (4) and ignoring small terms,

$$\frac{\partial H'}{\partial t} \cong \tfrac{1}{2}\rho C_i V \frac{\partial T'}{\partial t} - \rho\gamma_f \frac{\partial V'}{\partial t} \tag{47}$$

Thus, the heat change is split between a temperature change in the ice pack and a volume change due to melting or freezing. Comparing (37) and (47), note the temperature change in (47) is taken here as resulting from a portion of the heat added from (or lost to) the atmosphere [$A\,(q_i - q_r' - q_e' + q_h' + q_p')$]. The remainder of that heat is identified as Q_a:

$$Q_a = A\left(q_i - q_r' - q_e' + q_h' + q_p'\right) - \frac{1}{2}\rho C_i V \frac{\partial T'}{\partial t} \tag{48}$$

This heat (Q_a) and all of the heat added from the water body, Q_w, then result in changes to the ice pack volume (freezing or melting); from (37), (44), (47), (48):

$$Q_a + Q_w = -\rho\gamma_f \frac{\partial V'}{\partial t} = -\rho\gamma_f\left(\frac{\partial V}{\partial t} - S + E\right) \tag{49}$$

Consider a prismatic ice pack with surface area A and depth (or thickness) *D;* see Figure 20. The heat exchange between the atmosphere and the ice pack available for freezing or melting, Q_a, is taken as resulting in either melt (along the entire atmosphere-ice surface) or freezing (along the entire water-ice surface). The heat exchange between the water body and the ice pack, Q_w, is taken as resulting in changes along only the water/ice surface (either melt or freezing). After simplification (Croley and Assel, 1993),

$$\frac{\partial D}{\partial t} = \left\{-\frac{Q_a}{A + x_w D} I_{(-\infty,0]}(Q_a) - \frac{Q_a}{A + x_a D} I_{(0,\infty)}(Q_a) - \frac{Q_w}{A + x_w D}\right\}\frac{1}{\rho\gamma_f} + \frac{S}{A} - \frac{E}{A + x_a D} \tag{50}$$

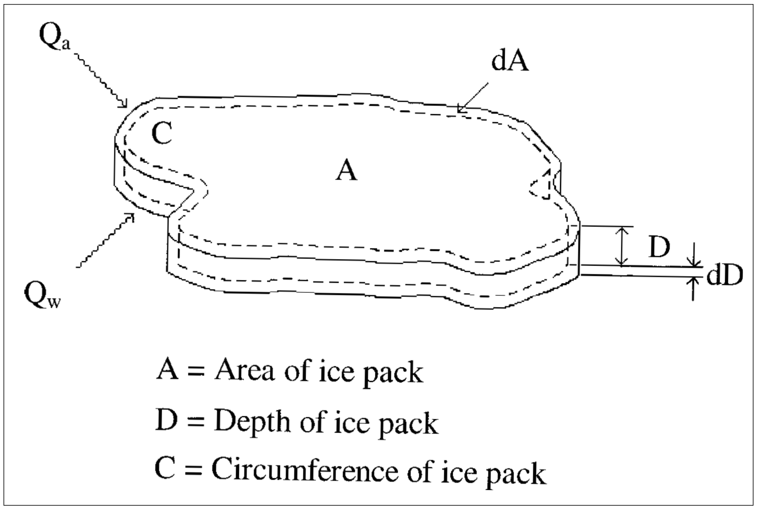


Figure 20.--Conceptual Prismatic Ice Pack.

$$\frac{\partial A}{\partial t} = \left\{ -\frac{x_w Q_a}{A + x_w D} I_{(-\infty,0]}(Q_a) - \frac{x_a Q_a}{A + x_a D} I_{(0,\infty)}(Q_a) - \frac{x_w Q_w}{A + x_w D} \right\} \frac{1}{\rho \gamma_f} - \frac{x_a E}{A + x_a D} \quad (51)$$

$$x_a = \tau_a A^{1/2} \quad (52)$$

$$x_w = \tau_w A^{1/2} \quad (53)$$

where $I_{(.,.)}(x)$ = indicator function (equal to unity if the quantity in parentheses, x, is within the indicated interval and equal to zero if not), τ_a and τ_w = empirical coefficients depending upon ice pack shape, the ratios of vertical to lateral changes along the atmosphere-ice interface and along the water-ice interface, and the buoyancy of ice. The change in total ice volume is, from (50) and (51):

$$\begin{aligned} \frac{\partial V}{\partial t} &= \frac{\partial (A\,D)}{\partial t} = A \frac{\partial D}{\partial t} + D \frac{\partial A}{\partial t} \\ &= (-Q_a - Q_w) \frac{1}{\rho \gamma_f} + S - E \end{aligned} \quad (54)$$

Note, (49) and (54) agree.

Equations (36)–(48), (50)–(54), and those for the component fluxes (21)–(35), may be solved simultaneously to determine the heat storage, the water and ice surface temperatures, and the ice pack extents. The Lake Evaporation and Thermodynamics Model is pictured schematically in Figure 21.

4.3.4 Calibration Procedure

Two calibrations are involved in applying the model in a particular setting. The first determines the first eight parameters (a, b, F, a', b', F', V_e, and h). The first seven parameters relate to superposition heat storage (Croley, 1992a) and the eighth parameter, h, reflects the effect of cloudiness on the atmospheric

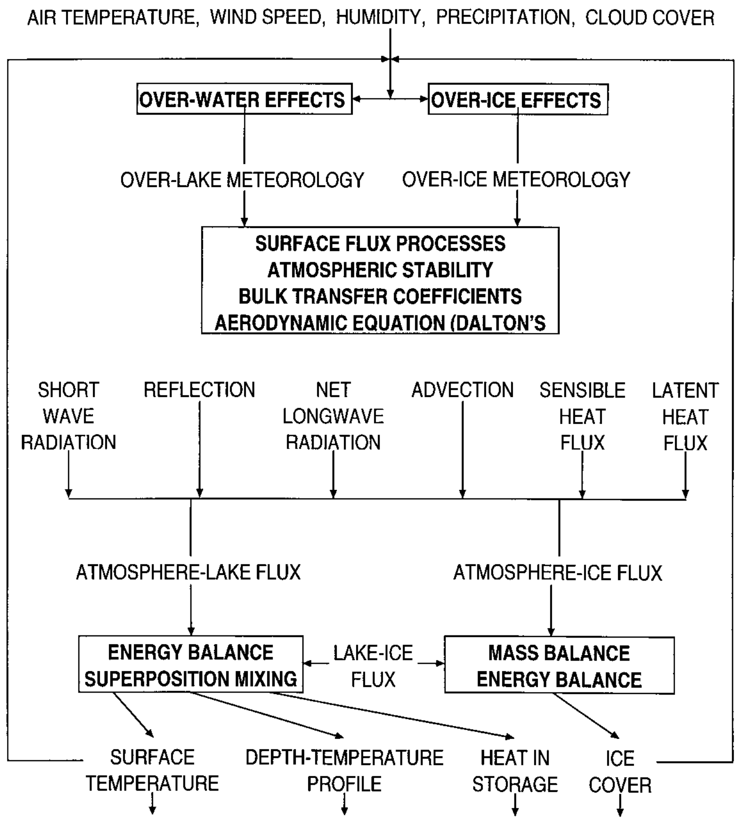


Figure 21.--Lake Evaporation and Thermodynamics Model Conceptual Schematic.

net long-wave radiation exchange (Croley, 1989a,b). This calibration minimizes daily water surface temperature root mean square error (RMSE) by using methods described elsewhere (Croley and Hartmann, 1984). Meteorology data for 1948–1985 and water surface temperature data on each of the Great Lakes, except Lake Michigan, were taken from airplane and satellite measurements, extended through August 1988, and prepared as described by Croley (1989a,b). Water surface temperature data for Lake Michigan from 1981 through 1985 were gleaned from areal maps prepared at the National Weather Service's Marine Predictions Branch (B. Newell, personal communication, 1990) and extended through August 1988 also. The second calibration determines the two parameters (τ_a and τ_w) that minimize daily ice cover RMSE with these same calibration techniques. Lake-averaged ice cover for model calibration was calculated from GLERL's digital ice cover data base (Assel, 1983). In most cases, less than 100% of a lake was observed on any given date. If less than 70% of the Lake Superior surface was observed, the ice cover for that date was not included in the model calibration. A subjective estimate of lake-averaged ice cover was made for the other Great Lakes if the data were insufficient.

Parameters are determined, in both cases, in automated systematic searches of the parameter spaces to minimize the RMSE between simulated and model outputs. Each parameter, selected in rotation, is searched until all parameter values converge to four digits, instead of searching only until the RMSE stabilizes. This simple search algorithm does not give unique optima for calibrated parameter sets because of synergistic relationships between parameters that allow parameter compensations to occur. However, the model concepts have been carefully chosen so that the parameters have physical significance; this allows them to be interpreted in terms of the thermodynamics they represent. Initialization of the model corresponds to identifying values from field conditions that may be measured; interpretations of a lake's thermodynamics then can aid in setting both initial and boundary conditions.

Prior to calibration or model use, the (spatial) average temperature-depth profile in the lake and the ice cover must be initialized. While the ice cover is easy to determine as zero during major portions of the year, the average temperature-depth profile in the lake is generally difficult to determine. If the model is to be used in forecasting or for short simulations, then it is important to determine these variables accurately prior to use of the model. If the model is to be used for calibration or for long simulations, then the initial values are generally unimportant. The effect of the initial values diminishes with the length of the simulation, and after 2–3 years of simulation, the effects are nil from a practical point of view.

Empirical coefficients of the evaporation, heat storage, and ice sub-models were calibrated in an iterative process that used the two calibrations sequentially in rotation. We used independent data (lake-averaged daily surface temperature for the lake thermodynamics and heat storage sub-models and lake-averaged daily ice cover for the lake ice cover sub-model). First we minimized the RMSE of daily water surface temperature by calibrating lake thermodynamics model parameters and holding the parameters for the ice cover sub-model constant. We then held lake thermodynamics model parameters constant and calibrated the parameters of the ice cover sub-model to minimize the RMSE of daily ice cover. Then we repeated the process until the RMSEs for both water surface temperatures and ice cover were not significantly reduced from the previous iteration.

4.3.5 Application

The results of the parameter calibration, as well as a few statistics on each of the Great Lakes, are summarized in Table 13. Statistics from the calibration and from an independent verification period are presented in Table 14. Turnovers (convective mixing of deep lower-density waters with surface waters as surface temperature passes through that at maximum density) occur as a fundamental behavior of GLERL's thermodynamic and heat storage model. Hysteresis between heat in storage and surface temperature, observed during the heating and cooling cycles on the lakes, is preserved. The model also correctly depicts lake-wide seasonal heating and cooling cycles, vertical temperature distributions, and other mixed-layer developments. There is good agreement between the actual and calibrated-model water surface temperatures; the RMSE is between 1.1 and 1.6°C on the large lakes [within 1.1–1.9°C for an independent verification period, 1966–79 (Croley, 1989a,b, 1992a)]. The RMSE for ice concentrations is between 12 and 23% for the joint calibration-verification period. There is also good agreement with 8 years of bathythermograph observations of depth-temperature profiles on Lake Superior, and 1 year of

Table 13.--Lake Evaporation and Thermodynamics Model Constants and Parameters.

	Lake					
	Superior	Michigan	Huron	Georgian	Erie	Ontario
Surface area, km^2	82,100	57,800	40,640	18,960	25,700	18,960
Volume, km^3	12,100	4,920	2,761	779	484	1,640
Average depth, m	147	85.1	67.9	41.1	18.8	86.5
a	$6.298 \times 10^{+0}$	$7.290 \times 10^{+0}$	$6.460 \times 10^{+0}$	$1.585 \times 10^{+0}$	$2.820 \times 10^{+0}$	$7.710 \times 10^{+0}$
b, m^{-1} s	3.298×10^{-3}	2.599×10^{-3}	2.810×10^{-3}	5.473×10^{-3}	5.430×10^{-3}	2.800×10^{-3}
F, km^3	$3.273 \times 10^{+3}$	$5.100 \times 10^{+2}$	$4.890 \times 10^{+3}$	$1.101 \times 10^{+3}$	$1.000 \times 10^{+2}$	$2.000 \times 10^{+2}$
a'	$2.019 \times 10^{+0}$	$1.158 \times 10^{+0}$	$3.829 \times 10^{+0}$	$1.471 \times 10^{+0}$	$2.610 \times 10^{+0}$	$4.000 \times 10^{+0}$
b', m^{-1} s	3.795×10^{-3}	2.301×10^{-3}	3.890×10^{-3}	1.103×10^{-2}	5.600×10^{-3}	5.110×10^{-3}
F', km^3	$5.113 \times 10^{+3}$	$4.000 \times 10^{+3}$	$6.789 \times 10^{+3}$	$8.943 \times 10^{+2}$	$1.000 \times 10^{+2}$	$4.600 \times 10^{+2}$
V_e, km^3	$1.200 \times 10^{+4}$	$5.006 \times 10^{+3}$	$8.010 \times 10^{+3}$	$9.748 \times 10^{+2}$	$8.490 \times 10^{+2}$	$2.000 \times 10^{+3}$
h	$1.299 \times 10^{+0}$	$1.068 \times 10^{+0}$	$1.150 \times 10^{+0}$	$1.223 \times 10^{+0}$	$1.290 \times 10^{+0}$	$1.200 \times 10^{+0}$
t_a	$9.011 \times 10^{+8}$	$9.001 \times 10^{+8}$	$9.119 \times 10^{+8}$	$9.279 \times 10^{+8}$	$9.988 \times 10^{+8}$	$9.010 \times 10^{+8}$
t_w	$8.002 \times 10^{+5}$	$2.003 \times 10^{+5}$	$1.080 \times 10^{+6}$	$4.437 \times 10^{+5}$	$9.202 \times 10^{+5}$	$8.001 \times 10^{+4}$

Table 14.--Lake Evaporation and Thermodynamics Model Calibration Statistics.

	Lake					
	Superior	Michigan	Huron	Georgian	Erie	Ontario
	CALIBRATION PERIOD STATISTICS Water Surface Temperatures (1980-1988)[a]					
Means Ratio[b]	1.00	1.01	0.98	1.01	1.03	0.99
Variances Ratio[c]	1.01	0.98	0.95	1.02	1.08	0.99
Correlation[d]	0.98	0.97	0.98	0.99	0.99	0.98
R.M.S.E.[e]	1.13	1.56	1.33	1.10	1.58	1.43
	Ice Concentrations (1960-1988)[f]					
Means Ratio[g]	0.92	0.72	0.70	0.98	1.15	0.39
Variances Ratio[h]	1.24	1.02	1.67	1.62	1.09	0.63
Correlation[i]	0.76	0.83	0.73	0.77	0.89	0.54
R.M.S.E.[j]	23.4	12.4	26.0	21.5	19.0	15.4
	VERIFICATION PERIOD STATISTICS Water Surface Temperatures (1966-1979)[k]					
Means Ratio[b]	0.96		1.03	0.98	1.05	0.94
Variances Ratio[c]	1.10		0.95	1.00	1.10	0.97
Correlation[d]	0.97		0.99	0.98	0.98	0.96
R.M.S.E.[e]	1.09		1.10	1.34	1.91	1.92

[a]Data between 1 January 1980 and 31 August 1988 for all lakes except Michigan and between 1 January 1981 and 31 August 1988 for Lake Michigan, with an initialization period for all lakes except Georgian Bay starting 1 January 1948 and 1 January 1953 for Georgian Bay.
[b]Ratio of mean model surface temperature to data mean.
[c]Ratio of variance of model surface temperature to data variance.
[d]Correlation between model and data surface temperature.
[e]Root-mean-square-error between model and data surface temperatures in degrees C.
[f]Data between 1 January 1960 and 31 August 1988 for all Great Lakes except Superior and between 1 March 1963 and 31 August for Lake Superior, with an initialization period for all lakes starting 1 January 1958.
[g]Ratio of mean model ice concentration to data mean.
[h]Ratio of variance of model ice concentration to data variance.
[i]Correlation between model and data ice concentration.
[j]Root-mean-square-error between model and data ice concentrations in %.
[k]Data between 1 January 1966 and 31 December 1979 for all lakes except Michigan with an initialization period for all lakes except Georgian Bay starting 1 January 1948 and 1 January 1953 for Georgian Bay.

independently-derived weekly or monthly surface flux estimates on Lakes Superior, Erie, and Ontario (2 estimates).

4.3.6 Calibration Issues

There were several problems in calibrating the model. First, it appears that the models are close to being over-specified in terms of the number of parameters used; i.e., there appear to be almost too many degrees of freedom allowed for the data sets used in the calibrations. The result is that the optimums are not unique and it is not possible to determine meaningful values of any additional parameters. Parameter compensation exists so that changes in one parameter can be offset by changes in other parameters with little change in the RMSE of the calibration. This made it difficult to determine an ice break-up model, not presented here, which had an additional three parameters. We had considered ice breaking and rejoining by developing a differential equation for the rate of change with time of the number of ice pieces as a function of wind stress, melting, and refreezing. We could not meaningfully calibrate this addition to the ice model with the ice cover data sets we had, and so we eliminated ice break-up from the model presented here. Perhaps when other parameters are reduced through model reformulations in calibrations at a later date, it will be possible to model and calibrate for ice break-up in a meaningful manner.

Second, optimizing parameters with regard to two objectives (minimizing RMSEs associated with water surface temperatures and ice cover) does not produce the same parameter sets. There seems to be a trade-off between the two objectives at times and RMSE of water temperatures decreases at the expense of ice cover RMSE and vice-versa.

The model has 10 parameters calibrated to match water surface temperatures and ice cover. Seven of them are defined in the superposition heat storage submodel. The number of empirical model parameters could perhaps be reduced by use of other one-dimensional mixed-layer heat-storage models (McCormick and Meadows, 1988; Hostetler and Bartlein, 1990). The critical limitation of such models for long-term hydrological forecasting and simulation is the lack of representative or accurate hourly hydrometeorological data over long periods. Secondarily, computer time can be excessive for such models in forecast or multi-year simulation environments.

4.4 Models Validity and Applicability

Although GLERL uses a daily resolution of data with their models, basin-wide processes of runoff, over-lake precipitation, and lake evaporation (described with models here) respond discernibly to weekly changes at best, and monthly is usually adequate for net supply and lake level simulation (this ignores short-term fluctuations associated with storm movement, which are not addressed in this study). Likewise, spatial resolution finer than about 1000–5000 km^2 (the present average resolution of GLERL's models and their applications) is unnecessary, for use with general circulation models (GCMs) of the atmosphere, and much can be done in assessing hydrology changes at resolutions of 100,000–1,000,000 km^2 with lumped versions of the models. This coarse spatial resolution is still much finer than present GCM grids.

The models were assessed partially by computing net basin supplies to the lakes (basin runoff plus over-lake precipitation minus over-lake evaporation) with historical meteorological data for 1951–80 and comparing to historical net basin supplies. The absolute average annual difference ranged from 1.6% to 2.7% on the deep lakes, while the Lake St. Clair and Lake Erie applications were 12.0% and 7.0% respectively; month-to-month differences showed more variation. These differences generally reflect poorer evaporation modeling on the shallow lakes and snowmelt and evapotranspiration model discrepancies for the other lake basins. While monthly differences were generally small, a few were significant. The low annual residuals were felt to be acceptable for use of these models in assessing changes from the current climate as they would be consistently applied to both a “present” and a “changed” climate. Further assessment of model deficiencies with comparisons to historical net basin supplies is difficult since the latter are derived from water budgets which incorporate all budget term errors in the derived net basin supplies.

There is some indication of model applicability outside of the time periods over which the models were calibrated as indicated above and in Tables 12 and 14. To assess the applicability of the process models to a climate warmer than the one under which they were calibrated and verified requires access to meteorological data and process outputs for the warmer climate, which unfortunately do not exist. Warm periods early in this century are not sufficiently documented for the Great Lakes. In particular, data are lacking on watershed runoff to the lakes, water surface temperatures, wind speed, humidity, cloud cover, and solar insolation.

It is entirely possible that the models are tied somewhat to the present climate; empiricism is employed in the evapotranspiration component of the LBRM and in some of the heat flux terms in the heat balance and lake evaporation model. Coefficients were determined or selected in accordance with the present climate. The models are all based on physical concepts that should be good under any climate; however, the assumption is made that they represent processes under a changed climate that are the same as the present ones. These include linear reservoir moisture storages, partial-area infiltration, lake heat-storage relations with surface temperature, and gray-body radiation. However, the calibration and verification periods for the component process models include a range of air temperatures, precipitation, and other meteorological variables that encompass much of the changes in these variables predicted for a changed climate. Even though the changes are transitory in the calibration and verification period data sets, the models appear to work well under these conditions.

5. GREAT LAKES CLIMATE CHANGE HYDROLOGICAL RESPONSES

GLERL integrated their hydrological process models into a system to estimate lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1992b; Croley and Hartmann, 1987; Croley and Lee, 1993). As mentioned earlier, they used this system to simulate the Great Lakes hydrology for historical meteorology and five transposed scenarios (scenarios 1 through 4 and scenario 5, which is scenario 3 corrected for lake effects). Behavior is characterized by looking at mean annual and seasonal values of each hydrological variable under each of the five climates tested, as well as the base case. Selected measures of the variability of each hydrological variable, for each of the five climates tested, also were calculated for annual and seasonal periods. These means and measures of variability are compared to those determined with the historical meteorology (which serves as a baseline for assessing shifts produced by other regimes). Seasonal steady-state behavior is exemplified here in figures for the Lake Superior basin and scenario 2, and summarized for all lakes and all climate-change scenarios for the entire period in annual tables.

5.1 Basin Meteorology

The annual cycles, of all meteorological variables, were averaged over the 1951–90 period and inspected. The annual air temperatures for the base case increase with decreasing latitude; see Table 15. The over-land air temperatures for all five transposed scenarios are higher than the base case throughout the annual cycle. The differences are greatest for the southern-most scenarios (3, 4, and 5) and for the northern-most lakes; see Table 15. The difference is smallest during the late summer or fall to early winter and largest during the late winter to early spring for all lakes and for all transposed scenarios; as an example, see Figure 22 for scenario 2 on the Superior basin. Changes in annual variability of air temperature were remarkably small. Table 16 shows the average annual steady state standard deviation of air temperature, depicting the variability from year to year in the 40-year period. The annual air temperature variability in Table 16 appears artificial and is the result of truncation of annual air temperatures to the nearest 0.1ºC before the standard deviation was calculated. Variability also changes little throughout the seasonal cycle on all lakes and for all scenarios; as an example see Figure 23 for scenario 2 on Lake Superior. The seasonal patterns remain with more variability associated with cooler temperatures.

Table 15.--Average Annual Steady-State Basin Meteorology Differences.

Basin	Overland Air Temperature (ºC) & Transferred Climate Absolute Differences[a]						Overland Precipitation (mm) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	2.3	6.9	6.8	10.4	10.9	10.4	817	-23%	6%	-20%	21%	-20%
Michigan	7.2	6.3	5.6	9.8	9.4	10.1	828	3%	39%	1%	59%	4%
Huron	7.1	5.8	4.6	9.8	9.1	10.3	813	26%	40%	48%	70%	51%
Georgian	4.3	7.0	5.7	10.4	9.8	10.6	908	2%	10%	30%	47%	31%
St. Clair	8.3	5.2	3.9	9.3	8.9	9.4	854	28%	33%	51%	61%	53%
Erie	9.1	6.1	4.4	9.4	8.2	9.4	913	31%	44%	37%	55%	39%
Ontario	7.2	6.2	6.5	9.3	9.7	9.3	934	26%	18%	49%	33%	49%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Table 16.--Average Annual Steady-State Basin Meteorology Variability Differences.

Basin	Overland Air Temperature Std. Dev. (ºC) & Transferred Climate Relative Change[a]						Overland Precipitation Std. Dev. (mm) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	0.80	-13%	-13%	-13%	-13%	-13%	83.8	27%	41%	52%	110%	52%
Michigan	0.60	17%	0%	0%	0%	0%	93.8	84%	64%	64%	124%	64%
Huron	0.60	17%	17%	0%	0%	-17%	89.1	127%	52%	153%	154%	153%
Georgian	0.70	0%	-14%	-14%	-14%	-14%	93.8	76%	17%	114%	101%	114%
St. Clair	0.60	17%	17%	0%	0%	0%	121.9	80%	22%	106%	90%	106%
Erie	0.60	0%	0%	-17%	0%	-17%	109.7	99%	51%	105%	103%	105%
Ontario	0.60	0%	-17%	0%	0%	0%	89.6	99%	71%	149%	96%	149%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Over-land precipitation shows much more variability than air temperature both among scenarios and among lake basins. Table 15 shows that, generally, precipitation is greater on all lakes except Superior for all five scenarios. The western-most scenarios (1, 3, and 5) generally increase precipitation much less than do the eastern-most scenarios (2 and 4) on the Superior, Michigan, and Erie basins. On the Lake Superior basin, the western most scenarios (1, 3, and 5) actually drop precipitation, relative to the base case, while the eastern-most raise precipitation. On the Huron, Georgian Bay, St. Clair, and Ontario basins, the southern-most scenarios (3, 4, and 5) show the largest precipitation increases of all the scenarios.

Throughout the annual cycle, all basins show a shift in seasonal precipitation to earlier peaks for all scenarios; as an example, see Figure 22 for scenario 2 on the Superior basin. Superior and Michigan maximum precipitation shifts from August and September to May or June, for all scenarios. Huron and Georgian Bay peaks shift from September to March through July, depending on the scenario. St. Clair, Erie, and Ontario peaks shift from June (Erie) or August (St. Clair and Ontario) to March through June, depending on the scenario.

Changes in annual variability of precipitation are also more pronounced than for air temperature; see Table 16. Generally, the southern-most scenarios, which are the wettest in Table 15, also are the more

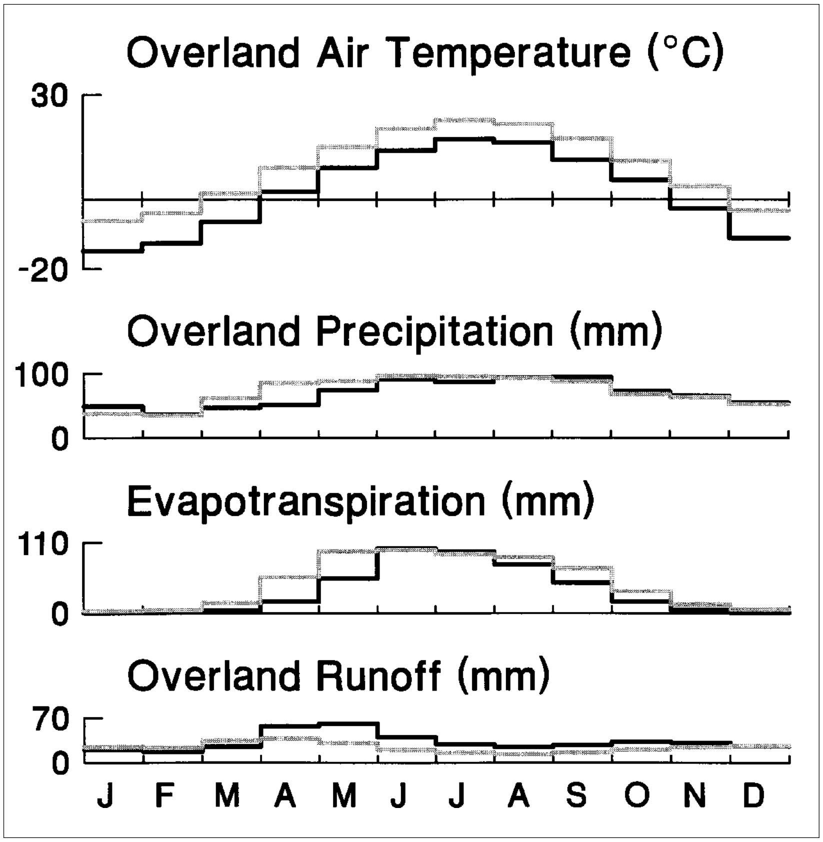


Figure 22.--Seasonal Lake Superior Basin Average Meteorology and Hydrology for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

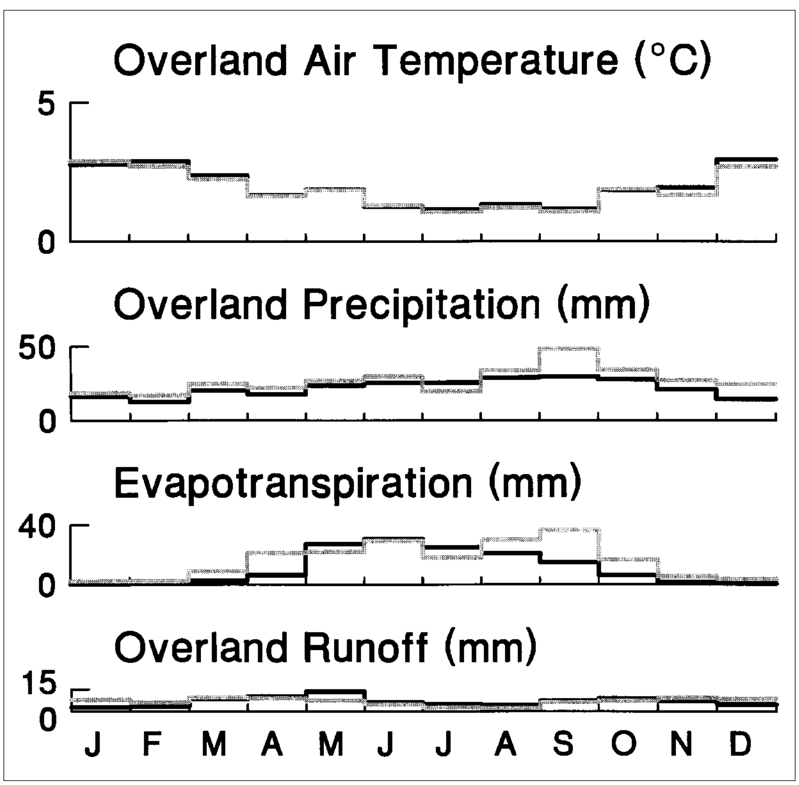


Figure 23.--Seasonal Lake Superior Basin Meteorology and Hydrology Standard Deviations for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

variable as seen in Table 16. This is expected since if precipitation is generally closer to its lower bound of zero, its variation must therefore be diminished too. The variation, as depicted by the standard deviation in Table 16, is greater than 100% for most of the southern-most scenarios (3, 4, and 5).

Seasonally, the variability of precipitation generally is high in the late summer and early fall; see Figure 23. This is similar to the base case pattern. For all scenarios except scenario 2, all lakes exhibit relatively lower variability in August with higher variability both before and after August. This local minimum in August may not be the smallest over the annual cycle, but the pattern is very pronounced. For scenario 2, this pattern is not so evident; no seasonal pattern of variability appears generally evident that applies to all lakes. Only scenario 2 involved no move east; the other climate scenarios were transposed from the west by some amount. The pattern of variability in these scenarios is consistent with the known spatial distribution of precipitation variability. For instance, Griffiths and Driscoll (1982, Fig. 7.22) show that the current Great Lakes region experiences a relative minimum of precipitation variability. Annual precipitation variability increases to the west and south of the Great Lakes.

5.2 Basin Hydrology

The increased air temperatures, consequent in all transposed climates, significantly alters the heat balance of the surface hydrology. As seen in Table 17, the snow pack is almost completely eliminated as the relative change varies among scenarios and lake basins from a 86% to a 99% drop in accumulated snow moisture. Furthermore, evapotranspiration increased significantly from 9% to 89%. For the eastern basins (Huron, Georgian Bay, St. Clair, Erie, and Ontario), the greatest increases in evapotranspiration and the greatest decreases in snowpack correspond to the southern-most transposed climates (3, 4, and 5), which have the greatest increases in average annual steady-state air temperatures. For the western basins (Superior and Michigan), the greatest increases in evapotranspiration come from the eastern-most scenarios (2 and 4), which have the greatest increases in precipitation for those basins. This implies that the eastern basins have moisture-limiting evapotranspiration where the effects of relatively less water availability (as reflected by precipitation) limit evapotranspiration more than in the western basins. The increased evapotranspiration and decreased snowpack give rise to less moisture available in the soil and groundwater zones. Table 17 shows a general lowering of soil moisture that is most acute for the western-most scenarios (1, 3, and 5) and a corresponding loss of groundwater storage in the same pattern. By adding snow water equivalent, soil moisture, groundwater, and surface storage (not shown in Table 17), the total moisture storage is computed as in Table 17. The pattern is the same there; the western-most scenarios show the most acute loss of moisture storage in the basin, but all scenarios show a general loss as compared to the base case. Note that the anomalously high groundwater (and consequently total moisture storage) for Georgian Bay in Table 17 is a result of unrealistic initial groundwater conditions used in the models to be consistent with earlier calibrations; recall they are arbitrary. On the other basins, estimated groundwater is much faster, compared to the Georgian Bay calibrations, and initial conditions are unimportant.

The net effect of the increased air temperatures, through increased evapotranspiration and decreased moisture storage in the basins, would be decreased runoff. While Table 17 does indeed show decreased runoff in many cases, there are other cases where runoff increases since moisture reductions are offset by precipitation increases. Recall that precipitation is generally higher under all of the transposed climates. Again, Table 17 shows the greatest runoff decreases for the northern-most basins and the western-most transposed scenarios.

Figure 22 depicts typical seasonal behavior for evapotranspiration and runoff. It appears that for all lake basins, for the western-most scenarios (1, 3, and 5), evapotranspiration has shifted earlier in the

Table 17.--Average Annual Steady-State Basin Hydrology Differences.

Basin	Snow Water Equivalent (mm) & Transferred Climate Relative Changes[a]						Soil Moisture (mm) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	50.6	-92%	-86%	-98%	-97%	-98%	41.4	-62%	-32%	-66%	-44%	-66%
Michigan	11.7	-91%	-87%	-98%	-97%	-98%	34.7	-46%	6%	-55%	0%	-55%
Huron	14.1	-90%	-87%	-98%	-98%	-99%	28.6	-33%	6%	-33%	-4%	-37%
Georgian	37.6	-94%	-93%	-98%	-98%	-98%	70.5	-48%	-23%	-35%	-16%	-34%
St. Clair	8.5	-87%	-90%	-98%	-99%	-98%	6.0	-32%	-9%	-45%	-38%	-46%
Erie	5.7	-89%	-88%	-98%	-96%	-98%	6.7	-40%	-17%	-55%	-38%	-55%
Ontario	15.7	-92%	-96%	-99%	-99%	-99%	20.8	-28%	-29%	-28%	-34%	-28%

Basin	Groundwater (mm) & Transferred Climate Relative Changes[a]						Total Basin Moisture (mm) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	148	-62%	-35%	-66%	-41%	-66%	297	-66%	-41%	-71%	-50%	-70%
Michigan	62	-45%	5%	-55%	-7%	-54%	115	-49%	-3%	-58%	-12%	-57%
Huron	146	-23%	10%	-17%	19%	-21%	193	-29%	3%	-25%	8%	-28%
Georgian	26857	-2%	-1%	-1%	0%	-1%	26977	-3%	-1%	-2%	0%	-2%
St. Clair	10	7%	26%	16%	26%	17%	28	-29%	-19%	-30%	-25%	-30%
Erie	9	-4%	-6%	-9%	-5%	-9%	24	-30%	-20%	-40%	-29%	-40%
Ontario	11	-27%	-27%	-21%	-23%	-21%	61	-37%	-42%	-36%	-48%	-36%

Basin	Overland Evapotranspiration (mm) & Transferred Climate Relative Changes[a]						Runoff as an Overland Depth (mm) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	423	9%	37%	19%	69%	19%	394	-57%	-27%	-62%	-31%	-61%
Michigan	506	22%	48%	22%	77%	26%	322	-28%	23%	-34%	29%	-29%
Huron	499	48%	52%	75%	89%	79%	314	-9%	21%	5%	39%	5%
Georgian	482	43%	40%	76%	89%	77%	418	-37%	-20%	-20%	0%	-18%
St. Clair	538	37%	41%	64%	77%	66%	315	13%	20%	29%	35%	30%
Erie	569	34%	41%	49%	66%	51%	344	26%	48%	17%	36%	19%
Ontario	473	52%	48%	88%	87%	88%	461	-1%	-14%	9%	-22%	9%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

seasonal cycle. The bulk of the annual evapotranspiration and the peak evapotranspiration occur earlier. This also appears true for the northern-most basins for scenarios 2 and 4; for the other basins in scenarios 2 and 4, while evapotranspiration increases, the seasonal pattern is not significantly changed. These shifts are due to the loss of snow moisture storage; the basins therefore have more soil moisture available for evapotranspiration during the winter than in the base case; see Figure 24. This affects the seasonal distribution of runoff as well. The shift in the seasonal peak runoff earlier in Figure 22 for scenario 2 on the Superior basin is typical of the behavior on all basins for all scenarios. The shift in runoff is therefore far more consistent across all basins and all scenarios than is the shift in evapotranspiration. Again, this results from the loss of snow moisture storage; basins have more winter runoff than the base case, contributing to the runoff shift.

Figure 24.--Seasonal Lake Superior Basin Average Moisture Storage for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

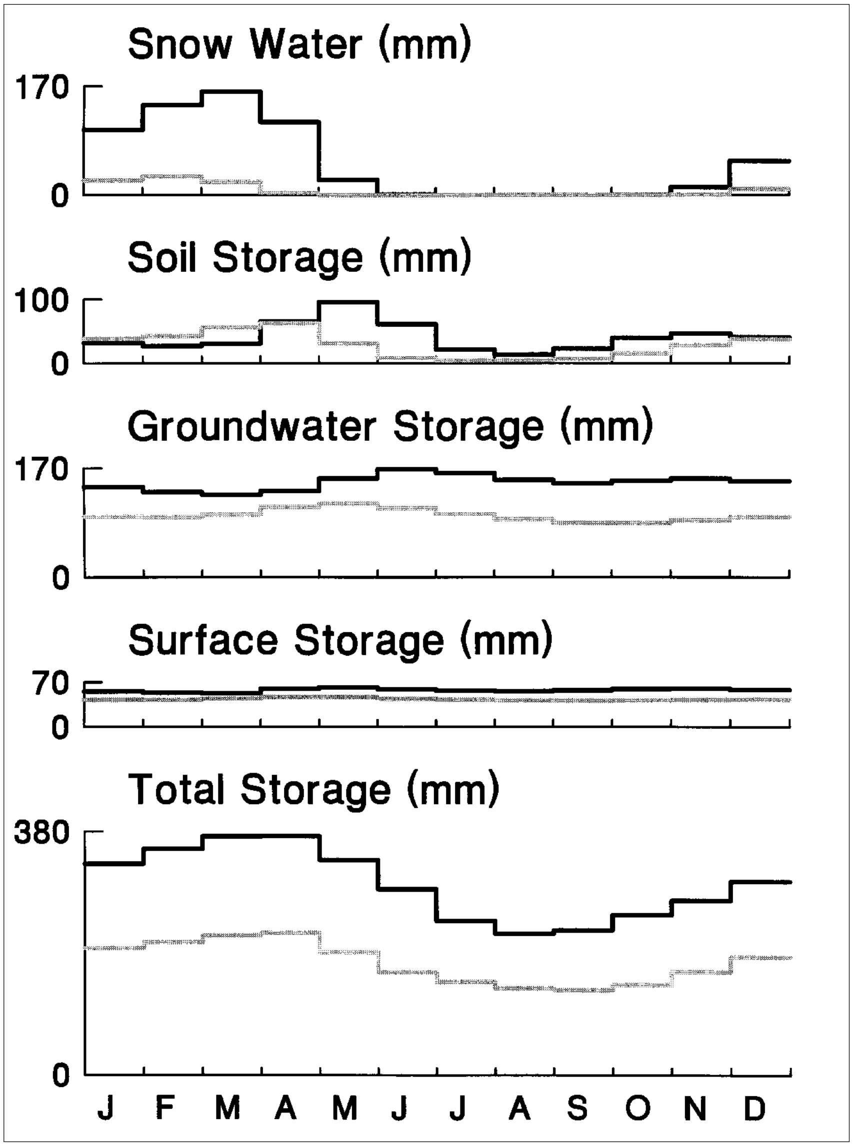


Table 18 and Figure 25 show expected changes in variability for basin moisture storage variables. Snow water variability is greatly decreased simply because snow water is greatly decreased toward its lower bound of zero. Relative changes in variability of soil moisture, groundwater, surface storage, and total basin moisture are small (less than 100%), with respect to both annual and seasonal values. Table 18 and Figure 23 also show evapotranspiration with more variability, generally, for the southern-most scenarios. This corresponds the greatest and most variable precipitation; see Tables 15 and 16 , respectively. Again, this results from the fact that evapotranspiration is a moisture-limited process; only the amount in storage can evaporate or transpire and where there is more variability in the moisture supply will there be more variability in the evapotranspiration amounts. There generally is not much change in basin runoff variability in Table 18 and Figure 25. There does appear to be a slight increase in runoff variability during the winter for almost all basins and scenarios, as in Figure 25 for scenario 2 on the Superior basin, corresponding to the absence of the snowpack and consequent runoff during the wintertime. The greatest consistent change in variability across all scenarios in both evapotranspiration and runoff, occur on Ontario, exposed to the most-eastern part of each scenario with its most variable precipitation.

Table 18.--Average Annual Steady-State Basin Hydrology Variability Differences.

Basin	Snow Water Equivalent Std. Dev. (mm) & Transferred Climate Relative Change[a]						Soil Moisture Std. Dev. (mm) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	11.0	-72%	-57%	-91%	-84%	-92%	5.30	-13%	0%	-13%	21%	-13%
Michigan	5.4	-83%	-70%	-94%	-93%	-96%	5.50	4%	31%	-25%	51%	-27%
Huron	7.1	-77%	-66%	-94%	-94%	-96%	4.40	25%	11%	5%	43%	-2%
Georgian	11.6	-81%	-78%	-93%	-90%	-93%	9.60	-3%	-26%	30%	21%	29%
St. Clair	5.9	-80%	-80%	-95%	-98%	-97%	1.10	18%	9%	-18%	-9%	-18%
Erie	3.7	-78%	-73%	-97%	-92%	-97%	1.00	-10%	-20%	-30%	10%	-30%
Ontario	10.3	-83%	-93%	-97%	-99%	-97%	2.30	39%	30%	61%	35%	61%
	Groundwater Std. Dev. (mm) & Transferred Climate Relative Change[a]						Total Moisture Storage Std. Dev. (mm) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	11.20	-1%	16%	4%	31%	4%	22.3	-7%	15%	-6%	29%	-6%
Michigan	6.40	8%	12%	-31%	27%	-33%	13.8	-1%	11%	-36%	27%	-36%
Huron	8.30	80%	8%	73%	104%	63%	12.8	59%	4%	41%	75%	33%
Georgian	82.40	-42%	8%	-60%	-54%	-79%	87.4	-50%	0%	-66%	-51%	-84%
St. Clair	2.10	43%	14%	57%	67%	57%	7.6	-25%	-45%	-29%	-25%	-30%
Erie	1.30	69%	8%	77%	62%	69%	5.0	-16%	-36%	-24%	-16%	-26%
Ontario	1.40	64%	43%	93%	64%	93%	11.2	-17%	-35%	-12%	-32%	-12%
	Overland Evapotranspiration Std. Dev. (mm) & Transferred Climate Relative Change[a]						Runoff as Overland Depth Std. Dev. (mm) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	38.5	80%	96%	144%	208%	144%	38.6	-1%	24%	6%	66%	7%
Michigan	49.2	143%	83%	139%	180%	141%	44.0	35%	57%	-6%	96%	-5%
Huron	47.2	172%	64%	237%	165%	246%	49.2	64%	40%	51%	147%	46%
Georgian	45.6	153%	69%	185%	146%	188%	45.8	9%	-26%	50%	69%	50%
St. Clair	67.5	84%	34%	129%	83%	130%	72.5	54%	15%	69%	90%	68%
Erie	62.5	89%	32%	111%	83%	111%	63.2	81%	58%	81%	109%	81%
Ontario	41.3	114%	115%	183%	162%	183%	55.2	115%	60%	151%	67%	151%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

5.3 Over-water Meteorology

The over-water air temperature, humidity, and wind speed differs from over land since the lower atmospheric layer is affected by the water surface over which it lies. The model corrections to over-land meteorological observations for over-water conditions depend heavily on the water surface temperature, which in turn is a function of the over-water meteorology and heat balance at the surface of the lake. Table 19 summarizes annual average steady-state over-water meteorology differences.

In general, the synergistic relationship that exists between air and water temperature in the transposed climates yields a general increase in both that follow the transposed climate patterns, similar to over-land behavior. As with over-land air temperatures, Table 19 shows that over-water air temperatures increase most for the southern-most scenarios (3, 4, and 5) but, because of the buffering effect of the water surface in contact with the over-lake air, the differences are not as great. In general, over-lake air temperatures do

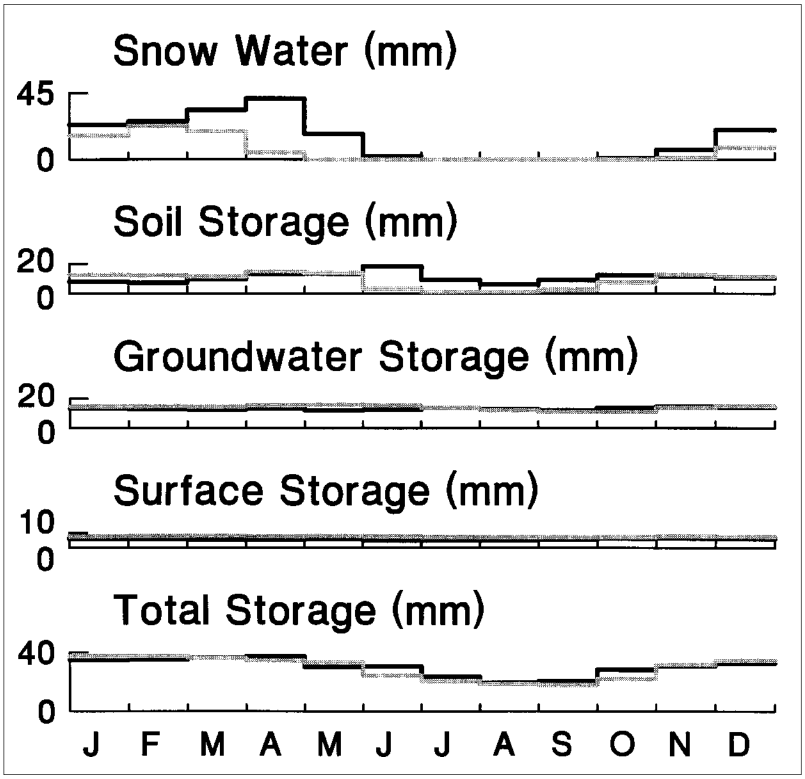


Figure 25.--Seasonal Lake Superior Basin Moisture Storage Standard Deviations for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

Figure 26.--Seasonal Lake Superior Average Overlake Meteorology for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

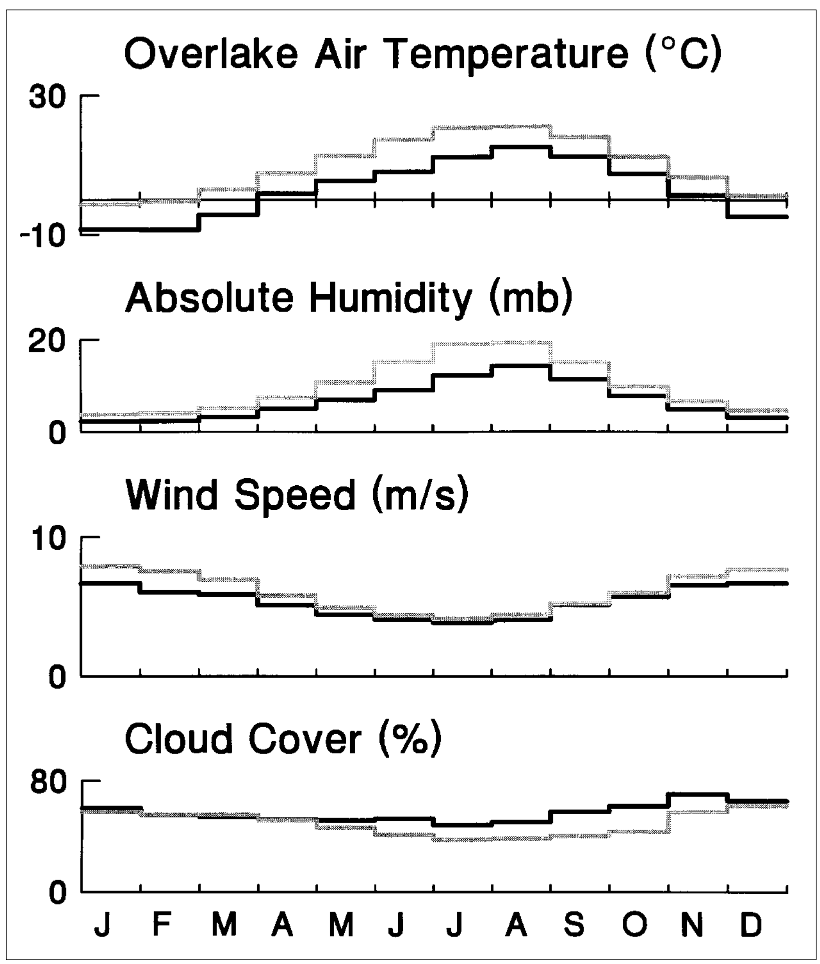


Table 19.--Average Annual Steady-State Overlake Meteorology Differences.

Basin	Air Temperature (ºC) & Transferred Climate Absolute Differences[a]						Humidity (mb) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	3.1	6.3	6.7	9.2	10.4	9.3	6.9	2.5	3.2	3.8	5.2	4.0
Michigan	7.8	4.8	5.0	8.3	9.4	8.4	9.7	1.8	2.2	3.9	5.5	3.9
Huron	6.6	5.6	5.8	10.1	9.9	11.3	8.7	2.6	2.6	5.8	5.6	7.0
Georgian	5.9	6.4	5.9	10.9	10.0	11.5	8.2	3.2	2.6	6.3	5.7	7.0
St. Clair	10.2	2.8	3.8	8.6	8.6	8.6	10.8	1.4	1.4	5.8	5.6	5.7
Erie	9.8	6.6	4.6	9.2	8.2	9.2	10.8	3.5	1.6	6.1	4.7	6.0
Ontario	7.8	5.8	5.3	9.9	9.2	9.9	9.2	3.2	2.6	6.3	5.5	6.3
	Cloud Cover (%) & Transferred Climate Relative Changes[a]						Wind Speed (m s^{-1}) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	56.7	-30%	-14%	-38%	-27%	-35%	5.36	15%	12%	23%	10%	27%
Michigan	42.8	-13%	11%	-19%	-6%	-12%	5.93	2%	-14%	-1%	-16%	5%
Huron	53.9	-23%	-1%	-26%	-20%	-21%	5.59	3%	-13%	-11%	-11%	-20%
Georgian	58.9	-28%	-7%	-32%	-26%	-29%	5.68	1%	-11%	-10%	-10%	-16%
St. Clair	49.8	-18%	-2%	-19%	-11%	-19%	5.33	3%	-25%	-8%	-21%	-8%
Erie	62.5	-36%	-22%	-34%	-30%	-34%	6.06	-13%	-26%	-14%	-16%	-14%
Ontario	59.0	-24%	-5%	-28%	-29%	-28%	5.74	-11%	-21%	-16%	-12%	-16%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

not rise as much as the over-land air temperatures and the distinction between southern and northern climates is not as pronounced. Over-water absolute humidity is increased for all scenarios; see Table 19. As with air temperatures, the southern-most climate scenarios show the greatest increase in over-lake humidity. Both over-lake air temperature and over-lake humidity show an increase and a slight shift earlier in the seasonal cycle, for all lakes and all scenarios; this is typified in Figure 26 for scenario 2 on Lake Superior. The shift reflects an interaction of the atmosphere with the lake's heat storage, which is discussed subsequently. This pattern is reflected in water surface temperatures as well, and is also discussed subsequently.

Cloud cover is reduced for all scenarios and all lakes, with the single exception of scenario 2 on Lake Michigan; see Table 19. The reduction is slightly more for the western-most scenarios (1, 3, and 5); there are generally fewer clouds to the extreme southwest of the Great Lakes. The seasonal variation of cloud cover is more difficult to ascertain in general. Figure 26 is typical of other scenarios and other lakes in that the variation of cloud cover is only approximately similar to the base case. Over-lake wind speed is increased on Lake Superior for all scenarios, and scenario 1 (6ºSx10ºW) gives slightly higher windspeeds on all lakes above Erie; see Table 19. However, for all other cases (scenarios 2, 3, 4, and 5 on all lakes below Superior), wind speeds are slightly decreased. In all cases, the differences appear slight. As typified in Figure 26 for Lake Superior under scenario 2, the seasonal variation of wind speed is very similar to the base case for all lakes and all scenarios.

Variability in over-lake meteorology is shown for annual values in Table 20 and depicted for seasonal values in the example of scenario 2 on Lake Superior in Figure 27. Over-lake air temperature variability is generally reduced on all lakes for all scenarios; Lake St. Clair is the only lake showing an increase for scenarios 2 and 4. Lake St. Clair, however, is a very shallow lake with almost no heat storage and hence

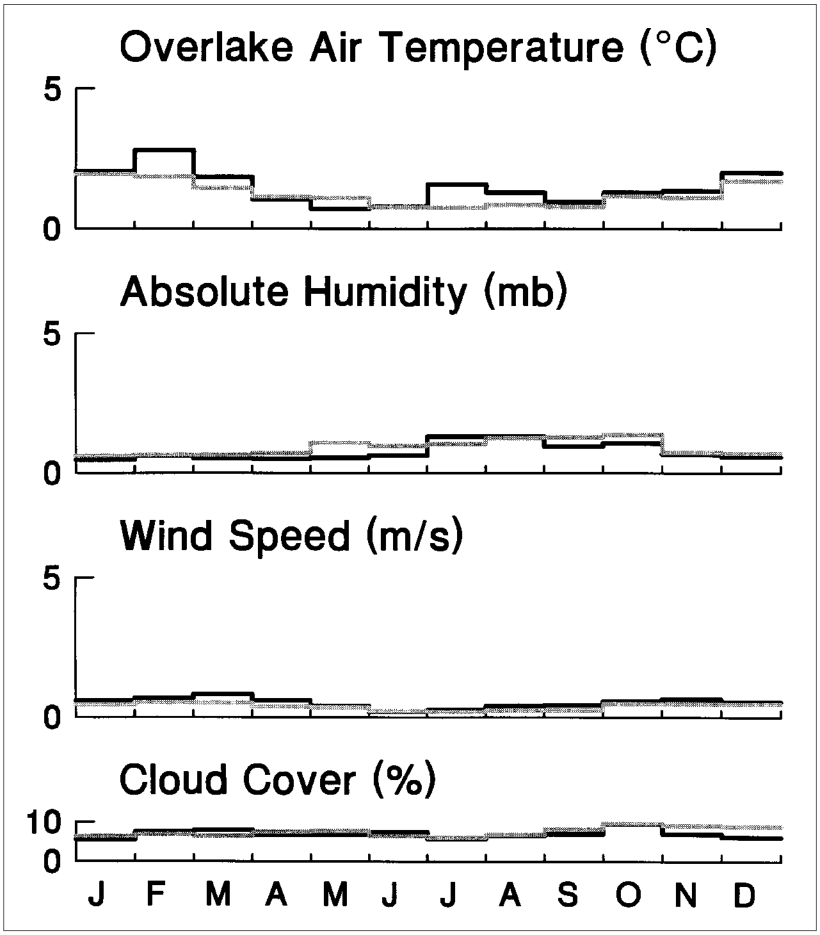


Figure 27.--Seasonal Lake Superior Average Overlake Meteorology Standard Deviations for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

very little effect on over-lake air temperatures. Seasonally, the variation of over-lake air temperatures appears to be smoother throughout the year than the base case with a minimum in the summer and a maximum in the winter, reflecting the steadying influence of lake heat storage on water surface temperature and therefore on over-lake air temperature. Since all transposed climates result in more heat in storage in the lakes, discussed subsequently, temperatures are higher and less variable. Figure 27 shows typical variation of air temperature. Figure 27 also depicts other typical change in variability for all lakes and all scenarios, in that wind speed is generally slightly less variable, and cloud cover is generally slightly more variable than the base case throughout the seasonal cycle. There also appears to be very little seasonality in the variability associated with either wind speed or cloud cover, neither in the base case nor in any of the transposed climate scenarios.

5.4 Lake Heat Balance

Lake heat balance changes are depicted in Table 21, and an example is given in Figure 28. Insolation changes in Table 21 largely reflect the cloud cover changes given earlier in Table 19; the western-most transposed climates (1, 3, and 5) transfer more heat into the lakes than do the eastern-most. The increase is spread throughout the annual cycle fairly uniformly for most lakes and all scenarios; Figure 28 is typical. Reflection changes are very small, relative to the insolation changes, with most of the difference coming in the winter-spring due to the absence of ice cover. The large changes in reflection on Lake St.

Table 20. Average Annual Steady-State Overlake Meteorology Variability Differences.

Basin	Air Temperature Std. Dev. (ºC) & Transferred Climate Relative Change[a]						Humidity Std. Dev. (mb) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	0.90	-33%	-33%	-44%	-44%	-44%	0.50	-20%	-20%	0%	0%	0%
Michigan	0.80	-25%	-13%	-50%	-38%	-50%	0.40	25%	25%	50%	50%	25%
Huron	0.70	-29%	-14%	-29%	-29%	-29%	0.40	25%	25%	50%	25%	50%
Georgian	0.70	0%	0%	-29%	-29%	-29%	0.30	67%	67%	100%	67%	133%
St. Clair	0.60	0%	33%	-17%	33%	-17%	0.40	50%	50%	50%	75%	50%
Erie	0.80	-38%	-13%	-38%	-38%	-38%	0.50	20%	20%	20%	0%	0%
Ontario	0.90	-33%	-33%	-44%	-33%	-44%	0.60	0%	-17%	-17%	0%	-17%
	Cloud Cover Std. Dev. (%) & Transferred Climate Relative Change[a]						Wind Speed Std. Dev. (m s^{-1}) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	2.80	29%	21%	39%	32%	39%	0.40	-50%	-50%	-25%	-50%	-25%
Michigan	2.00	90%	95%	95%	65%	95%	0.20	0%	-50%	0%	0%	0%
Huron	3.20	22%	9%	12%	-3%	12%	0.20	0%	0%	-50%	-50%	-50%
Georgian	2.30	74%	39%	52%	35%	52%	0.30	-67%	-33%	-33%	-33%	-33%
St. Clair	3.40	12%	-12%	9%	3%	9%	0.40	25%	-25%	-25%	-50%	-25%
Erie	5.20	-37%	-42%	-38%	-40%	-38%	0.20	0%	50%	0%	0%	0%
Ontario	3.60	6%	28%	-14%	-22%	-14%	0.20	0%	0%	-50%	0%	-50%

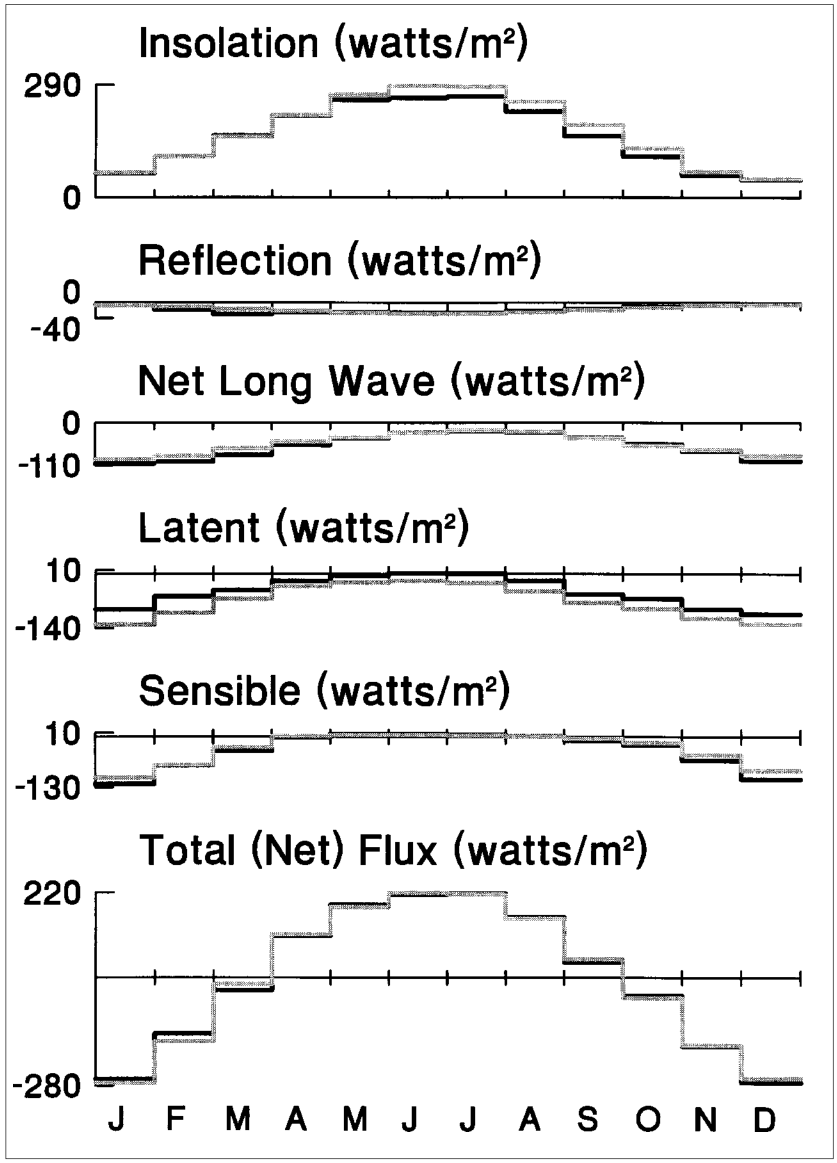
[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Table 21.--Average Annual Steady-State Lake Heat Flux Differences.

Basin	Insolation (w m^{-2}) & Transferred Climate Absolute Differences[a]						Reflection (w m^{-2}) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	158	25	13	31	24	29	-18	0	1	-1	0	-1
Michigan	172	8	-5	13	5	10	-19	1	2	0	1	0
Huron	165	17	-2	19	15	17	-18	0	2	0	0	0
Georgian	159	21	3	25	20	24	-23	5	7	5	5	5
St. Clair	166	13	1	14	7	14	-30	3	8	12	13	12
Erie	148	33	19	31	26	31	-20	2	3	2	2	2
Ontario	148	21	2	22	22	22	-15	-2	0	-2	-2	-2
	Net Long Wave Exchange (w m^{-2}) & Transferred Climate Absolute Differences[a]						Latent Heat Flux (w m^{-2}) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	-61	-3	6	0	6	2	-45	-27	-25	-42	-40	-42
Michigan	-81	3	5	8	11	10	-50	-20	-10	-32	-26	-33
Huron	-73	2	6	8	10	9	-48	-23	-13	-35	-33	-36
Georgian	-64	2	6	7	8	9	-50	-28	-18	-41	-37	-43
St. Clair	-59	0	2	12	12	12	-69	-16	-8	-35	-28	-35
Erie	-41	-7	-8	2	1	2	-70	-30	-14	-38	-32	-38
Ontario	-60	-2	2	6	4	6	-51	-21	-11	-34	-33	-34
	Sensible Heat Flux (w m^{-2}) & Transferred Climate Absolute Differences[a]						Net Heat Flux (w m^{-2}) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	-34	6	6	13	12	13	0	0	0	0	1	0
Michigan	-21	8	9	13	12	15	0	0	0	0	0	0
Huron	-25	4	7	10	10	12	0	0	0	0	0	0
Georgian	-21	0	3	6	5	7	0	0	0	0	0	0
St. Clair	-5	1	-2	0	-2	0	0	0	0	0	0	0
Erie	-16	4	1	5	5	5	0	0	0	0	0	0
Ontario	-21	4	6	9	10	9	0	0	0	0	0	0

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Figure 28.--Seasonal Lake Superior Average Lake Heart Fluxes for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.



Clair are atypical because the changes in ice cover, and hence reflection, are most dramatic on Lake St. Clair. Net long wave exchange increases slightly, implying more heat stays in the lakes. Sensible heat exchange also increases even more than the net long-wave exchange, again implying more heat stays in the lakes. Table 21 shows that the long-term average increases in both are more pronounced on all lakes for the southern-most scenarios (3, 4, and 5). The overall increases in heat storage in the lakes thus far discussed are balanced by increases in evaporation, shown in Table 21 as a decrease in latent heat transferred into the lake. These evaporation increases are rather large compared to the base case. The seasonal patterns of net long wave exchange, sensible heat transfer, and latent heat transfer are very similar to the base case; see Figure 28. The changes shown in these variables, summarized in Table 21, are distributed fairly uniformly throughout the seasonal cycle; again Figure 28 is fairly typical of the pattern of changes observed on all lakes in all scenarios. The annual total heat flux should remain close to zero for all scenarios, as in the base case, indicating that there is no long-term heat storage in the lakes, and energy conservation is satisfied. Table 21 shows this, and Figure 28 indicates that the change in the seasonal

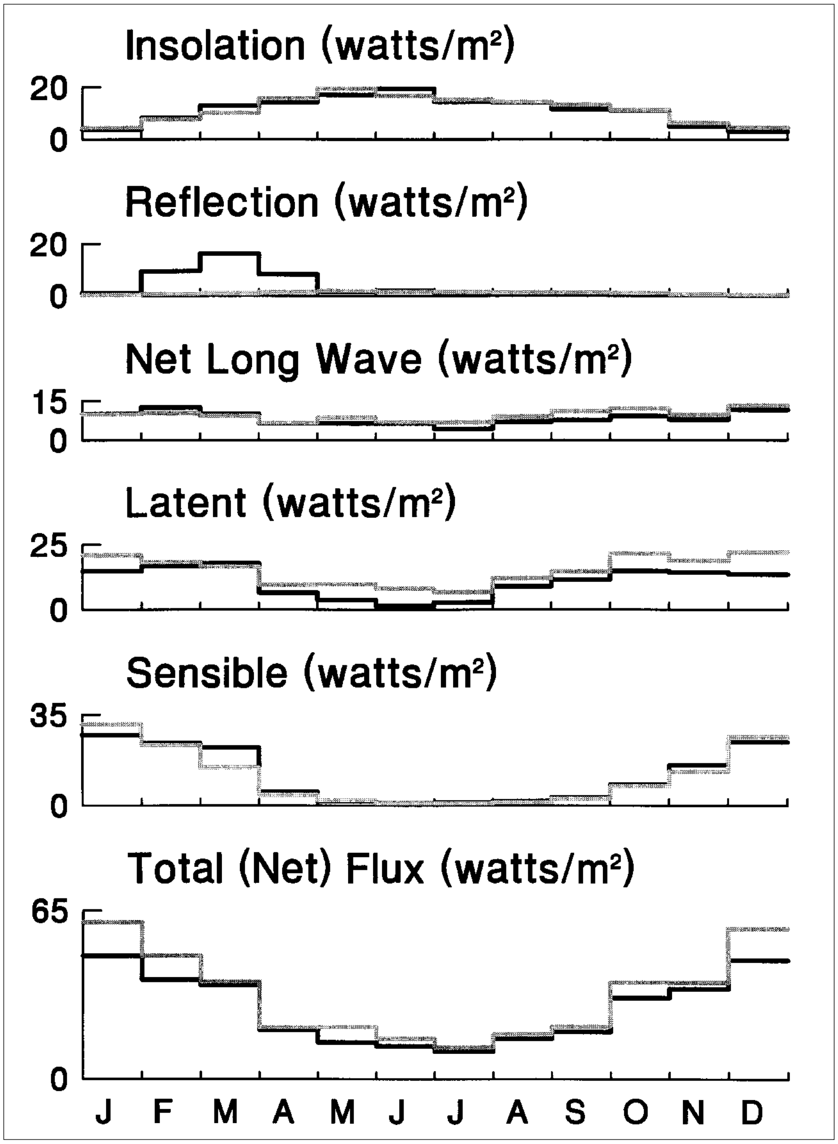


Figure 29.--Seasonal Lake Superior Lake Heat Fluxes Standard Deviations for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

variation of total heat flux is very similar to the base case; increased heat fluxes are balanced largely by evaporation.

Changes in variability in lake heat balance variables are summarized in Table 22 and typified in Figure 29 for scenario 2 on Lake Superior. As seen in Table 22, insolation and net long wave exchange are more variable than in the base case, although the relative change is less than 100%. The increased variability in insolation reflects the increased variability in cloud cover shown in Table 20. In particular, trends in cloud cover are reflected directly in solar insolation; the reduced variability under all scenarios of Lake Erie insolation and net long wave radiation in Table 22 matches the reduced variability of cloud cover for Erie in Table 20. The reduced variability of Lake Ontario insolation in Table 22 matches the reduced variability of cloud cover in Table 20. There does not seem to be pronounced patterns in the shifted seasonal cycles for the variability of either insolation or net long wave exchange; see Figure 29.

The variability of reflection in Table 22 is greatly reduced, reflecting the greatly reduced ice cover (discussed subsequently) that exists under the transposed climates. Without ice cover, reflection is from

Table 22.--Average Annual Steady-State Lake Heat Flux Variability Differences

Basin	Insolation Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]						Reflection Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	5.00	10%	8%	14%	10%	14%	2.60	-81%	-81%	-77%	-77%	-77%
Michigan	3.40	59%	82%	76%	62%	76%	1.80	-72%	-67%	-67%	-67%	-67%
Huron	4.80	23%	27%	27%	10%	27%	2.90	-79%	-79%	-79%	-83%	-79%
Georgian	4.00	55%	42%	52%	33%	52%	3.30	-82%	-79%	-82%	-85%	-82%
St. Clair	5.60	2%	-2%	13%	4%	13%	1.70	18%	76%	-29%	-35%	-29%
Erie	8.10	-31%	-32%	-32%	-35%	-32%	2.90	-79%	-79%	-79%	-83%	-79%
Ontario	5.50	9%	42%	-9%	-18%	-9%	0.80	-25%	0%	-38%	-38%	-38%

	Net Long Wave Exchange Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]						Latent Heat Flux Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	3.30	9%	12%	21%	30%	21	4.60	15%	13%	22%	2%	24%
Michigan	1.70	24%	29%	29%	29%	24%	5.20	4%	2%	10%	-6%	12%
Huron	2.20	36%	32%	32%	32%	36%	5.00	10%	14%	6%	8%	0%
Georgian	2.40	58%	33%	50%	46%	54%	4.80	19%	15%	12%	8%	12%
St. Clair	2.90	31%	28%	10%	31%	10%	4.00	10%	25%	-2%	0%	-2%
Erie	4.20	-2%	5%	-7%	0%	-7%	6.00	-42%	-23%	-32%	-33%	-32%
Ontario	2.80	29%	39%	21%	32%	21%	4.70	13%	15%	-4%	9%	-4%

	Sensible Heat Flux Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]						Net Heat Flux Std. Dev. ($w\ m^{-2}$) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	4.80	2%	-13%	-10%	-23%	-10%	8.10	19%	6%	15%	4%	16%
Michigan	3.20	12%	-9%	-16%	-31%	-19%	6.90	9%	-6%	1%	-6%	1%
Huron	3.80	-3%	-29%	-37%	-34%	-58%	7.60	4%	-1%	-4%	7%	-18%
Georgian	3.40	9%	-24%	-26%	-26%	-32%	6.60	20%	14%	14%	18%	11%
St. Clair	0.90	67%	67%	33%	67%	33%	2.30	13%	-13%	-17%	-22%	-17%
Erie	2.60	-12%	-23%	-19%	-31%	-19%	5.30	0%	-4%	8%	-2%	8%
Ontario	3.10	-3%	-23%	-26%	-39%	-26%	7.30	7%	-8%	-4%	1%	-4%

[a]Scenario #1 is 6ºSxx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

the water surface only, and the fraction of incoming radiation reflected remains nearly constant throughout the seasonal cycle. Figure 29 shows that the change in reflection variability is concentrated in the winter-spring period where ice cover would affect reflection.

Sensible heat transfer is seen to be less variable almost everywhere, for all scenarios, with the exception of Lake St. Clair. Again, the absence of any real heat storage on Lake St. Clair precludes the filtering effect possible with such storages on meteorology and heat transfers. The seasonal patterns of fluctuation of sensible flux variabilities are very similar to the base case; see Figure 29. The variability changes are mostly concentrated in the cool part of the season. Latent heat transfer is seen in Table 22 to be only slightly more variable than the base case for all scenarios on all lakes except for Lakes Erie (all scenarios) and Ontario (scenarios 3 and 5). These exceptions probably correspond to the cloud-cover variability changes, and the corresponding variability changes in insolation and net long wave exchange, already noted. Seasonally, the variability increase appears to be spread throughout the annual cycle in Figure 29, resulting in less seasonality than with the base case variability. The net effect of variability changes in all of the heat balance components for a lake is a slight shift in variability in the total flux. For Lake Superior and scenario 2, Figure 29 shows this variability increases slightly in the cool months.

The heat budget gives rise to increased water surface temperatures as seen in Figure 30 and summarized in Table 23. Stored heat increases between 23% and 180% on the average over the Great Lakes, depending on the transposed climate considered; see Table 23. The largest relative heat increases are seen to occur for the southern-most climates (scenarios 3, 4, and 5), but are substantial in all cases. The stored heat appears as a constant amount higher throughout the seasonal cycle, since we are looking at steady-state conditions; see Figure 30. The increased heat in storage also means that ice formation will be greatly reduced over winter on the deep Great Lakes. Ice cover is practically eliminated under all transposed climate scenarios on all lakes but Lake St. Clair; since that lake has very little heat storage capacity, ice formation is not affected as much as elsewhere. The average steady-state increase in water surface temperatures for all transposed climate scenarios on all lakes range from 2.2ºC on Lake St. Clair (scenario 1) to 11.5ºC on Lake Huron (scenario 5). The heat storage capacity of a lake influences the increase in water surface temperatures that can almost be seen in Figure 30. Water surface temperatures are seen to peak earlier on deep lakes under the transposed climates than under the base case. Again, the southern-most transposed climate scenarios (3, 4, and 5) result in the greatest water surface temperature increases. The increased heat in storage is sufficient to cause increased lake evaporation on all lakes under all scenarios, even though wind speeds and humidity, by themselves, would not increase evaporation. (Wind speed and humidity changes, in some cases, would decrease lake evaporation, all other things being equal.) Table 23 shows increases in annual lake evaporation of 12%–96%, depending on the lake and the scenario. Again, note that the southern-most transposed climate scenarios result in the largest increases in lake evaporation over the base case.

The variabilities associated with the lake heat balance variables are summarized in Table 24 and depicted for Lake Superior, scenario 2, in Figure 31. The stored heat exhibits some increase in variability for all scenarios, in Table 24, for the deep lakes only. The variability appears to be spread more uniformly across the seasonal cycle in every transposed climate scenario than in the base case, largely as a result of the disappearance of the ice pack. The dip in total heat storage variability during the winter-spring period, associated with the presence of the ice pack, is eliminated. Figure 31 is typical in this regard. Also, since the ice pack is not present anymore (see Figure 30), the variability associated with the ice pack is zero (ice pack stays at a constant zero value); see Figure 31. This means a relative change of 100% in the standard deviation of ice cover in Table 24.

The increased heat storage in the lakes (see Table 23) results in a greater thermal inertia for each lake and the water surface temperature is less variable; see Table 24. Figure 31 shows that for Lake Superior under scenario 2, the variability associated with water surface temperature is spread more uniformly throughout the season than was true for the base case. This is true for other scenarios and other lakes and reflects, again, the absence of the ice cover under the transposed scenarios. Not seen well in Figure 31, but more pronounced on other lakes and other scenarios, is a shift from a peak variability of water surface temperature from the summer to the spring. This shift is the result of a change in the fundamental behavior of heat storage in the lake.

5.5 Lake Thermal Structure

The deep lakes (Superior, Michigan, Huron, Georgian Bay, and Ontario) show water surface temperatures that stay above 3.98ºC throughout the annual cycle in some years. Figure 32 illustrates this for 1961 for Lake Superior under both the base case climate and under the scenario 2 transposed climate. This means that buoyancy-driven turnovers of the water column do not occur in the same way as they do at

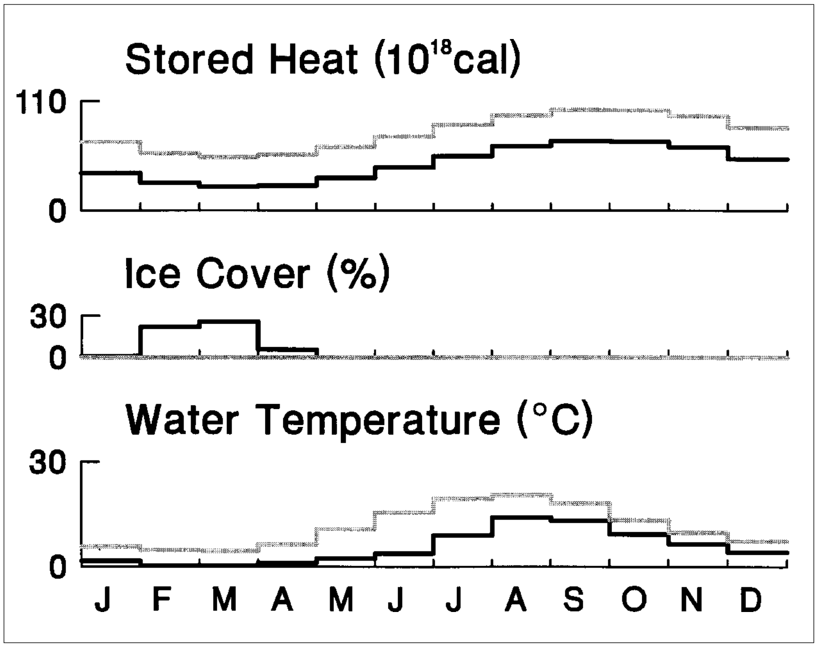


Figure 30.--Seasonal Lake Superior Average Lake Heat Storage Characteristics for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

Table 23.--Average Annual Steady-State Lake Heat Balance Differences

Basin	Stored Heat (10^{17} cal) & Transferred Climate Relative Changes[a]						Ice Cover (%) & Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	47.1	53%	66%	112%	146%	118%	4.6	-100%	-100%	-100%	-100%	-100%
Michigan	23.9	34%	36%	98%	126%	111%	2.6	-99%	-100%	-100%	-100%	-100%
Huron	15.9	58%	65%	138%	137%	180%	3.4	-100%	-100%	-100%	-100%	-100%
Georgian	4.8	66%	62%	138%	125%	149%	15.7	-99%	-99%	-100%	-100%	-100%
St. Clair	0.0	23%	37%	64%	66%	64%	35.0	-25%	-49%	-96%	-97%	-96%
Erie	4.9	54%	41%	82%	71%	82%	14.3	-100%	-99%	-100%	-100%	-100%
Ontario	8.4	68%	62%	157%	144%	157%	0.9	-100%	-100%	-100%	-100%	-100%

	Water Surface Temperature (°C) & Transferred Climate Absolute Differences						Lake Evaporation Depth (mm) & [a]Transferred Climate Relative Changes[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	5.5	5.1	5.9	7.5	9.2	7.6	569	61%	57%	94%	91%	96%
Michigan	8.7	4.1	4.6	7.4	9.0	7.3	640	39%	20%	65%	54%	67%
Huron	8.0	5.0	5.4	9.7	9.5	11.5	612	48%	28%	74%	70%	76%
Georgian	7.7	5.5	5.2	10.2	9.3	10.9	634	57%	36%	85%	75%	88%
St. Clair	10.9	2.2	3.6	8.0	8.4	8.0	888	24%	12%	52%	42%	52%
Erie	11.0	6.2	4.8	8.9	7.8	8.9	895	44%	20%	56%	47%	56%
Ontario	8.6	6.1	5.6	10.2	9.2	10.2	645	42%	21%	68%	66%	68%

[a]Scenario #1 is 6°Sx10°W; #2 is 6°Sx0°W; #3 is 10°Sx11°W; #4 is 10°Sx5°W; #5 is #3 with lake effects.

Table 24.--Average Annual Steady-State Lake Heat Balance Variability Differences.

	BasinStored Heat Std. Dev. (10^{17} cal) & Transferred Climate Relative Change[a]						Ice Cover Std. Dev. (%) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	3.10	23%	61%	55%	87%	58%	5.50	-100%	-100%	-100%	-100%	-100%
Michigan	1.50	7%	67%	47%	80%	40%	3.40	-97%	-100%	-100%	-100%	-100%
Huron	1.10	18%	73%	27%	64%	18%	5.90	-100%	-100%	-100%	-100%	-100%
Georgian	0.30	33%	67%	67%	67%	67%	6.80	-96%	-88%	-100%	-100%	-100%
St. Clair	0.00	0%	0%	0%	0%	0%	2.90	34%	155%	0%	-7%	0%
Erie	0.30	-33%	0%	0%	0%	0%	8.00	-100%	-93%	-100%	-100%	-100%
Ontario	0.80	50%	50%	25%	38%	25%	1.80	-100%	-100%	-100%	-100%	-100%
	Water Surface Temperature Std. Dev. (ºC) & Transferred Climate Relative Change[a]						Lake Evaporation Depth Std. Dev. (mm) & Transferred Climate Relative Change[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	0.70	-14%	-29%	-43%	-29%	-43%	59.4	14%	12%	20%	3%	24%
Michigan	0.80	-25%	0%	-50%	-38%	-50%	67.3	4%	1%	8%	-6%	11%
Huron	0.60	-17%	0%	-17%	-17%	-33%	64.1	11%	15%	7%	9%	1%
Georgian	0.50	20%	40%	20%	0%	20%	62.0	19%	13%	12%	8%	12%
St. Clair	0.60	0%	17%	0%	17%	0%	51.1	12%	27%	-1%	2%	-1%
Erie	0.70	-29%	0%	-29%	-29%	-29%	78.5	-43%	-25%	-33%	-35%	-33%
Ontario	1.00	-40%	-40%	-60%	-50%	-60%	60.2	13%	15%	-3%	9%	-3%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Table 25.--Average Characteristics of Turnovers/Reversals.

Basin	Fraction Dimictic[a]						Interarrival Times[a]					
	Base	#1	#2	#3	#4	#5	Base	#1	#2	#3	#4	#5
Superior	100 %	78 %	24 %	0 %	0 %	0 %	182 d	205 d	293 d	365 d	365 d	365 d
Michigan	100 %	75 %	50 %	0 %	0 %	0 %	182 d	208 d	241 d	364 d	364 d	365 d
Huron	100 %	18 %	2 %	0 %	0 %	0 %	182 d	308 d	356 d	365 d	365 d	365 d
Georgian	100 %	87 %	66%	0 %	0 %	0 %	182 d	195 d	221 d	365 d	365 d	364 d
St. Clair	100 %	100 %	100 %	93 %	79 %	93 %	182 d	182 d	182 d	187 d	202 d	187 d
Erie	100 %	56 %	63 %	0 %	7 %	0 %	182 d	234 d	221 d	365 d	342 d	365 d
Ontario	100 %	0 %	0 %	0 %	0 %	0 %	182 d	364 d	365 d	365 d	365 d	365 d

	Monomictic Reversal Water Temperature[a]					
	Base	Scenario #1	Scenario #2	Scenario #3	Scenario #4	Scenario #5
Superior	-	4.3ºC	4.5ºC	6.1ºC	7.4ºC	6.3ºC
Michigan	-	4.2ºC	4.7ºC	6.9ºC	8.6ºC	7.6ºC
Huron	-	4.8ºC	5.7ºC	9.5ºC	9.6ºC	12.4ºC
Georgian	-	4.3ºC	4.7ºC	8.0ºC	7.3ºC	8.9ºC
St. Clair	-	-	-	4.6ºC	4.8ºC	4.6ºC
Erie	-	5.5ºC	5.5ºC	7.9ºC	7.0ºC	7.9ºC
Ontario	-	5.8ºC	5.9ºC	10.7ºC	9.9ºC	10.7ºC

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

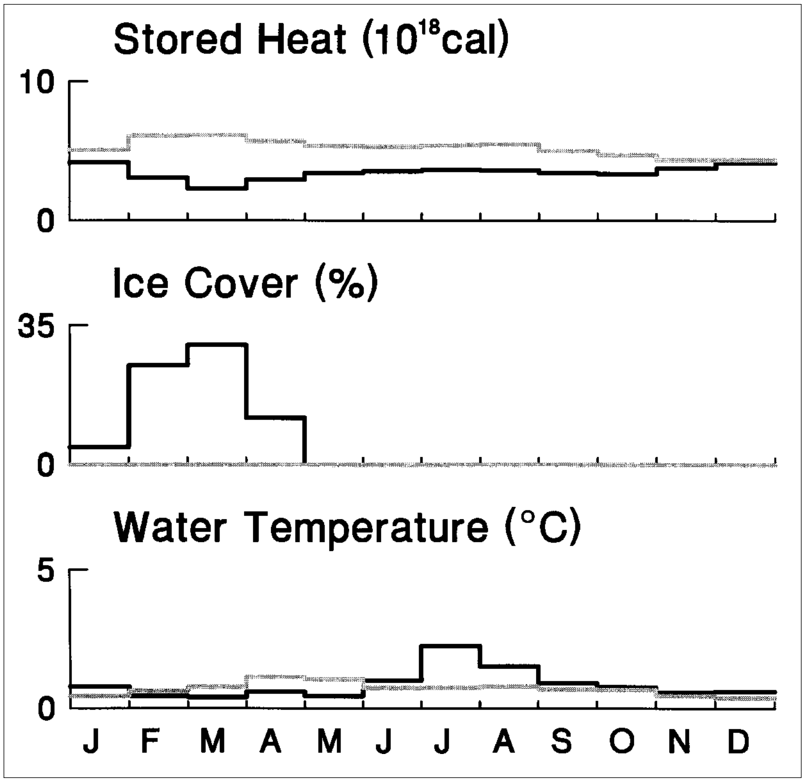


Figure 31.--Seasonal Lake Superior Lake Heat Storage Characteristics Standard Deviation for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

present. In some years, the large lakes are changed from dimictic lakes (turnovers occur twice a year as water temperatures pass through the point of maximum density, 3.98ºC) to monomictic lakes (maximum turnover occurs at the temperature "reversal" where temperatures stop declining and start rising again, and the minimum temperature is greater than 3.98ºC). Figure 32 shows that the base case temperature profile for Lake Superior passed through 3.98ºC in June 1961 and approached, in December, the January 1962 transition. Under the scenario 2 transposed climate, temperatures remain above 3.98ºC but approach a vertical profile in March. This represents a change from dimictic to monomictic.

Table 25 shows that the large lakes remain dimictic under the transposed climates only between 0% and 87% of the time, depending on the lake and the scenario. The largest change is associated with Lake Ontario, which is the furthest south of the deep lakes. Least affected are Lakes Erie and St. Clair, which are very shallow and have relatively little heat storage. The southern-most transposed climates (scenarios 3, 4, and 5) show the largest shifts, with all deep lakes becoming 100% monomictic. As the lakes change to one reversal per year in some years, instead of two turnovers per year, the interarrival times of the maximum mixing extent increase. Table 25 illustrates that the average interarrival time grows to a full year for the deep lakes under the southern-most scenarios. Table 25 also illustrates the monomictic reversal temperature is, of course, well above the point of maximum water density. Again, the southern-most transposed climates show the highest monomictic reversal water temperatures. Since Lake St. Clair remains 100% dimictic under both scenarios 1 and 2, there are no entries for that lake for monomictic reversal water temperature in Table 25.

The timing of maximum turnovers or temperature reversal shifts. Table 26 shows the time increases between the spring turnover and the fall turnover (for dimictic behavior). The spring turnover occurs earlier and the fall turnover occurs later in the annual cycle. For monomictic behavior, the single maxi-

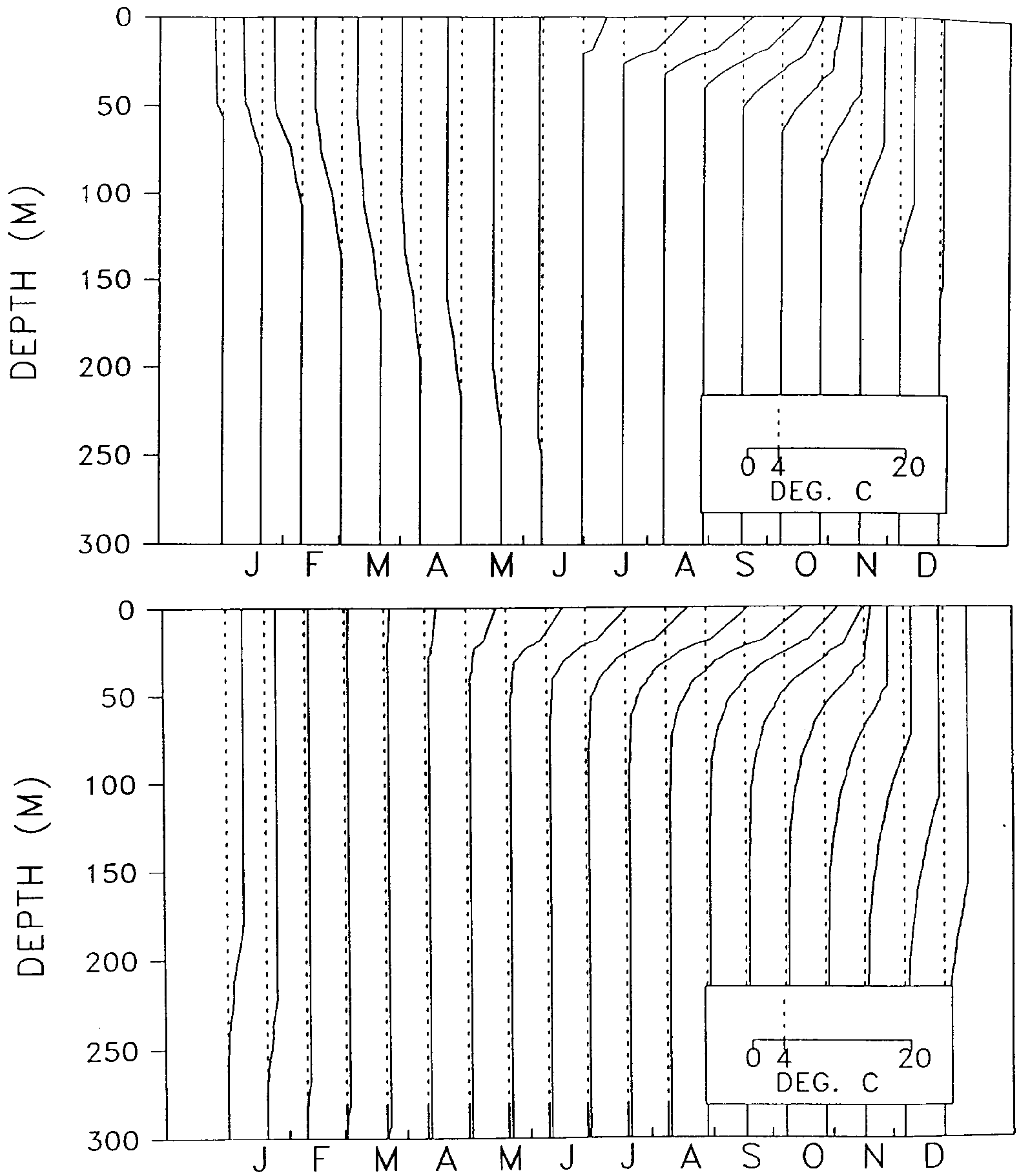


Figure 32.--Steady-State Lake Superior Temperature-Depth Profiles for (top) the Base Case Climate and (bottom) the Scenario 2 Transposed Climate (6ºS×0ºW).

Table 26.--Average Dates and depths of Maximum Turnover or Temperature Reversal

Basin	Dimictic Dates[a]											
	Base		Scenario #1		Scenario #2		Scenario #3		Scenario #4		Scenario #5	
	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall
Superior	27 Jun	24 Dec	16 Apr	12 Feb	10 Apr	22 Feb	-	-	-	-	-	-
Michigan	23 May	31 Dec	31 Mar	5 Feb	6 Apr	6 Feb	-	-	-	-	-	-
Huron	21 May	13 Jan	25 Mar	24 Feb	19 Mar	3 Feb	-	-	-	-	-	-
Georgian	26 May	29 Dec	2 Apr	2 Feb	3 Apr	6 Feb	-	-	-	-	-	-
St. Clair	04 May	19 Nov	8 Apr	22 Nov	25 Mar	1 Dec	7 Feb	27 Dec	8 Feb	26 Dec	6 Feb	27 Dec
Erie	25 Apr	24 Dec	25 Feb	31 Jan	10 Mar	29 Jan	-	-	28 Feb	12 Dec	-	-
Ontario	21 May	14 Jan	-	-	-	-	-	-	-	-	-	-

	Dimictic Depths[a]											
	Base		Scenario #1		Scenario #2		Scenario #3		Scenario #4		Scenario #5	
	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall
Superior	238 m	156 m	134 m	264 m	99 m	275 m	-	-	-	-	-	-
Michigan	123 m	105 m	55 m	175 m	47 m	145 m	-	-	-	-	-	-
Huron	229 m[b]	229 m[b]	68 m	207 m	0 m	229 m[b]	-	-	-	-	-	-
Georgian	164 m[b]	157 m	132 m	147 m	125 m	136 m	-	-	-	-	-	-
St. Clair	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]	6 m[b]
Erie	63 m	64 m[b]	19 m	59 m	29 m	60 m	-	-	13 m	45 m	-	-
Ontario	231 m	202 m	-	-	-	-	-	-	-	-	-	-

	Monomictic Dates[a]						Monomictic Depths[a]					
	Base	#1	#2	#3	#4	#5	Base	#1	#2	#3	#4	#5
Superior	-	14 Mar	19 Mar	14 Mar	14 Mar	13 Mar	-	298 m	306 m	304 m	291 m	291 m
Michigan	-	8 Mar	4 Mar	1 Mar	28 Feb	5 Mar	-	225 m	195 m	226 m	205 m	234 m
Huron	-	11 Mar	13 Mar	8 Mar	7 Mar	12 Mar	-	224 m	224 m	216 m	216 m	220 m
Georgian	-	25 Feb	8 Mar	26 Feb	28 Feb	24 Feb	-	142 m	147m	164 m[b]	158 m	161 m
St. Clair	-	-	-	5 Jan	5 Jan	5 Jan	-	-	-	6 m[b]	6 m[b]	6 m[b]
Erie	-	9 Feb	17 Feb	9 Feb	13 Feb	9 Feb	-	61 m	64 m[b]	60 m	60 m	60 m
Ontario	-	10 Mar	11 Mar	5 Mar	5 Mar	5 Mar	-	244 m[b]	232 m	228 m	234 m	233 m

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.
[b]Maximum average depth of the lake.

mum turnover occurs even earlier in the year than the dimictic turnovers. These are consequences of greater heat storage in, and heat inputs to, the lakes.

Temperature-depth profiles, as in Figure 32, for every day of a single model year can be combined and depicted as depth-time plots of temperature isolines; see Figure 33 for an example on Lake Superior for the base case and scenario 2. Then, not only are the turnover timing changes depicted between the base case and the transposed climate, but depth changes are more apparent as well. Table 26 also summarizes the maximum depths at turnover in the lakes. Dimictic spring turnovers exhibit shallower average depths under the transposed climate scenarios than under the base case conditions, and fall turnovers are deeper on Lakes Superior and Michigan. Monomictic turnovers are generally even deeper on Lakes Superior, Michigan, and Ontario. On Lakes Huron, Georgian Bay, and Erie, both dimictic and monomictic turnover depths are reduced under the transposed climates as compared to the base case. Dimictic turnovers do not occur at all on Lake Ontario under any of the transposed climates. Dimictic turnovers do occur on Lake St. Clair, with relatively no heat storage, for all transposed climates. Lake

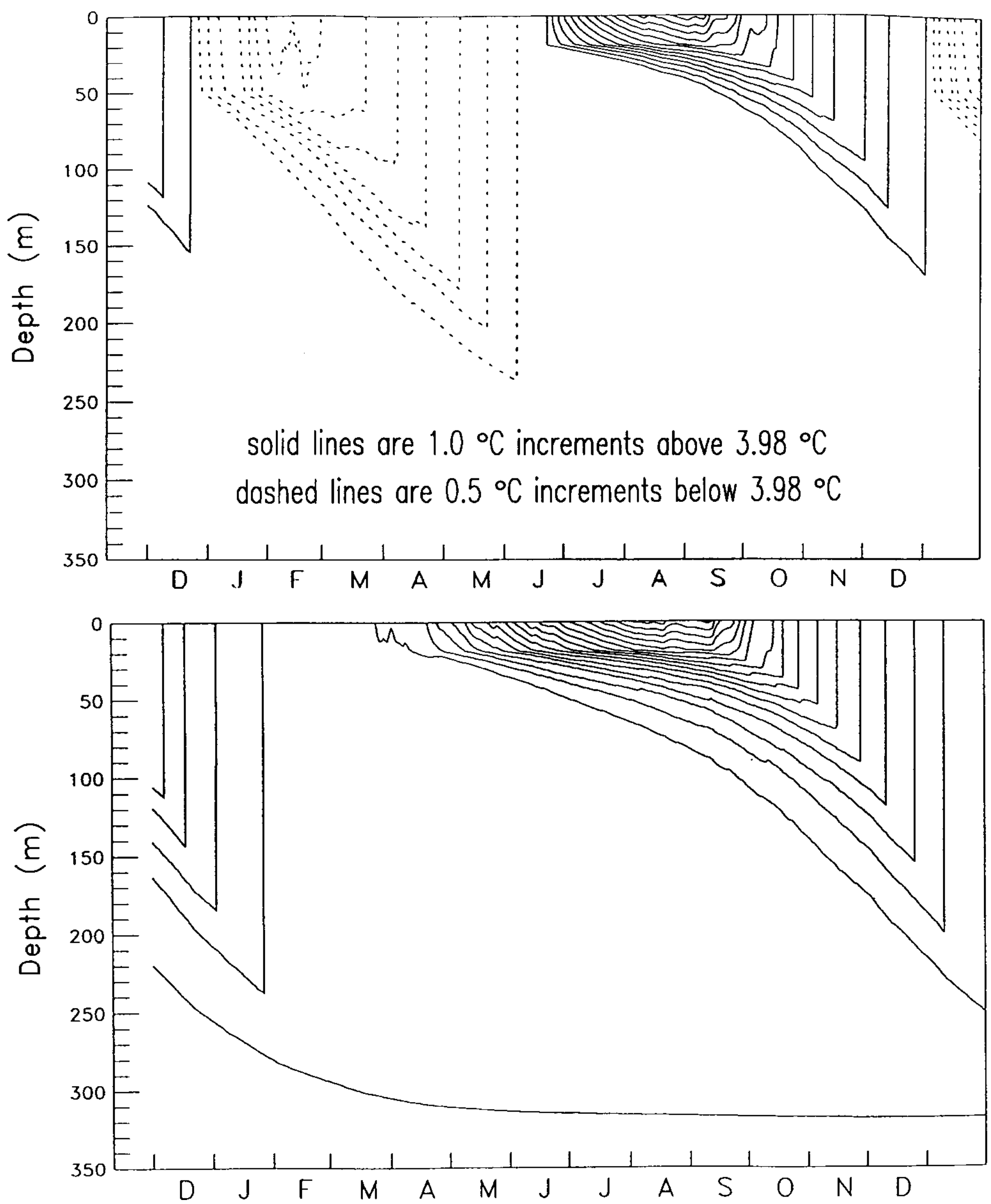


Figure 33.--Steady-State Lake Superior Depth-Time Temperature Isolines for (top) the Base Case Climate and (bottom) the Scenario 2 Transposed Climate (6ºS×0ºW).

Erie also has more dimictic turnovers than the deep lakes under all scenarios, as it shows dimictic turnovers occurring under scenario 4.

There is a normal hysteresis observed in graphs of lake heat plotted with surface temperature, such as in Figure 34 for Lake Superior under the base case. This reflects the mixing of heat at depth. Surface temperatures rise quickly, and heat storage follows after the spring turnover. When surface temperatures then begin to drop in the fall, stored heat does not initially. Then heat storage drops more slowly. Similar behavior occurs after the fall turnover, and both result in the characteristic double “loop” in the plot. Under the warmer climate change scenario, temperatures sometimes never drop below that at maximum

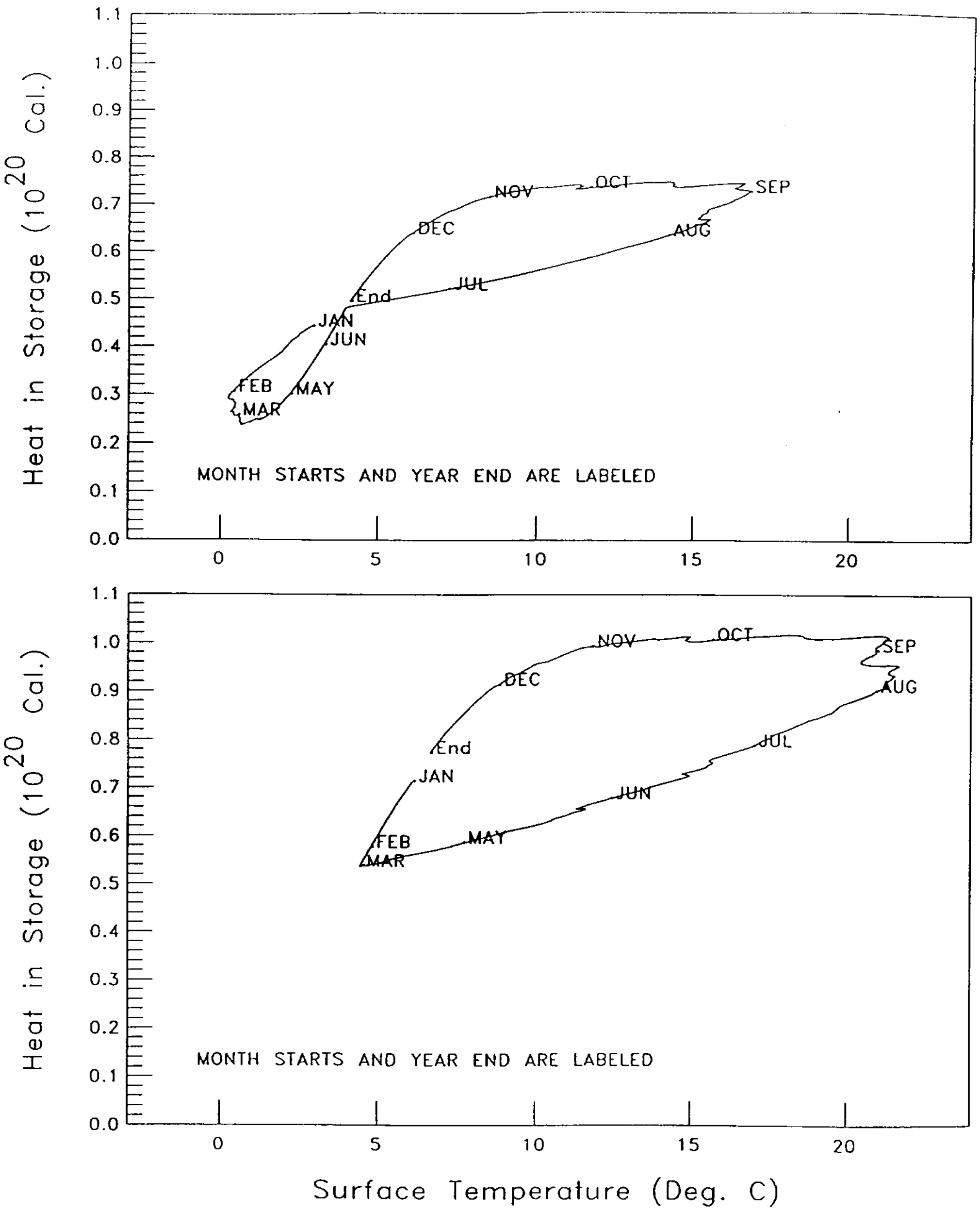


Figure 34.--Steady-State Lake Superior Heat-Temperature Hysteresis for (top) the Base Case Climate and (bottom) the Scenario 2 Transposed Climate (6ºS×0ºW).

density (3.98ºC). This results in only one hysteresis loop, but it is much larger. Figure 34 shows this typical behavior for the transposed climate of scenario 2.

Quinn and den Hartog (1981) noted in their studies on Lake Ontario during the International Field Year on the Great Lakes that evaporation events can take place in which a majority of annual evaporation occurs within a few days. By estimating the evaporation time series on each lake under each scenario, it is possible to address the question, “to what extent is evaporation an event-oriented process?” Inspection of the time series of meteorology and estimated evaporation on all of the Great Lakes illustrates the nature of the process. For example, under the base case, significant evaporation on Lake Superior begins in

August and continues through April (evaporation is generally July through March on the other Great Lakes except Lake Erie where ice cover restricts significant evaporation to July through December or January). During this period, lake evaporation appears to be highly variable, depending to a large extent on wind speed. Individual daily estimated evaporation peaks often do correspond to daily wind speed peaks, but the extent of this correspondence changes during the evaporation season. While wind speed fluctuates daily about an underlying constant during the period September through March on Lake Superior, estimated evaporation fluctuates daily like wind speed but generally increases from September to December and then generally decreases. As wind speed increases and humidity drops during the fall and early winter, estimated evaporation increases. Then, dropping water temperatures lower the vapor deficit over water, and estimated evaporation generally drops during late winter and spring. Superimposed on this general behavior are the fluctuations corresponding to changes in wind speed and humidity that are associated with the passage of air masses. Thus, evaporation may be separated into two components: a steady component rising and falling throughout the fall and winter, and a highly variable component corresponding to passage of individual air cells (events).

Inspection of the simulations for each of the Great Lakes, for each of the climates considered here, reveals that a third of the estimated annual evaporation occurs in about one tenth of the annual cycle (not necessarily continuously); see Table 27. One half of the annual evaporation total occurs in about one fifth of the annual cycle, and two thirds occur in about one quarter to one third of the annual cycle. While the bulk of the annual evaporation still occurs in a small part of the year, Table 27 shows that the total annual evaporation is spread a little more throughout the annual cycle for each of the transposed climates, as compared to the base case. While all annual evaporation under the base case occurs in 79.8%–88.9% of the annual cycle, depending on the lake, it occurs under the transposed climates in 92.3%–98.2% of the annual cycle, depending on the lake and the scenario.

5.6 Lake Water Balance

Over-lake precipitation, runoff, and lake evaporation sum algebraically as the net basin supply to the lake and are presented again in Table 28 for convenience. The observations, presented previously, on over-land precipitation equally apply for over-lake precipitation. Both are estimated in the same way as over basin (both land and lake) precipitation, which is assumed to be the same over both land and lake; recall lake effects are ignored. Since over-lake precipitation is taken here as the same as over-land, Table 28 shows the same relations for the transposed climates vs. the base case precipitation as does Table 15. Likewise, basin runoff is the same as in Table 17, except that it is expressed as an over-lake depth rather than as an over-land depth. Lake evaporation is simply repeated from Table 23 for convenience. Also, precipitation, runoff, and evaporation are presented as absolute differences from the base case, rather than as relative changes. Net basin supply in Table 28 is generally less than the base case for all transposed climate scenarios for the deep lakes (Superior, Michigan, Huron, Georgian Bay, and Ontario). The exceptions among the deep lakes are Michigan and Huron under scenarios 2 and 4. Also, the shallow lakes (St. Clair and Erie) show generally higher net basin supplies for all the transposed climate scenarios with the exception being a slight decrease on Erie under scenario 3.

In addition to the absence of trends in annual net basin supplies relative to the base case, that apply for all of the transposed climate scenarios, there are no seasonal characteristics of the transposed climates that apply across all lakes and all scenarios. Therefore, the seasonal plot in Figure 35 for Lake Superior under scenario 2 is not typical of the other lakes. For Lake Superior, net basin supplies are lower throughout the seasonal cycle under all scenarios than under the base case, except for March under scenarios 2 and 4. They also peak at the same time of the year (May) under all scenarios as in the base case.

Table 27.--Minimum Fractions of the Year for Occurrence of Annual Lake Evaporation Fractions.

Annual Evap-oration	Lake Superior[a]						Lake Michigan[a]					
Fract. %	Base %	S #1 %	S #2 %	S #3 %	S #4 %	S #5 %	Base %	S#1 %	S#2 %	S#3 %	S#4 %	S #5 %
5	1.1	1.0	1.2	1.2	1.3	1.2	1.0	1.1	1.0	1.1	1.1	1.1
10	2.4	2.4	2.7	2.7	2.9	2.6	2.2	2.4	2.3	2.5	2.6	2.4
20	5.4	5.7	6.2	6.2	6.6	6.1	5.2	5.8	5.5	5.9	6.1	5.9
33.3	10.2	11.0	11.8	11.9	12.4	11.8	9.9	11.2	10.6	11.8	12.0	11.8
50	17.6	19.3	20.1	20.6	21.3	20.5	17.2	19.8	18.6	21.2	21.3	21.4
66.7	26.7	29.8	30.5	31.8	32.3	31.7	26.5	31.2	29.1	33.8	33.6	34.3
80	36.0	41.1	41.7	44.5	44.4	44.4	36.4	43.7	40.5	47.3	46.5	48.1
90	45.7	54.0	54.0	59.5	58.0	59.5	47.2	57.2	53.2	61.1	59.7	62.2
95	53.2	64.8	63.9	71.3	69.0	71.1	55.6	67.4	62.9	71.2	69.1	72.4
100	79.8	95.8	94.4	98.2	97.4	97.5	82.9	94.4	92.3	96.0	94.5	96.6

Annual Evap-oration	Lake Huron[a]						Georgian Bay					
Fract. %	Base %	S #1 %	S #2 %	S #3 %	S #4 %	S #5 %	Base %	S#1 %	S#2 %	S#3 %	S#4 %	S #5 %
5	1.0	1.1	1.1	1.2	1.1	1.2	1.0	1.1	1.1	1.2	1.1	1.1
10	2.3	2.4	2.5	2.7	2.5	2.6	2.2	2.4	2.5	2.7	2.5	2.5
20	5.2	5.6	5.8	6.3	5.8	6.2	5.0	5.5	5.7	6.3	6.0	5.8
33.3	9.9	10.7	11.0	12.0	11.3	12.2	9.5	10.4	11.0	12.0	11.6	11.3
50	17.0	18.4	18.9	20.8	19.9	22.2	16.4	18.2	19.0	21.0	20.3	20.2
66.7	26.0	28.2	29.0	32.1	31.2	35.7	25.0	28.2	29.5	32.9	32.0	32.3
80	35.6	39.0	40.0	44.6	44.0	49.9	34.7	39.3	40.9	45.8	44.8	45.5
90	46.1	51.5	52.6	58.6	57.9	63.9	46.3	52.1	53.7	59.4	58.4	59.2
95	54.6	61.9	62.9	68.8	68.3	73.5	56.1	62.5	63.7	69.3	68.2	68.8
100	85.1	92.7	95.6	97.0	96.5	97.7	88.2	92.8	95.0	96.2	95.6	96.8

Annual Evap-oration	Lake Erie[a]						Lake Ontario[a]					
Fract. %	Base %	S #1 %	S #2 %	S #3 %	S #4 %	S #5 %	Base %	S#1 %	S#2 %	S#3 %	S#4 %	S #5 %
5	0.9	1.1	1.1	1.2	1.2	1.2	1.0	1.2	1.1	1.2	1.2	1.2
10	2.1	2.6	2.6	2.7	2.7	2.7	2.2	2.6	2.6	2.8	2.6	2.8
20	4.9	6.1	6.2	6.6	6.4	6.6	5.1	6.0	6.0	6.5	6.2	6.5
33.3	9.3	12.0	12.2	13.1	12.6	13.1	9.8	11.6	11.5	12.8	11.9	12.8
50	16.4	21.3	21.7	23.2	22.5	23.2	16.9	20.3	20.2	22.6	21.2	22.6
66.7	25.9	33.2	33.7	35.7	34.9	35.7	26.0	31.4	31.4	35.4	33.5	35.4
80	36.7	45.5	46.0	48.0	47.2	48.0	35.7	43.5	43.5	48.8	46.9	48.8
90	48.8	57.8	58.3	60.2	59.3	60.2	46.3	56.5	56.4	62.1	60.8	62.1
95	58.5	66.8	67.2	69.0	68.1	69.0	54.8	66.3	66.3	71.4	70.8	71.4
100	88.9	94.2	96.1	94.2	94.5	94.2	84.2	94.8	96.9	96.8	96.6	96.8

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Table 28.--Average Annual Steady-State Lake Water Balance Differences.

	BasinOverlake Precipitation (mm) & Transferred Climate Absolute Differences[a]						Runoff as Overwater Depth (mm) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	817	-188	50	-163	171	-161	615	-353	-164	-379	-191	-377
Michigan	828	22	322	6	486	36	645	-183	151	-216	185	-190
Huron	813	209	325	393	570	413	390	-37	82	18	153	19
Georgian	908	21	95	269	426	283	1803	-676	-360	-360	-6	-325
St. Clair	854	242	282	439	525	455	4454	560	874	1271	1548	1338
Erie	913	286	399	342	501	353	810	210	392	140	293	151
Ontario	934	242	166	461	312	461	1701	-19	-238	151	-374	151

	Lake Evaporation Depth (mm) & Transferred Climate Absolute Differences[a]						Net Basin Supply (mm) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	569	347	327	536	515	545	863	-887	-441	-1078	-535	-1083
Michigan	640	252	127	415	347	430	833	-413	347	-625	325	-584
Huron	612	295	169	454	426	468	590	-123	238	-43	297	-37
Georgian	634	359	229	538	475	560	2076	-1014	-495	-629	-55	-601
St. Clair	888	216	106	465	376	463	4420	586	1049	1245	1697	1329
Erie	895	391	180	498	421	498	828	105	612	-17	373	6
Ontario	645	272	138	438	423	438	1990	-49	-210	174	-485	174

[a]Scenario #1 is 6°Sx10°W; #2 is 6°Sx0°W; #3 is 10°Sx11°W; #4 is 10°Sx5°W; #5 is #3 with lake effects.

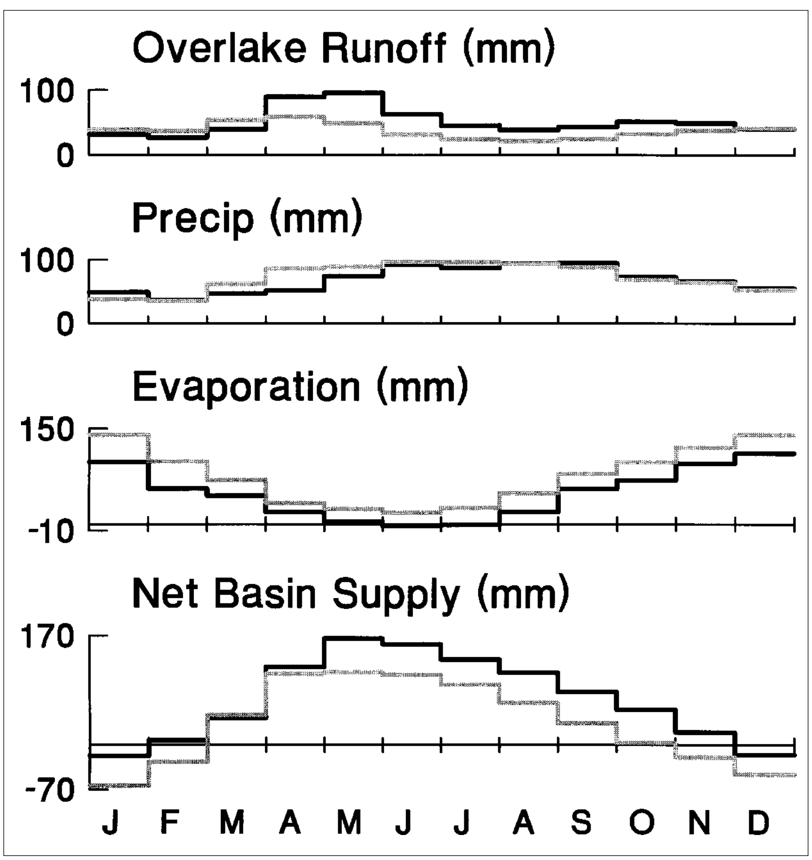


Figure 35.--Seasonal Lake Superior Average Net Basin Supply Components for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

On Lake Michigan, they were lower throughout the seasonal cycle under the western-most scenarios (1, 3, and 5) than under the base case, except for February under scenario 5. Lake Michigan under the eastern-most scenarios (2 and 4) showed a marked increase in seasonal net basin supplies over the base case from November to May and a slight decrease from June to October. The time of the peak net basin supply on Lake Michigan shifted from April to May under the western-most scenarios (1, 3, and 5), and to March for the eastern-most scenarios (2 and 4).

On Lake Huron, seasonal net basin supplies became slightly higher during the winter-spring part of the seasonal cycle with a shift in the peak from April to May under the western-most scenarios, and they became more than slightly higher during the winter-spring part of the seasonal cycle with a shift in the peak from April to March under the eastern-most scenarios.

On Georgian Bay, seasonal net basin supplies were lower throughout the year under scenario 1 and only slightly higher during February and March and lower throughout the rest of the year under scenario 2; both scenarios saw no shift in the peak net basin supply (March) under these northern-most scenarios. For the southern-most scenarios (3, 4, and 5), Georgian Bay net basin supplies were higher during December through March but much lower during the rest of the year, with a shift in the peak net basin supply from April to March.

On Lake Erie under all transposed climate scenarios, net basin supplies were higher from October or November to May or June; there were shifts in the peak from April to March under scenarios 1, 2, and 4, and to February under scenarios 3 and 5.

On Lake Ontario, there is a shift in the peak net basin supplies under all transposed climate scenarios from April to March. Also the net basin supplies are greater during December through March under the western-most scenarios (1, 3, and 5) than under the base case; for the eastern-most scenarios (2 and 4), the net basin supplies are greater than the base case during January through March (a minor exception is August of scenario 2 where it is only slightly higher).

The variabilities associated with the net basin supplies and its components are depicted seasonally in Figure 36 for Lake Superior under scenario 2, and annual values are summarized in Table 29 for all lakes and scenarios. With the single exception of Georgian Bay under scenario 2, the annual variability of net basin supplies increases on all lakes under all scenarios; see Table 29. Seasonally, the variability is distributed across the seasonal cycle approximately as it is in the base case, but it is larger. There does appear to be generally greater increases in variability over the late fall-winter-early spring part of the seasonal cycle, relative to the late spring-summer-early fall part. Figure 36 shows these two observations only partially for Lake Superior under scenario 2, but they are fairly general across all lakes and scenarios.

Table 30 summarizes the changes in the hydrological and net basin supply components for the entire Great Lakes basin; they were computed by converting the equivalent depths of Table 28 to annual flow rates on each lake and adding them over all the lakes. The changes from the base case are also expressed relatively in Table 30. Also expressed relatively are changes from other studies that used other GCMs (Croley, 1990, 1993a); they are provided for comparison. Net basin supplies to the Great Lakes taken as a whole are seen to drop to about one half under the first and third scenarios (the western-most scenarios). This drop in net basin supply seems to result from the increases in over-lake evaporation and over-land evapotranspiration (reducing subsequent runoff to the lakes). While evaporation and evapotranspiration have increased just as significantly under the transposed climates of scenarios 2 and 4 as well, the precipitation increases (both over-land and over-lake) for scenarios 2 and 4 compensate, and the

Table 29.--Average Annual Steady-State Lake Water Balance Variability Differences.

	BasinOverlake Precipitation Std. Dev. (mm) & Transferred Climate Absolute Differences[a]						Runoff as Overwater Depth Std. Dev. (mm) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	84	27%	41%	52%	110%	52%	60	-1%	24%	6%	65%	7%
Michigan	94	84%	64%	64%	124%	64%	88	35%	57%	-6%	96%	-5%
Huron	89	127%	52%	153%	154%	153%	61	64%	40%	51%	147%	45%
Georgian	94	76%	17%	114%	101%	114%	198	9%	-26%	49%	69%	49%
St. Clair	122	80%	22%	106%	90%	106%	25	2227%	639%	2871%	3740%	2841%
Erie	110	99%	51%	105%	103%	105%	149	81%	58%	81%	108%	81%
Ontario	90	99%	71%	149%	96%	149%	204	115%	60%	151%	67%	151%

	Lake Evaporation Depth Std. Dev. (mm) & Transferred Climate Absolute Differences[a]						Net Basin Supply Std. Dev. (mm) & Transferred Climate Absolute Differences[a]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	59.4	14%	12%	20%	3%	24%	164	21%	32%	39%	80%	41%
Michigan	67.3	4%	1%	8%	-6%	11%	203	62%	53%	38%	100%	38%
Huron	64.1	11%	15%	7%	9%	1%	165	103%	47%	108%	133%	104%
Georgian	62.0	19%	13%	12%	8%	12%	303	30%	-17%	65%	70%	64%
St. Clair	51.1	12%	27%	-1%	2%	-1%	142	456%	112%	591%	24%	586%
Erie	78.5	-43%	-25%	-33%	-35%	-33%	268	87%	51%	86%	99%	86%
Ontario	60.2	13%	15%	-3%	9%	-3%	304	102%	57%	141%	74%	141%

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.

Table 30.--Average Annual Steady-State Great Lakes Basin Hydrology Summary.

Scenario	Overland Precipitation (m^3 s^{-1})		Evapo transpiration (m^3 s^{1})		Basin Runoff (m^3 s^{1})		Overlake Precipation (m^3 s^{1})		Overlake Evaporation (m^3 s^{1})		Net Basin Supply (m^3 s^{1})	
Base	13855		7814		6206		6554		4958		7803	
6ºS x 10ºW	14643	+6%	10201	+31%	4674	-25%	6767	+3%	7394	+49%	4048	-48%
6ºS x 0ºW	17167	+24%	11198	+43%	6154	-1%	8169	+25%	6615	+33%	7708	-1%
10ºS x 11ºW	16236	+17%	11563	+48%	4877	-21%	7379	+13%	8699	+75%	3556	-54%
10ºS x 5ºW	20095	+45%	13907	+78%	6308	+2%	9482	+45%	8364	+69%	7426	-5%
CCC[a]		-2 %		22 %		-32 %		0 %		32 %		-46 %
GISS[b]		2 %		21 %		-24 %		4 %		27 %		-37 %
GFDL[c]		1 %		19 %		-23 %		0 %		44 %		-51 %
OSU[d]		6 %		19 %		-11 %		6 %		26 %		-23 %

[a]Canadian Climate Centre (Croley, 1993a).
[b]Goddard Institute for Space Studies GCM (Croley, 1990).
[c]Geophysical Fluid Dynamics Laboratory GCM (Croley, 1990).
[d]Oregon State University GCM (Croley, 1990).

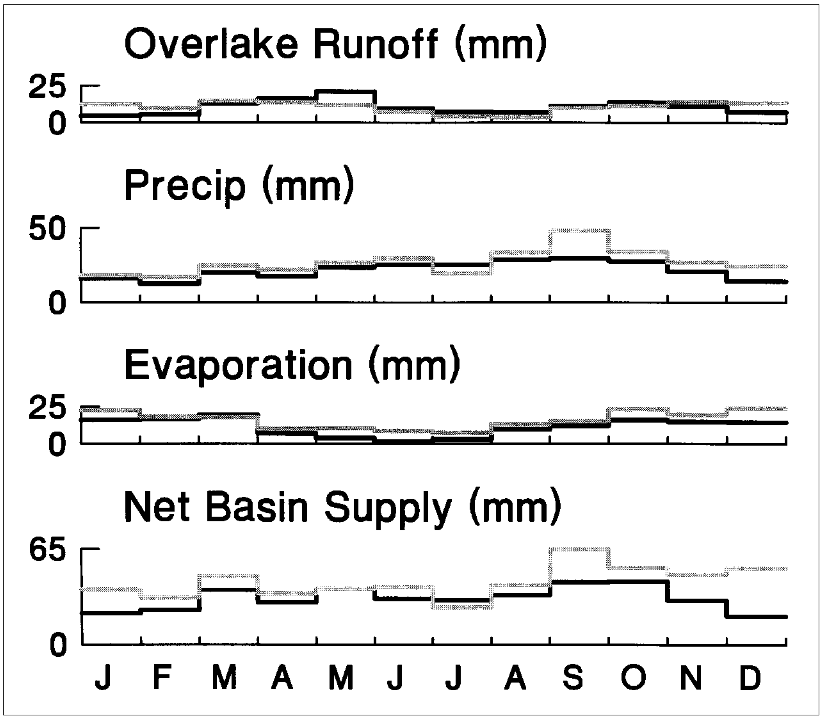


Figure 36.--Seasonal Lake Superior Net Basin Supply Components Standard Deviations for Scenario 2 (6°S×0°W). Black - Base; Gray - Scenario 2.

net basin supplies for scenarios 2 and 4 are very close to the base case. The results of scenarios 1 and 3 then, are similar to the earlier studies with GCMs, also reported in Table 30, in that the net supplies drop by about half due to increased basin evapotranspiration and lake evaporation; precipitation increases are not present however.

5.7 Hydrological Sensitivities

Without temperatures below freezing, the snowpack is insensitive to precipitation. Although the steady-state scenarios on different lakes show different estimates of precipitation change, each shows increases in air temperatures that significantly reduce the snowpack. Thus, even if precipitation increases more than suggested by these transposed climates, the snowpack will be much reduced under warmer climates. Similarly, although affected by actual changes in precipitation, the Great Lakes basin experiences reduced soil moisture storage with only minor exceptions among the lake basins and transposed climate scenarios. The combined effect on basin runoff to the lakes is mixed; the western-most scenarios (1, 3, and 5) show the greatest reduction in basin runoff, while the eastern-most (scenarios 2 and 4) show little overall change in runoff over the entire Great Lakes basin but more lake by lake variation. Both soil moisture and runoff peak shortly after snow melt and then drop throughout the summer and fall due to high evapotranspiration; each transposed climate scenario produces earlier snow melt and a longer period of evapotranspiration. Soil moisture and runoff are most sensitive to precipitation in midsummer when at annual minimums. Overall, runoff appears sensitive to both air temperature (as affects evapotranspiration and snow pack development) and precipitation. This is in contrast to earlier GCM studies (Croley, 1990, 1993a) where soil moisture and runoff appeared insensitive to precipitation, largely as a result of the limited range of precipitation present in the GCM scenarios.

Lake evaporation increased substantially on all lakes under all transposed climate scenarios. Interestingly, this occurs in the face of increased humidities on all lakes under all scenarios and decreased wind speeds on most lakes and scenarios, which would ordinarily reduce the evaporation. However, there is such a large increase in the heat that is input and stored in the lakes (through reductions in cloud cover and loss of ice cover) that water surface temperatures are so much higher under the transposed climates than under the base case. This increases the vapor pressure deficit between the water and overlying atmosphere to the point that the net effect is increased evaporation.

Precipitation changes associated with the transposed climate scenarios were larger and more variable than those considered in earlier GCM-based studies of climate change impacts in the Great Lakes basin. The results reflect this; changes in net basin supply are more variable across the transposed climates than was observable formerly in the GCM-based studies, and precipitation is seen to have a more pronounced effect on the behavior of the lakes under changed climates than before. Depending on precipitation changes, net basin supplies to the lakes vary from almost no annual difference to about half. However, even when there is little change in the supplies on an annual basis, there are considerable changes in the components of net supply, which compensate so the total changes little.

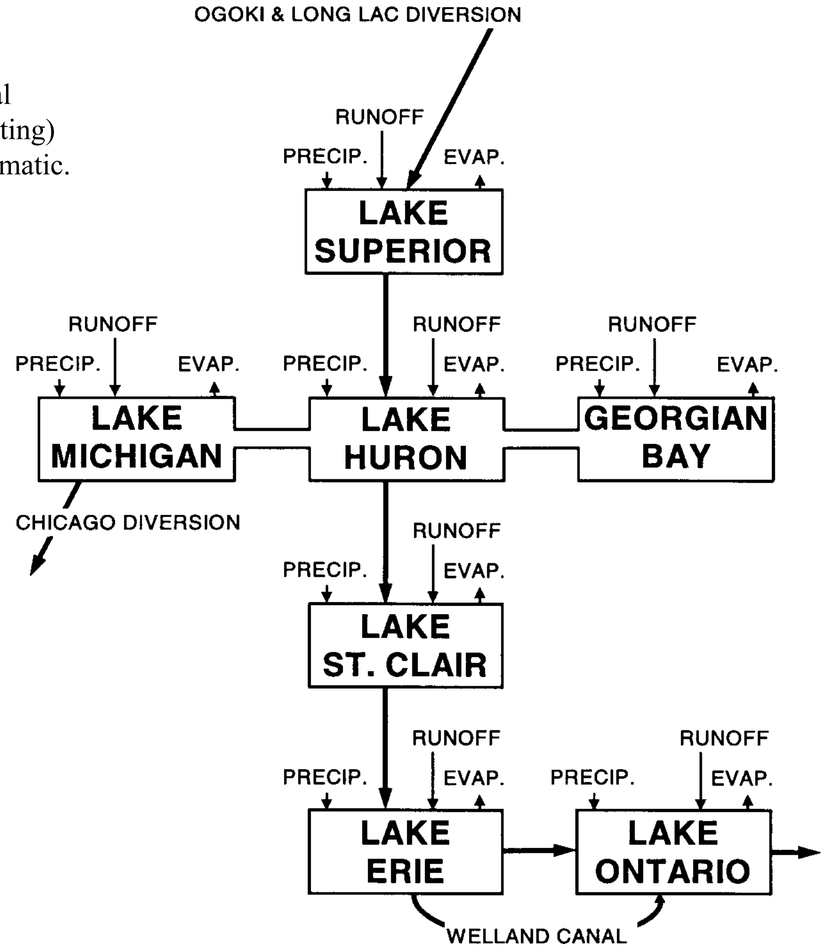
These results should be received with caution as they are, of course, dependent on arbitrary transposed climates and there are uncertainties associated with their transposition to the Great Lakes. There are also model uncertainties that affect the results. However, the linkage method used with GCMs in earlier studies is completely avoided. Those studies changed historical meteorology to match average changes in GCM outputs observed from simulations of $1xCO_2$ and $2xCO_2$ atmospheres. Those studies simply changed the magnitude of meteorological time series without affecting their temporal or spatial structures. Therefore, changes in variabilites that would take place under a change climate were not addressed, and seasonal timing differences in the GCMs for the changed climate were not reproduced. Now, however, all meteorological variabilities, and their temporal and spatial structures, are fully considered since alternate climates are transposed to the Great Lakes. This study with transposed climates allowed consideration of changed seasonal meteorological patterns and observation of seasonal changes induced by storage effects both in the values of the various variables and in their variabilities. Changes in annual variability also are now clear whereas in the GCM studies they were not.

The higher air temperatures under the transposed climate scenarios lead to higher over-land evapotranspiration and lower runoff to the lakes with earlier runoff peaks since snowpack is reduced up to 100% and the snow season is eliminated. This also results in a reduction in available soil moisture. Water temperatures increase and peak earlier; heat resident in the deep lakes increases throughout the year. Mixing of the water column diminishes, as most of the lakes become mostly monomictic, and lake evaporation increases. Without biannual turnovers, hypolimnion chemistry may be altered; oxygen may be depleted, releasing nutrients and metals from lake sediments. The lakes may experience more than a single winter turnover if temperature gradients are small and winds are strong enough to induce mixing (Hutchinson, 1957). Ice formation is greatly reduced over winter on the deep Great Lakes, and lake evaporation increases; average steady-state net supplies drop where precipitation increases are modest.

6. GREAT LAKES CLIMATE CHANGE LAKE LEVEL RESPONSES

Great Lakes levels and flows have been simulated for a variety of studies, including changed climates (Quinn, 1988; International Joint Commission, 1976; Hartmann, 1990; Lee et al., 1994). The basic procedure is to determine lake levels and connecting channel flows by routing the simulated water supplies through the Great Lakes system with a hydrological response model (Hartmann, 1987; Quinn, 1978). In addition to net basin supplies, monthly diversions and consumptive uses data (International Great Lakes Diversions and Consumptive Uses Study Board, 1981) are also input to the model. GLERL's Hydrological Response Model (see Figure 37) consists of regulation plans, channel routing dynamics, and water balances, combined to estimate lake levels and connecting channel flows from water supplies to the

Figure 37.--Hydrological Response (Channel Routing) Model Conceptual Schematic.



lakes. Lake Superior is regulated by Plan 1977–A (International Lake Superior Board of Control, 1981, 1982) and Lake Ontario by Plan 1958–D (International St. Lawrence Board of Control, 1963). The regulation plans were modified (Lee et al., 1994) and now have extreme condition operation rules. The modifications provided the robustness for the plans to handle the wider range of outflows expected during climate change and stochastic hydrological studies of the Great Lakes basin than were used in the derivation of the Plans. In addition several minor modifications were made to allow the models to function under the extreme high and low lake levels and flows expected under severe transposed climates. Middle lake outflows are represented with stage-fall-discharge equations as functions of lake levels or of lake level differences between lakes. Flow retardation from ice and weeds are given by monthly median retardation values. Constant diversions are used for the Ogoki, Long Lac, and Chicago diversions, and monthly means are used for Welland Canal diversions. Each lake storage, with all inflows and outflows, is described by mass continuity equations. The system of equations is solved numerically.

A base case using the existing available data was run for the period 1953–1990 because of a lack of historical Georgian Bay evaporation data prior to 1953. Following the base case, runs were made for each of the five transposed scenarios. All model runs were initialized by running to steady state conditions using Basis of Comparison conditions (Lee, 1993). Impacts were assessed by comparing the lake levels and flow outputs from the base case with those of the various scenarios. Comparisons of interest include impacts on the seasonal lake level cycles, mean lake level changes, and changes in connecting channel flow characteristics. The analysis follows on a lake by lake basis.

6.1 Lake Superior

The annual levels for the base and transposed climate runs for Lake Superior are depicted in Figure 38 and summarized in Table 31. Scenario 3 is not included since an equilibrium state was not reached. As shown in Table 28, the hydrology for scenario 3 resulted in negative net basin supplies with Lake Superior becoming a terminal lake (no outflows) at an undetermined elevation. Figure 38 shows the pattern, repeated on the other lakes, where scenarios 2 and 4 track well together, with Scenario 1 about 2 m lower. Each of the scenarios exhibits much greater variability than the base climate. This is also illustrated in Table 32 by a greater than twofold increase in the scenario standard deviations and range of levels as compared with the base conditions. This comparison effectively demonstrates the relatively low climatic variability of the northern Great Lakes region when compared with much of the rest of the country. No record high lake levels would be set on Lake Superior for any of the scenarios.

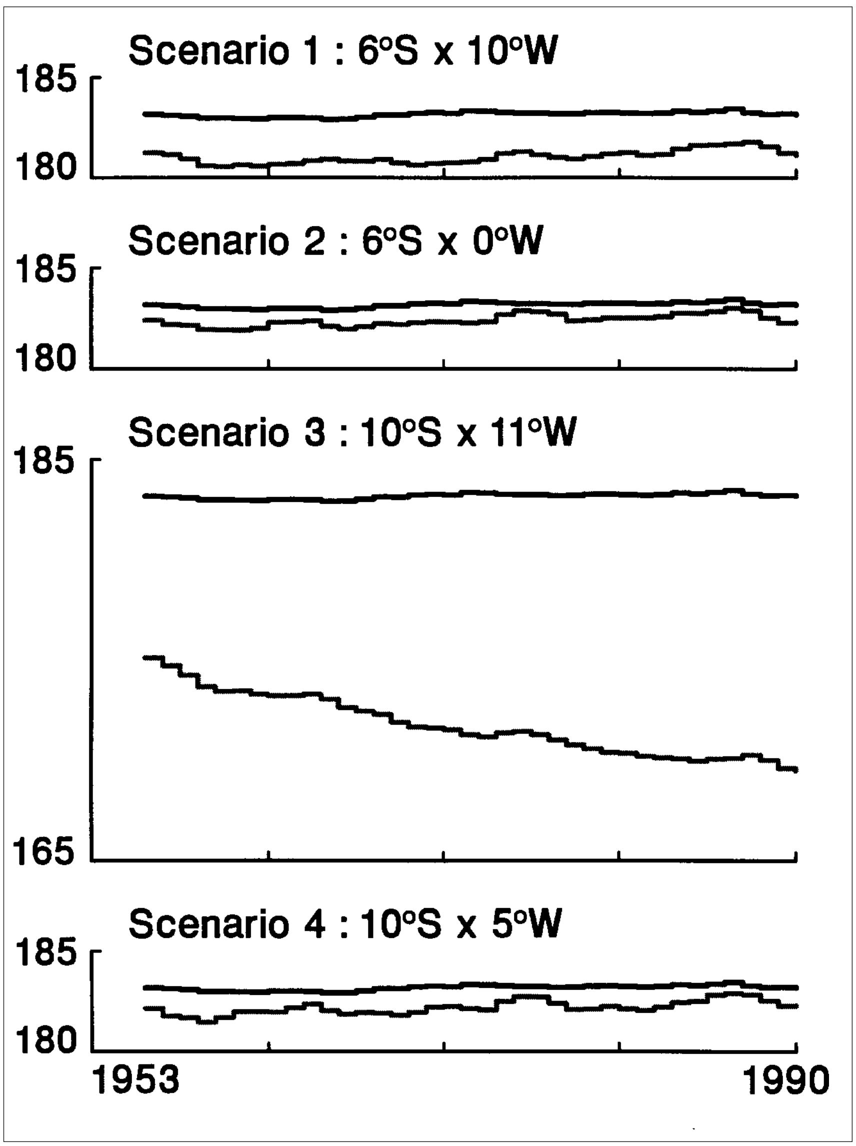


Figure 38.--Annual Lake Superior Water Levels (m), 1953-1990. Black - Base; Gray - Scenario.

Table 31.--Average Annual Steady-State Lake Level (IGLD55[a]) and Outflow Differences.

Basin	Lake Water Level (m) & Transferred Climate Absolute Differences[b]						Lake Outflow (m^3s^{-1}) & Transferred Climate Relative Change[b]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	183.17	-2.12	-0.75	-[c]	-0.97	-[c]	2390	-96%	-48%	-[c]	-58%	-[c]
Michigan	176.65	-3.33	-0.23	-3.49	-0.23	-3.36	-[d]	-[d]	-[d]	-[d]	-[d]	
Huron	176.65	-3.33	-0.23	-3.49	-0.23	-3.36	5818[d]	-66%[d]	-9%[d]	-68%[d]	-7%[d]	
Georgian	176.65	-3.33	-0.23	-3.49	-0.23	-3.36	-[d]	-[d]	-[d]	-[d]	-[d]	
St. Clair	175.22	-2.59	-0.08	-2.74	-0.11	-2.63	5973	64%	-8%	-65%	-6%	
Erie	174.32	-2.14	+0.01	-2.29	-0.04	-2.19	6642	-56%	+1%	-59%	-1%	
Ontario	74.66	-1.48	-0.03	-1.51	+0.03		7848	-48%	-1%	-48%	-4%	

[a]International Great Lakes Datum of 1955.
[b]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.
[c]Lake Superior becomes a terminal lake under these conditions.
[d]Outflow is from Lakes Michigan and Huron (including Georgian Bay), treated as one lake hydraulically.

Table 32.--Average Annual Steady-State Lake Level and Outflow Variability Differences.

Basin	Lake Water Level Std. Dev. (m) & Transferred Climate Absolute Differences[b]						Lake Outflow Std. Dev. (m^3s^{-1}) & Transferred Climate Relative Change[b]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	0.14	+136%	+100%	-[b]	+143%	-[b]	249	-45%	+12%	-[b]	+44%	-[b]
Michigan	0.38	+50%	-5%	+13%	+47%		-[c]	-[c]	-[c]	-[c]	-[c]	
Huron	0.38	+50%	-5%	+13%	+47%		511[c]	-10%[c]	-2%[c]	-39%[c]	+48%[c]	
Georgian	0.38	+50%	-5%	+13%	+47%		-[c]	-[c]	-[c]	-[c]	-[c]	
St. Clair	0.32	+47%	-6%	+22%	+50%		526	-10%	-5%	-38%	+48%	
Erie	0.29	+45%	-7%	+28%	+45%		608	-5%	-3%	-22%	+50%	
Ontario	0.09	+433%	0%	+411%	+144%		741	+1%	+2%	-5%	+42%	

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.
[b]Lake Superior becomes a terminal lake under these conditions.
[c]Outflow is from Lakes Michigan and Huron (including Georgian Bay), treated as one lake hydraulically.

Table 33.--Average Annual Steady-State Lake Level Range and Variability Differences.

Basin	Lake Water Level (m) & Transferred Climate Absolute Differences[b]						Lake Outflow (m^3s^{-1}) & Transferred Climate Relative Change[b]					
	BASE	#1	#2	#3	#4	#5	BASE	#1	#2	#3	#4	#5
Superior	0.31	+39%	+19%	-[b]	+29%	-[b]	0.08	+38%	+13%	-[b]	+75%	-[b]
Michigan	0.34	+29%	+18%	+24%	+59%		0.08	+88%	+50%	+117%	+125%	
Huron	0.34	+29%	+18%	+24%	+59%		0.08	+88%	+50%	+117%	+125%	
Georgian	0.34	+29%	+18%	+24%	+59%		0.08	+88%	+50%	+117%	+125%	
St. Clair	0.30	+50%	+37%	+60%	+97%		0.07	+143%	+86%	+129%	+143%	
Erie	0.38	+55%	+42%	+71%	+89%		0.08	+150%	+113%	+113%	+163%	
Ontario	0.56	+32%	+7%	+64%	+38%		0.14	+71%	+21%	+107%	+57%	

[a]Scenario #1 is 6ºSx10ºW; #2 is 6ºSx0ºW; #3 is 10ºSx11ºW; #4 is 10ºSx5ºW; #5 is #3 with lake effects.
[b]Lake Superior becomes a terminal lake under these conditions.

This sensitivity of Lake Superior to scenarios 1 and 3 is due primarily to it being the only lake that experienced lower precipitation for any scenario than for the base case. Lake Superior is the western-most lake in the system. A major precipitation gradient running from north to south just west of the lake results in a much dryer transposed climate, from western transpositions, than occur for the more eastern-most lakes yielding dramatically reduced, sometimes negative, net basin water supplies.

In general the seasonal cycle for the transposed scenarios is more pronounced than for the base case as shown in Figure 39 for scenario 2. The average range between maximum and minimum yearly values as well as the standard deviation of the range show small to moderate increases; see Figure 40 and Table 33. The months in which the average annual high and low levels occur remain the same for both the base and each scenario with the exception of scenario 4 where the peak month moves from September to August. In many cases a 1 month shift in the seasonal high or low is not significant when using monthly mean data.

Annual Lake Superior outflows for all scenarios also are given in Table 31. With Lake Superior being regulated, the outflows from the lake through the St. Marys River differ markedly for the transposed scenarios as compared to the base case. The average outflows are reduced by about 50% or greater. As is seen in Figure 41, scenario 1 has no outflow during a large portion of the time period, thus becoming intermittently a terminal lake.

6.2 Lake Michigan-Huron

The Lake Michigan-Huron water levels, in Figure 42, track much the same as for Lake Superior with scenarios 2 and 4 tracking along with the base case, while scenarios 1 and 3 track together about 3 m lower; see also Table 31. As shown in Table 32, most scenarios demonstrate greater annual variability than the base case with scenarios 1 and 4 particularly large. Scenario 2 is notable for having smaller annual variability than the base case. Scenario 4 would also result in new record levels on the lake.

The seasonal analysis, in Table 33, shows increased range and greatly increased variability for all scenarios as compared with the base. The annual maximum shifts by 2 months from August to June for scenarios 3 and 4, and the minimum shifts from February to December for scenario 4.

There are greatly reduced outflows for scenarios 1 and 3 in Table 31 while only around a 10% reduction for scenarios 2 and 4. The change in Michigan-Huron outflows through the St. Clair River is not as extreme as for the Superior outflows. Only scenario 4 has a larger range in outflows and variability than the base case; see Table 32.

6.3 Lake St. Clair

The Lake St. Clair water levels in Figure 43 and summarized in Table 31 follow the continuing pattern with scenarios 2 and 4 tracking along with the base case, while scenarios 1 and 3 track together about 2 m lower. Scenario 4 would set new record highs, while scenario 1 approaches the base record. As shown in Tables 32 and 33, all but scenario 2 would result in increased annual variability. Scenario 4 would also result in a record low lake level.

All of the transposed scenarios result in a greatly increased seasonal range from between 10% and 100%, as seen in Table 33. The variability of the range increases from about 100% to 140%. There are also significant shifts in the timing of the annual maximum level from July to May and from the minimum level from February to November.

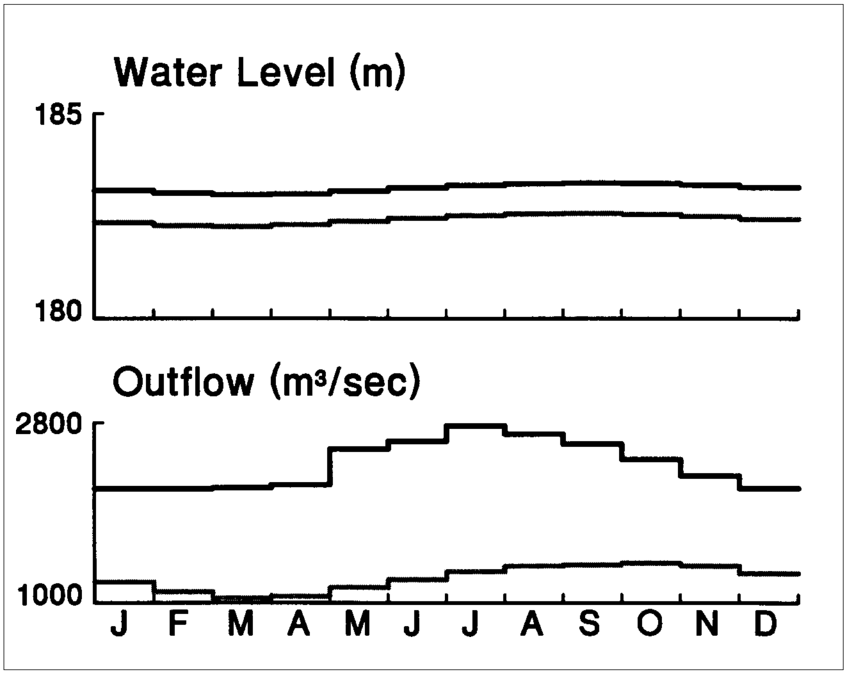


Figure 39.--Seasonal Lake Superior Average Water Levels and Outflows for Scenario 2 (6ºS×0ºW).

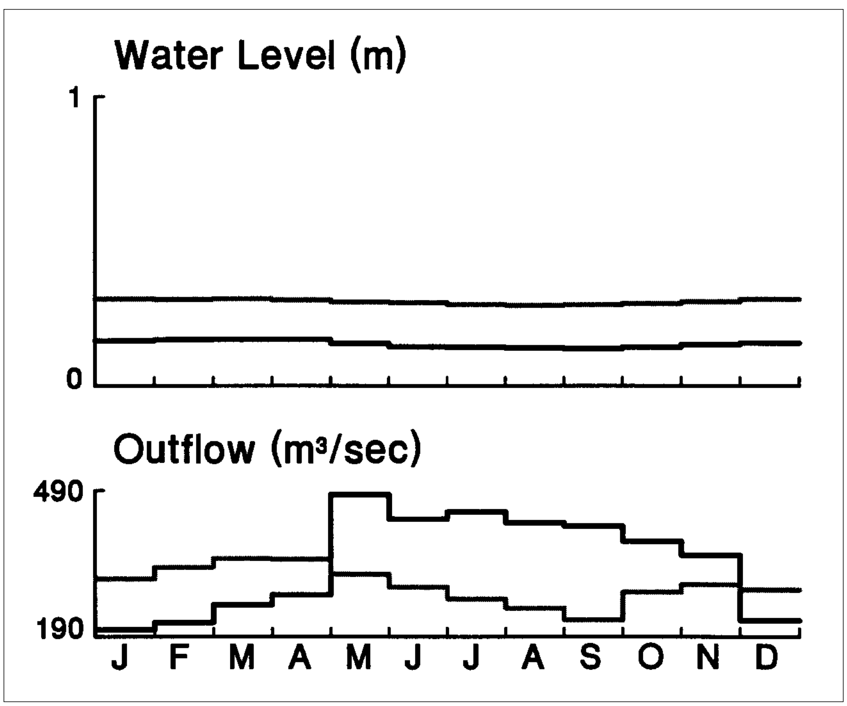


Figure 40.--Seasonal Lake Superior Water Levels and Outflows Standard Deviations for Scenario 2 (6ºS×0ºW).

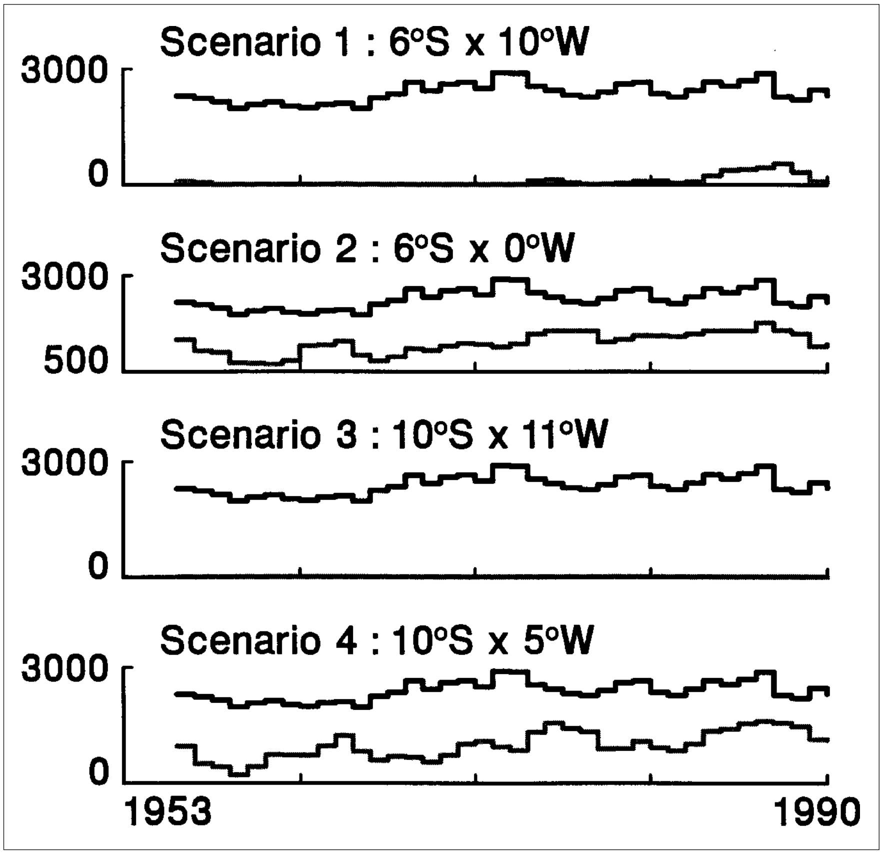


Figure 41.--Annual Lake Superior Outflows ($m^3 s^{-1}$), 1953-1990. Black - Base; Gray - Scenario.

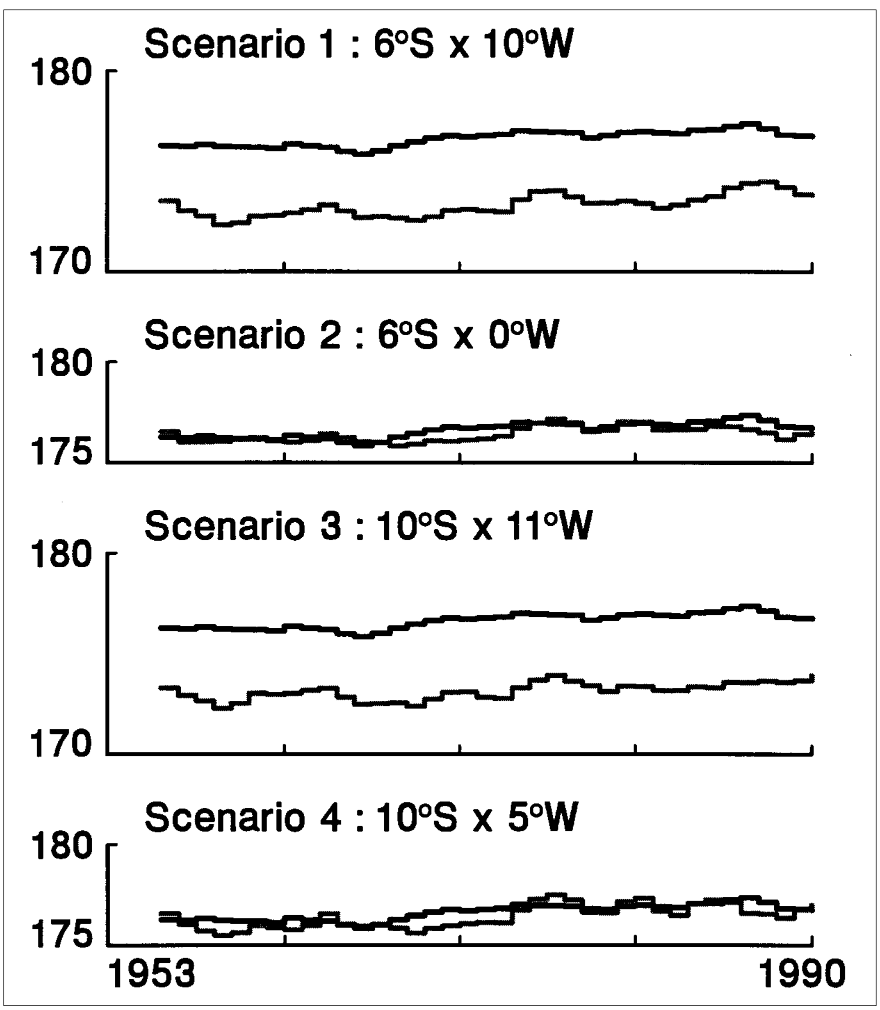


Figure 42.--Annual Lake Michigan-Huron Water Levels (m), 1953-1990. Black - Base; Gray - Scenario.

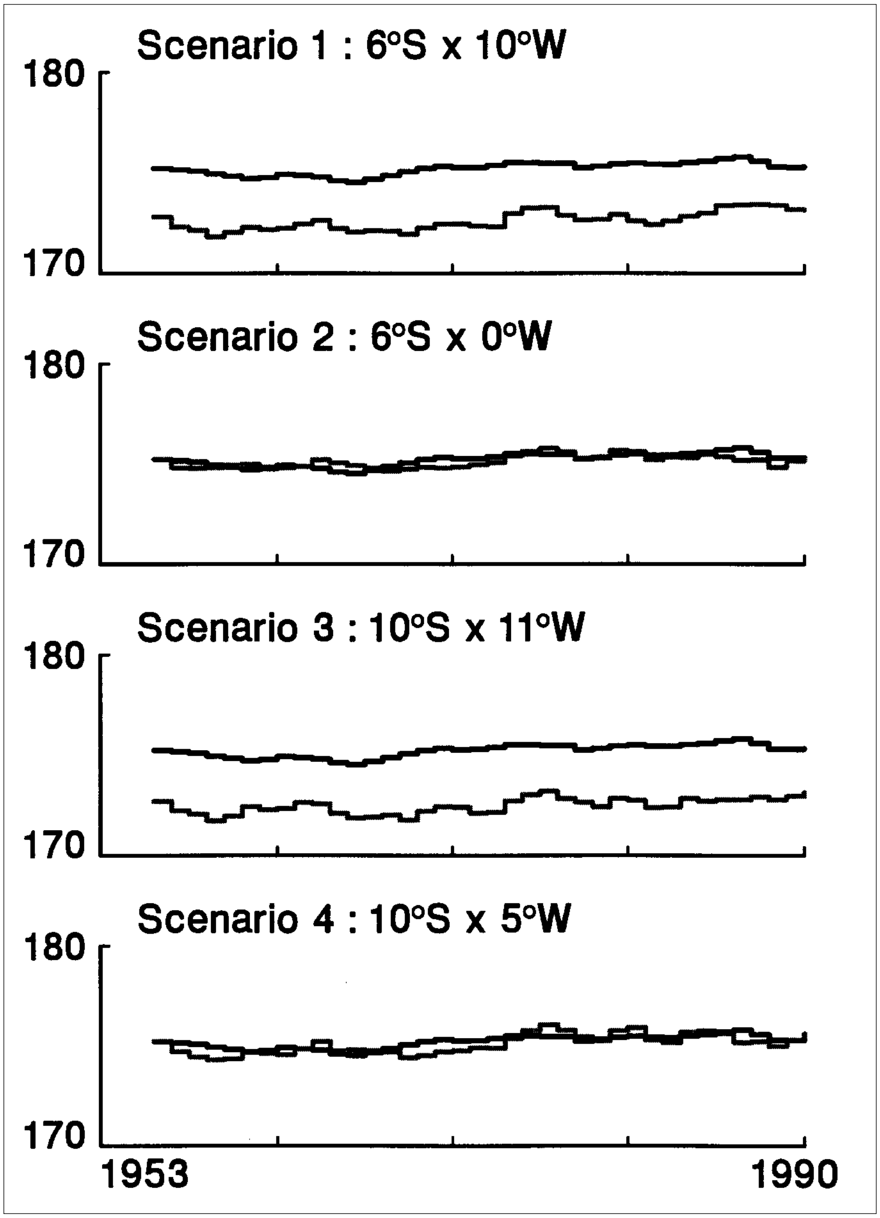


Figure 43.--Annual Lake St. Clair Water Levels (m), 1953-1990.
Black - Base; Gray - Scenario.

The Lake St. Clair outflows comparison, in Table 31, is essentially the same as for the Michigan-Huron outflows. This is because Lake St. Clair can be considered as a wide spot in the connecting channel between Lake Michigan-Huron and Lake Erie. Only slight changes in the standard deviation and range occur, as compared to Lake Michigan-Huron, resulting from local inflows into Lake St. Clair and backwater effects from Lake Erie.

6.4 Lake Erie

The Lake Erie water levels, in Figure 44, track similarly to the other lakes. New record high levels would be set by scenarios 2 and 4, while new record lows would be set by scenario 4. Both scenarios 1

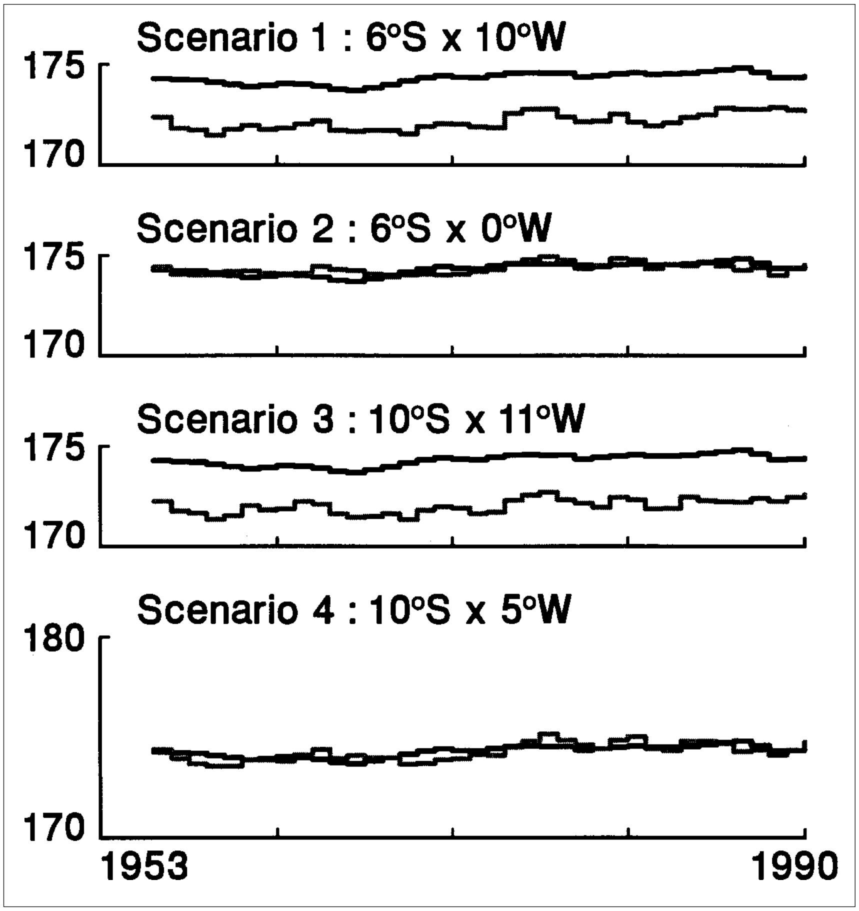


Figure 44.--Annual Lake Erie Water Levels (m), 1953-1990. Black - Base; Gray - Scenario.

and 3, as shown in Table 31, are consistently about 2.2 m below the base case. All but scenario 2 show increased variability and range in lake levels over the base case; see Tables 32 and 33.

Lake Erie, in Table 33, demonstrates a greatly enhanced seasonal cycle under all transposed climates with greatly increased range and variability of the range. For scenario 4, the increase in range is about 90% with an increased standard deviation of about 160%. However the timing of the seasonal maximum and minimum remain about the same.

The Lake Erie outflows, in Table 31, show the results of greatly enhanced precipitation over the Lake Erie basin for scenarios 2 and 4. The range in outflows for scenario 4 increases by about 50% over the base case while decreasing by about 25% for scenarios 1 and 3, as seen in Table 33. The variability, however, is increased only in scenario 4.

6.5 Lake Ontario

Lake Ontario is an interesting case because of the complexity and reaction of the regulation plan to major changes in water supplies. As mentioned earlier, the version of the regulation plan used in this study has been modified to make it more robust to changes in both high and low water supplies. This is illustrated in Figure 45 where in years 1956–1958 and 1967–1969 sharp rises in water levels occur as artifacts of the regulation. This results from holding back water on the lake; see Figure 46. It is also interesting that scenario 2 has similar level characteristics as the base case, while scenario 4 has about the same annual mean but over twice the variability and range as the base case; see Tables 31, 32, and 33.

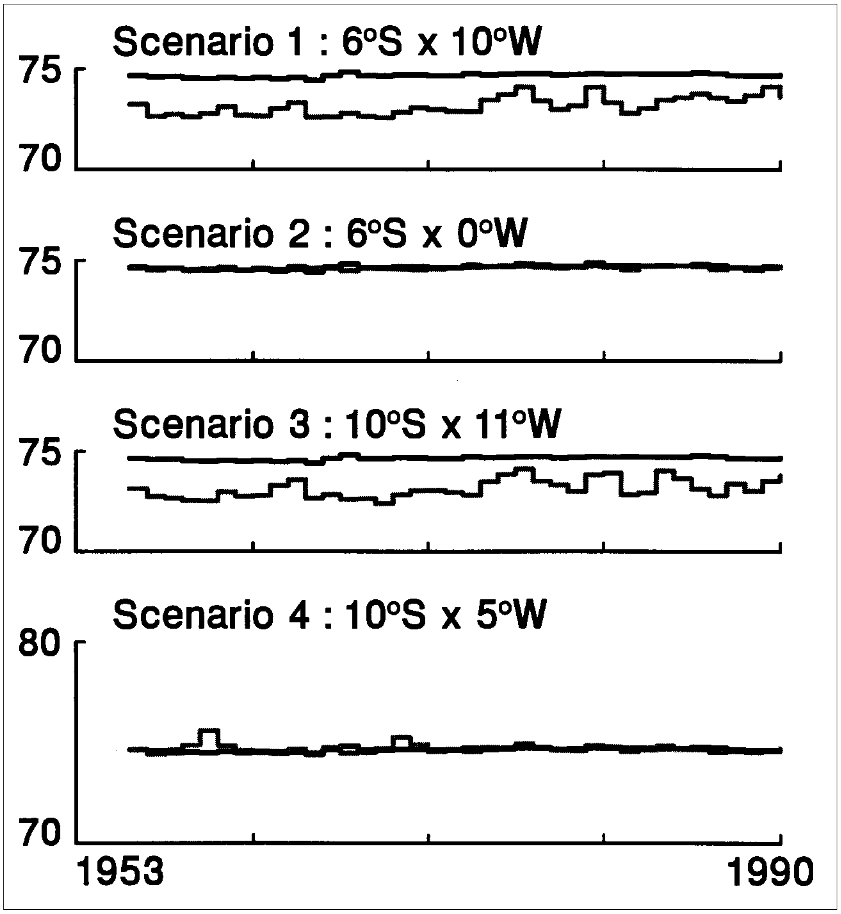


Figure 45.--Annual Lake Ontario Water Levels (m), 1953-1990. Black - Base; Gray - Scenario.

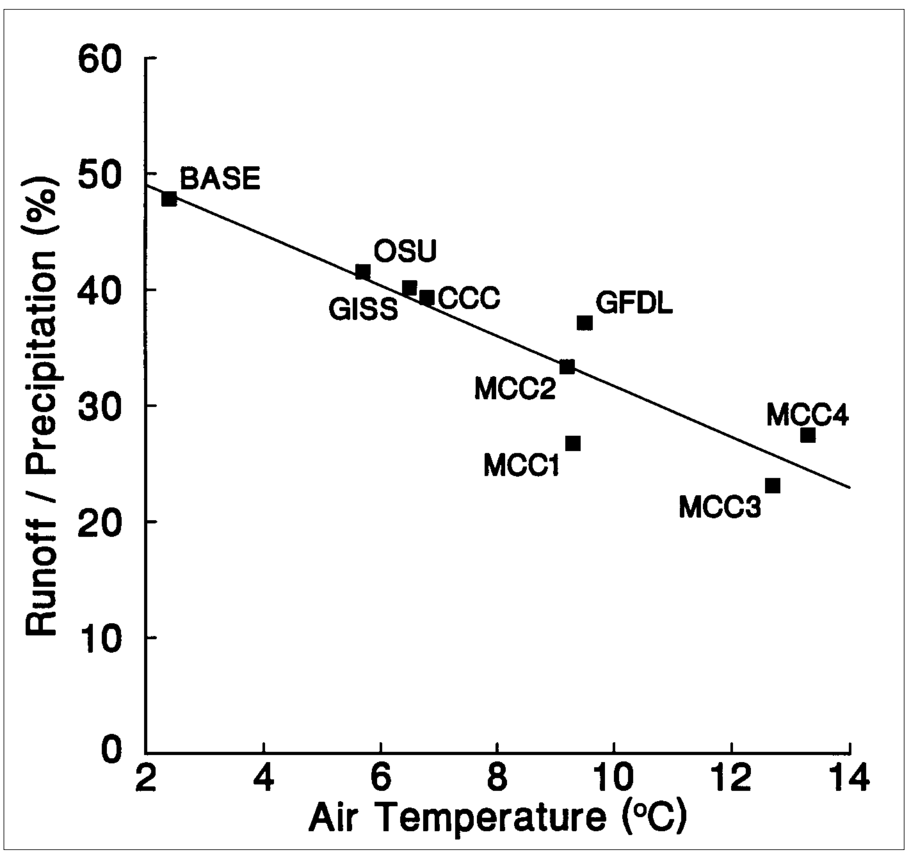


Figure 46.--Annual Lake Ontario Outflows ($m^3\ s^{-1}$), 1953-1990. Black - Base; Gray - Scenario.

Record high water levels would occur under both scenarios 2 and 4. Extreme range and variability are also noted for the low supply scenarios 1 and 3. This is particularly noticeable during the 1972–1990 time period.

All transposed scenarios exhibit a more pronounced seasonal cycle than the base case; see Table 33. The magnitude of the seasonal cycle increases with the average range for scenario 3 being about 50% greater than the base case. The standard deviation of the range also increases with scenario 3 being about 100% larger than the base. The timing of the seasonal cycle remains about the same as the base case.

The Lake Ontario outflows, in Table 31, are consistent with the other outflows showing greatly decreased outflows under scenarios 1 and 3 with moderately decreased outflows under scenarios 2 and 4. The standard deviations are similar to the base case for all scenarios except for scenario 4 which has a standard deviation about 50% greater than the base case. Of particular note is that the maximum and minimum flows for scenario 4 are about 10% higher and lower respectively than the base case.

6.6 Summary

The impacts on levels and flows were slightly more severe under scenario 4 than for scenario 2; see Table 31. The upper lakes were down slightly; most were down by less than 1 m. The levels of Lakes St. Clair and Erie were virtually unchanged, as compared to the Base run. Lake Ontario was down slightly for scenario 2 and up the same amount (.03 m) in scenario 4. It is worth noting that even under scenario 2, the least extreme of all four scenarios, the routing models predict that Lake Superior would be down 0.75 m and Lake Michigan-Huron would be lower by over 0.2 m.

The effects on outflows were also more pronounced in scenario 4 than for scenario 2. The flow in the St. Marys River was cut in half in both scenarios. The outflows from Lake Michigan-Huron, St. Clair, and Ontario were reduced slightly. The outflow from Lake Erie was almost unchanged in both scenarios.

The impacts on levels and flows were more dramatic for scenario 3. The most severe impacts are on Lake Superior, which becomes a terminal lake with no outflow. Lake Michigan-Huron drops over 3 m; Lakes St. Clair, Erie, and Ontario are down over 2 m. Outflows are about one third of normal in the St. Clair and Detroit Rivers. Flows are slightly better (about 40% of normal) in the Niagara River and just over 50% of normal in the St. Lawrence River.

Similar but less severe results are experienced under scenario 1. In this scenario, the most severe impact is on Lake Michigan-Huron, which is down over 3 m as compared to the base run. The other lakes are all down over 2 m. Again, the St. Marys River is shut off after a few years of low flow. The flows in the St. Clair, Detroit, and Niagara Rivers are about 33% of normal. The flow in the St. Lawrence River is about 50% of normal.

7. MAIN FINDINGS AND IMPLICATIONS

7.1 Interrelationship Constraints

Scenarios 1 and 2 represent the upper end of the range of GCM predictions for temperature increases under doubled-CO_2 conditions. Scenarios 3 and 4 represent conditions much more extreme than any GCM prediction of temperatures for doubled-CO_2 conditions. These rather extreme scenarios were purposely chosen to clearly delineate the direction of the response of the Great Lakes to climate change. We recognize that the climate transposition approach has limitations. A principal limitation is that several of the climatic variables that influence the lake are interrelated. Major interrelations are:

- Precipitation generally increases with temperature; therefore in most basins and most scenarios, these scenarios represent wetter conditions. This is not inconsistent with recent GCM results, which generally predict that precipitation will either remain the same or increase slightly in the Great Lakes basin.

- Atmospheric water vapor content increases with precipitation. These two variables are highly correlated in the present climate of eastern North America.

- These interrelationships constrain the range of conditions investigated in this study. However, despite this limitation, an analysis of the results for individual lakes reveals that a wide range of conditions were sampled using the approach chosen for this study.

7.2 Lake Evaporation Increases

All scenarios produce significant increases in lake evaporation. From an energy standpoint, the energy for increased evaporation is derived from four sources. Most important is the decrease in cloud cover that results in increased incoming solar radiation and, on average, accounts for about half of the evaporation increase. The second source is increased downward long-wave radiation emitted by the atmosphere, a result of the higher temperatures; this accounts for 10–15% of the effect. These increased radiative sources result in a greater accumulation of heat in the lakes. A third important factor is a change in the partitioning of energy between sensible and latent heat flux. As a result of the Clausius-Clapeyron relationship, the ratio of sensible to latent heat flux (Bowen ratio) decreases in all scenarios and accounts for about one third of the effect. A fourth factor is a decrease in lake ice cover. Because of the higher temperatures, nearly all lakes remain ice-free throughout the winter. Thus, the average albedo during the winter and early spring months is lower, increasing the amount of solar radiation absorbed by the lake. However, this makes a minor (1–2%) contribution.

The increase in downward long-wave radiation, the change in the partitioning between sensible and latent heat flux, and the decrease in lake ice cover result from fundamental physical principals, and they will almost certainly be a feature of any climatic state warmer then current conditions in the basin. Thus, there will be a considerable positive pressure on lake evaporation. However, the changes in cloud cover in these scenarios may not be realized in a future warmer climate. Thus, the increases in lake evaporation in these scenarios would be smaller if cloud cover does not decrease. However, most GCM predictions do result in significant increases in lake evaporation (Croley, 1990, 1992b, 1993a).

Another interesting aspect of lake evaporation is that it is highly event oriented. A large proportion of evaporation occurs during Arctic cold air outbreaks in the cold season. In all scenarios for all lakes, the relative contribution of these events to total lake evaporation increases. Although the increase may be unique to these specific scenarios, these events are also important in the current climate. This indicates that accurate future estimates of lake evaporation will require accurate estimates of the number and severity of cold air outbreaks.

7.3 Soil Moisture and Runoff Reductions

Many scenarios result in lower soil moisture and reduced runoff despite higher precipitation. As a result of the higher temperatures and longer growing seasons, the four scenarios produced a more vigorous over-land hydrological cycle. Total annual evapotranspiration from the ground and the vegetation increases in all scenarios. Also, higher temperatures significantly reduce total snowfall. In the current climate, the spring snowmelt runoff season is very important to the total lake hydrology. In the four scenarios, the snowmelt season is shorter and less significant. The above results are likely to be a feature of any warmer climate. This means that in a warmer climate, greater precipitation is required to maintain runoff at present levels.

Figures 47 and 48 illustrate the interrelationships between air temperature and runoff/precipitation ratios and lake evaporation changes for the Lake Superior basin, the most sensitive basin for this study. The figures also include data from the previously mentioned EPA and IJC studies. The relationship among these variables is amazingly linear. The figure shows that substantial increases in precipitation would be required to maintain the runoff equivalent to the present climate. For example, a change in temperature of 5–6ºC requires an increase in precipitation of 20–30% to maintain current runoff levels.

7.4 Net Basin Supply Decreases

Warmer climates result in large negative pressures on net basin water supply. Net basin supply (NBS) is comprised of the sum of over-lake precipitation and surface runoff into the lake, minus lake evaporation. The previous two findings have indicated that warmer climates will likely lead to increases in lake evaporation and decreases in runoff. Thus, significantly greater precipitation is required to maintain NBS at current levels of the Great Lakes. A summary of the results of these scenarios for NBS on Lake Superior (Figure 49) illustrates this. There is a very coherent relationship among NBS changes and the changes in temperature and precipitation. The relationships in Figures 47, 48, and 49 suggest that for temperature changes of 5–6ºC, precipitation increases of 20–30% may be required to maintain Lake Superior NBS at today’s levels. If annual mean temperatures were to increase with no compensating increases in precipitation, it is highly likely that NBS levels would fall significantly. Of particular interest to Lake Superior is the precipitation required to maintain outflow from the lake under the various scenarios. This is illustrated in Figure 50. Precipitation below the equilibrium line may result in Lake Superior becoming a terminal lake.

7.5 Net Basin Supply Variability Increases

These scenarios produce much higher variability in NBS. Interannual variability in NBS, expressed as an over-lake depth, ranges from 140 to 300 mm under current climate changes. These scenarios produce average increases of about 60% in warm scenarios 1 and 2 and about 90% in very warm scenarios 3 and 4. These increases are primarily due to increases in precipitation variability in these scenarios. The Great Lakes currently experience lower precipitation variability than that of locations to the west and south of the basin. Thus, all scenarios have increased precipitation variability. Kunkel et al. (1993) have pointed out that a major contributor to interannual precipitation variability in the Great Lakes region is infrequent large multi-day precipitation events. Thus, accurate estimates of precipitation variability expected in future climates will require an accurate simulation of the frequency and magnitude of these infrequent large events.

The changes in lake evaporation variability are rather small (< 20%) and are thus not a major factor in changing NBS variability. However, in some scenarios, the simulated increases in runoff variability are caused partially by higher variability in evapotranspiration. This is a result of the longer growing season that results in a greater exposure to soil moisture stress. However, this is a minor factor compared to the contribution of precipitation variability.

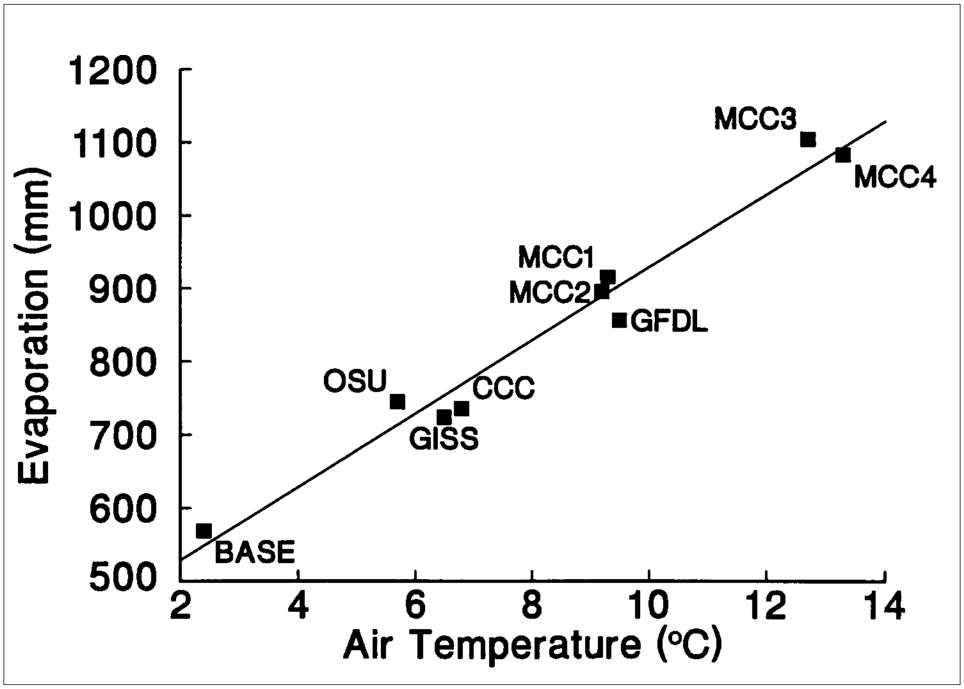


Figure 47.--Average Annual Lake Superior Basin Effective-Runoff Variation with Temperature.

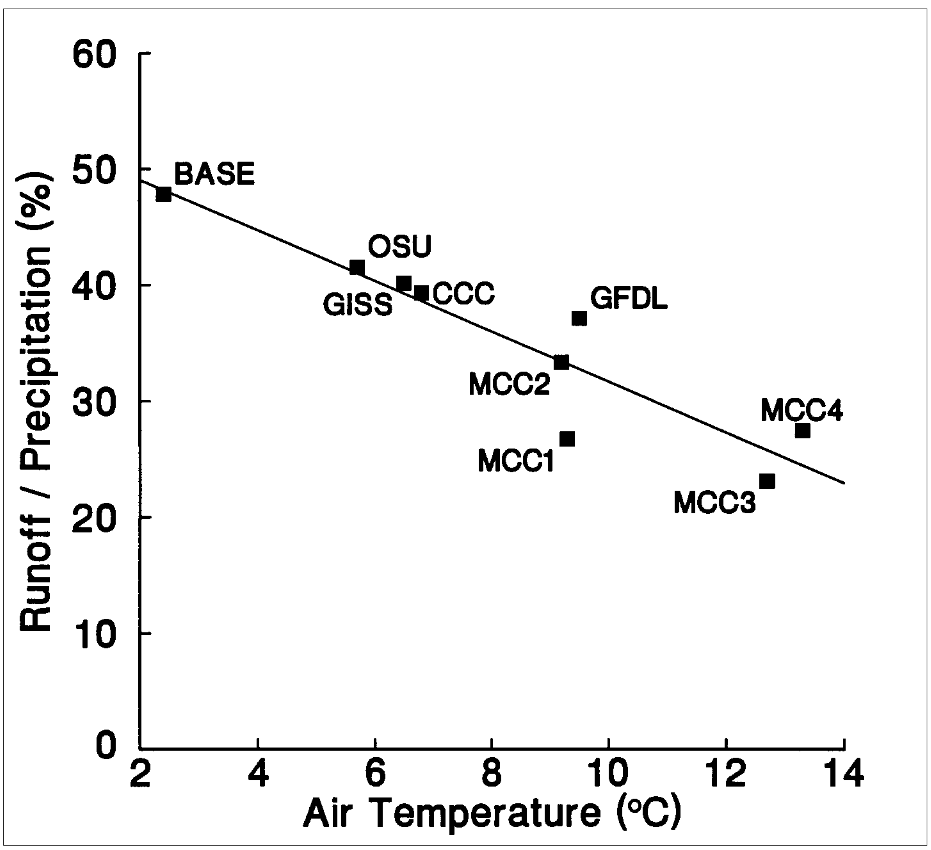


Figure 48.--Average Annual Lake Superior Evaporation Variation with Temperature.

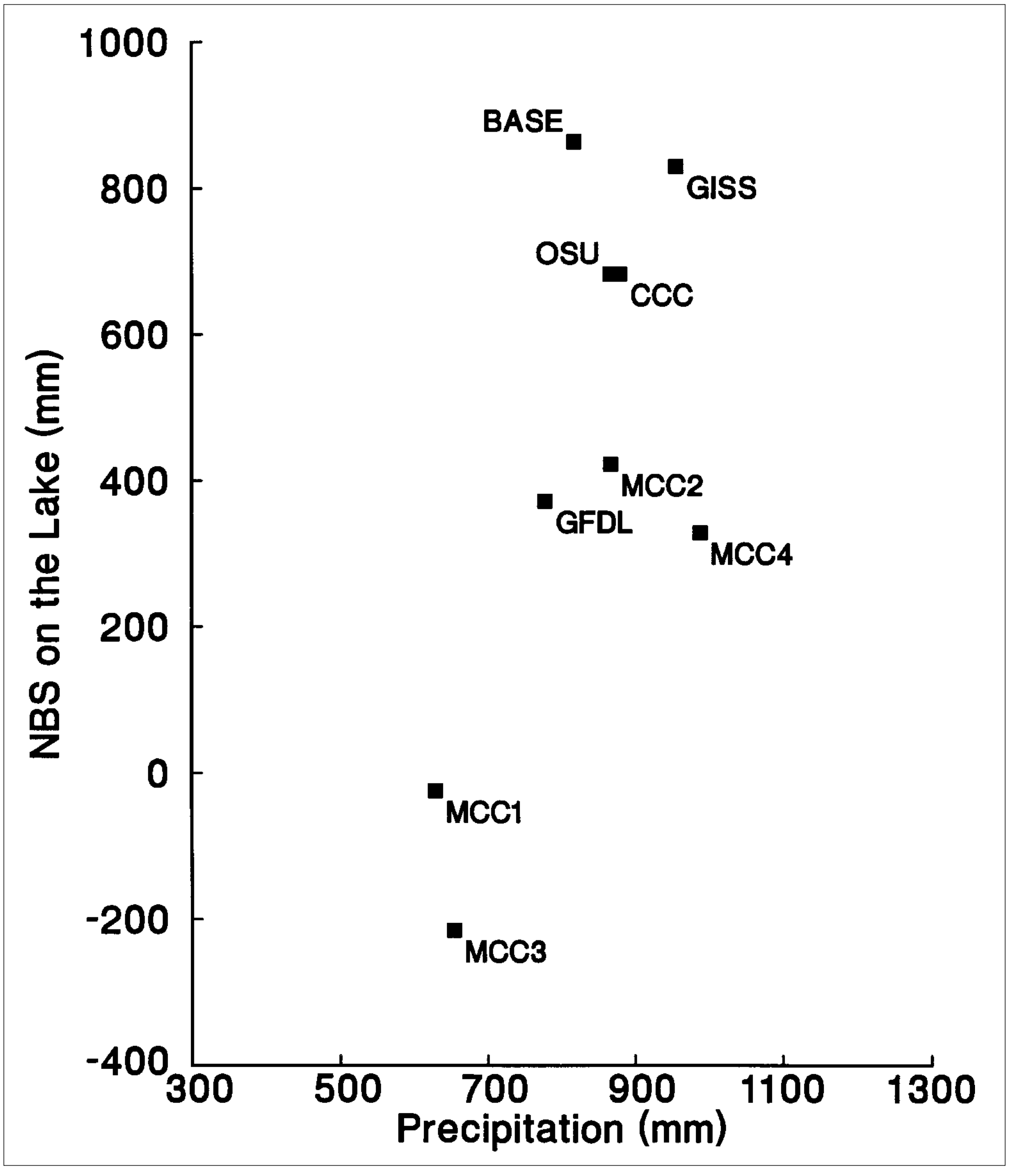


Figure 49.--Average Annual Lake Superior Net Basin Supply Variation with Precipitation.

7.6 Reduced Turnover Frequency

Warmer climates can result in reduced frequency of buoyancy-driven water column turnovers. In many of these scenarios, lake surface water temperatures often do not fall to 3.98ºC (the temperature of the maximum density of water) during the colder half-year. As a result, buoyancy-driven vertical turnovers of the water column change from a frequency of two times per year to once per year. Since this is related to a fundamental physical property of fresh water, it is highly likely that this will occur in any future climate that is sufficiently warm. This could result in significant environmental impacts, since these turnovers are important for nutrient distribution, oxygenation of lake water, and so forth.

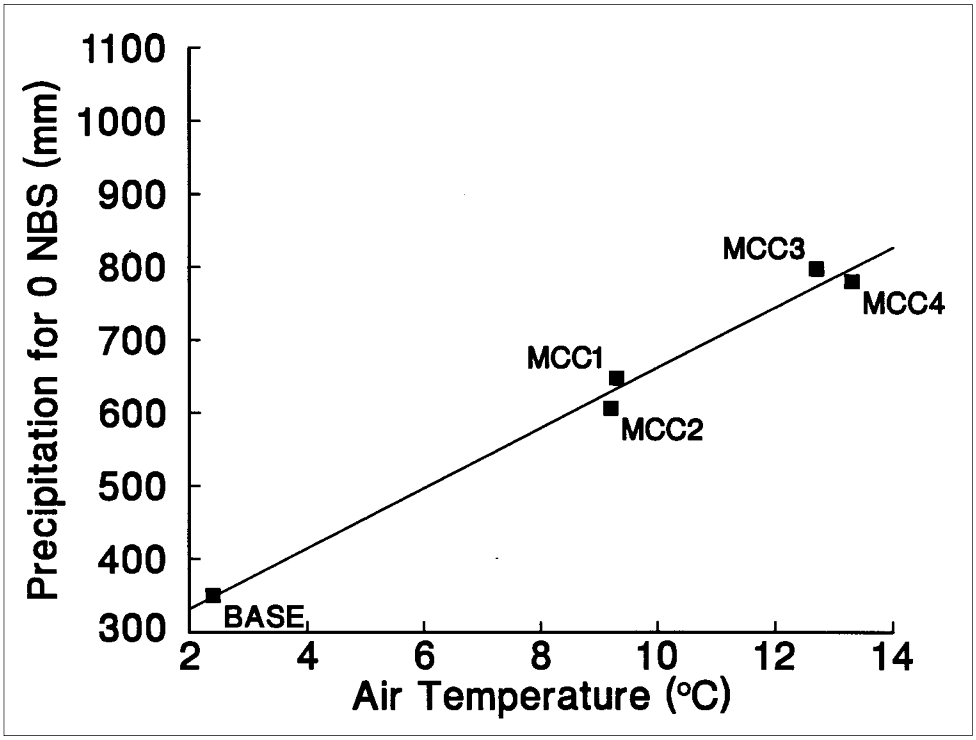


Figure 50.--Average Annual Lake Superior Precipitation Required for Zero Net Basin Supply Variation with Temperature.

7.7 Lake Effects

Lake effects on regional climate have negligible hydrological effects. We utilized existing spatial and quantitative measures of lake effects on various climate conditions to modify climate data for one of the scenarios. We tested lake effects on basin hydrology by calculating outcomes with and without lake effects present. The differences on runoff and lake levels for the various Great Lake basins were negligible. We did not attempt a modeling investigation to ascertain how much lake effects might change in warm-wet scenarios like 3 or 4, but the lack of differences suggest that huge changes in lake effects would be required to significantly alter the hydrological results.

7.8 Lake Levels and Outflows

Great Lakes water levels are lower and more variable under the transposed climates. A warmer climate over the Great Lakes basin, whether wetter or dryer, would have major impacts on Great Lakes water levels and flows in the connecting channels. Superior is the most sensitive lake in the system when looking at climate transposition. Under scenarios 1 and 3, the lake would have negative water supplies for all or part of the time. Under scenario 3, Lake Superior would become a terminal lake. In other scenarios, both record high and record low lake levels would be achieved. In addition most scenarios indicated greater variability of lake levels, both seasonal and interannual, than exist in the present regime.

From an adaptive viewpoint, the individual lakes, with the exception of Lake Superior during negative water supply scenarios, could be regulated to maintain water levels at about the long term average. However, there is no way to maintain the connecting channel flows about their long means. There would also likely be a major effort to divert additional water into Lake Superior from the Hudson Bay watershed under low or negative water supply scenarios.

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9. NOTATION

a	=	wind parameter, T > 3.98°C [evaporation model empirical parameter]
a'	=	wind parameter, T < 3.98°C [evaporation model empirical parameter]
a_s	=	proportionality constant for snowmelt per degree-day [runoff model empirical parameter]
A	=	area of the ice surface
A_b	=	area of the watershed
Aw	=	area of the open-water (ice-free) lake surface
α_{dp}	=	deep percolation coefficient [runoff model empirical parameter]
α_{gw}	=	groundwater coefficient [runoff model empirical parameter]
α_{int}	=	interflow coefficient [runoff model empirical parameter]
α_{per}	=	percolation coefficient [runoff model empirical parameter]
α_{sf}	=	surface outflow coefficient [runoff model empirical parameter]
b	=	wind parameter, T > 3.98°C [evaporation model empirical parameter]
b'	=	wind parameter, T < 3.98°C [evaporation model empirical parameter]
b_1	=	empirical constant
b_2	=	empirical constant
β	=	partial linear reservoir coefficient
β_{eg}	=	groundwater zone evapotranspiration coefficient [runoff model empirical parameter] (= 0)
β_{el}	=	lower zone evapotranspiration coefficient [runoff model empirical parameter] (= 0)
β_{es}	=	surface zone evapotranspiration coefficient [runoff model empirical parameter]
β_{eu}	=	upper zone evapotranspiration coefficient [runoff model empirical parameter]
C_i	=	specific heat of ice
C_p	=	specific heat of air at constant temperature
C_w	=	specific heat of water
C_E	=	bulk evaporation coefficient over water
C_E'	=	bulk evaporation coefficient over ice
C_H	=	sensible heat coefficient over water
C_H'	=	sensible heat coefficient over ice
D	=	ice pack depth (thickness)
DD	=	degree-days per day
Δ	=	time increment of mass balance computation period
e	=	evaporation or evapotranspiration rate
e_p	=	rate of evaporation or evapotranspiration still possible
e_w	=	over-water evaporation rate
e_w'	=	over-ice evaporation rate
E	=	volumetric rate of evaporation from ice
E_g	=	evapotranspiration from the groundwater zone storage
E_l	=	evapotranspiration from the lower soil zone storage
Es	=	evapotranspiration from the surface storage

E_u = evapotranspiration from the upper soil zone storage
ε_w = emissivity of water
ε_a = emissivity of the atmosphere
f = infiltration rate
$f_{k,m}$ = ratio of surface temperature rise on day k from heat added on day m to that heat addition
F = representing lake volume at which a heat addition is uniformly fully mixed, T > 3.98°C [evaporation model empirical parameter]
F' = representing lake volume at which a heat addition is uniformly fully mixed, T < 3.98°C [evaporation model empirical parameter]
GZM = content of groundwater zone
γ_f = latent heat of fusion
γ_v = latent heat of vaporization
H = heat stored in the lake
H' = heat stored in the ice pack
H_s = nonlatent heat released to the atmosphere
η = parameter relating cloudiness to atmospheric long-wave radiation [evaporation model empirical parameter]
K = units and proportionality constant
$LSZM$ = moisture content of lower soil zone
m = daily snowmelt rate
m_p = daily potential snowmelt rate
$M_{k,m}$ = mixing volume size on day k of heat added on day m
n = number of days in the mass balance computation period
n_s = daily net supply rate to the watershed surface
N = fraction of sky covered in clouds
p = precipitation rate
q = specific humidity of the air over the water
q' = specific humidity of the air over the ice
q_0 = unit (per unit area) cloudless sky short-wave radiation rate
q_e = unit evaporative (latent and advected) heat transfer rate
qe' = unit evaporative (latent and advected) heat transfer rate from ice pack
qh = unit sensible heat transfer rate
q_h' = unit sensible heat transfer rate to ice pack
q_i = unit incident short-wave radiation rate
q_p = unit precipitation heat advection rate to water surface
q_p' = unit precipitation heat advection rate to ice pack
q_r = unit reflected short-wave radiation rate
q_r' = unit reflected short-wave radiation rate to ice pack
q_w = specific humidity of saturated air at temperature of water
q_w' = specific humidity of saturated air at temperature of ice
$q\uparrow$ = long-wave radiation emitted by the water body
$q\downarrow$ = long-wave radiation from the atmosphere absorbed by the water surface
Q = basin outflow volume for n days
Q_a = heat flux between atmosphere and ice pack used for freezing or melting
Q_l = net long-wave radiation exchange rate
Q_w = total heat flux between the water body and the ice pack
Q_I = net heat advection to the lake from surface flows

θ_i = sum of all surface inflows to lake
θ_o = sum of all outflows from lake
r_a = reflectivity of the water surface
r_r = daily solar insolation at the watershed surface
ρ = density of ice
ρ_a = density of air
ρ_w = density of water
S = volumetric rate of snow falling on ice
SNW = water content of the snowpack
SS = content of surface storage zone
σ = Stephan-Bolzman constant
t = time
T = water surface temperature
T' = ice surface temperature
T_a = air temperature
T_a' = over-ice air temperature
T_b = a base scaling temperature [runoff model empirical parameter]
T_{max} = maximum daily air temperature
T_{min} = minimum daily air temperature
τ = daily extra-terrestrial solar radiation
τ_a = parameter reflecting ice pack shape, vertical-lateral change ratios along atmosphere-ice boundary, and ice buoyancy [evaporation model empirical parameter]
τ_w = parameter reflecting ice pack shape, vertical-lateral change ratios along water-ice boundary, and ice buoyancy [evaporation model empirical parameter]
U = wind speed over water
$USZC$ = capacity of the upper soil zone (= 2 cm)
$USZM$ = moisture content of upper soil zone
V = volume of the ice pack
V' = volume of ice formed by only by freezing or melting
V_c = lake volume (capacity)
V_e = equilibrium lake volume approached as a limit by mixing [evaporation model empirical parameter]
W = daily wind movement
X = ratio of hours of bright sunshine to maximum possible
Ψ = total heat available for evapotranspiration during the day
Z = volume of water in storage