GREAT LAKES HYDROLOGY UNDER TRANSPOSED CLIMATES*

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Abstract. Historical climates, based on 43 years of daily data from areas south and southwest of the Great Lakes, were used to examine the hydrological response of the Great Lakes to warmer climates. The Great Lakes Environmental Research Laboratory used their conceptual models for simulating moisture storages in, and runoff from, the 121 watersheds draining into the Great Lakes, over-lake precipitation into each lake, and the heat storages in, and evaporation from, each lake. This transposition of actual climates incorporates natural changes in variability and timing within the existing climate; this is not true for General Circulation Model-generated corrections applied to existing historical data in many other impact studies. The transposed climates lead to higher and more variable over-land evapotranspiration and lower soil moisture and runoff with earlier runoff peaks since the snow pack is reduced up to 100%. Water temperatures increase and peak earlier. Heat resident in the deep lakes increases throughout the year. Buoyancy-driven water column turnover frequency drops and lake evaporation increases and spreads more throughout the annual cycle. The response of runoff to temperature and precipitation changes is coherent among the lakes and varies quasi-linearly over a wide range of temperature changes, some well beyond the range of current GCM predictions for doubled CO_2 conditions.

1. Introduction

If climate warming should occur, it will impact Great Lakes water supply components and basin storages of water and heat; these impacts must be understood before secondary impacts can be assessed. Climate variability is of particular concern since it is a key problem for shipping, power production, and resource management. The Great Lakes Environmental Research Laboratory (GLERL) have been using their Great Lakes hydrology models with adjusted data sets, derived from general circulation model (GCM) outputs, to assess regional effects of climate changes. However, these studies revealed little about the sensitivity of water supplies to changes in variability (Kunkel et al., 1998) since spatial and temporal variabilities were the same (or proportional to each other) in the adjusted data sets as in the historical data. Furthermore, the use of GCM outputs forced the use of inappropriately large spatial scales for studying Great Lakes climate change impacts.

The companion paper (Kunkel et al., 1998) describes the rationale for this study and the climates used as input to the hydrologic models. To briefly summarize,

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the objectives were twofold. One was to determine the sensitivity of Great Lakes water supply variability to changes in climate variability. The second was to gain a more fundamental understanding of the response of the system to different climate conditions, including conditions well beyond any current GCM prediction of future climate change, in order to identify thresholds of climate conditions where impacts on water supplies become severe or where the response of the system becomes nonlinear. To achieve the wide range of climates desired in this study, a transposition approach using data from stations to the south and west of the Great Lakes Basin, was employed. Four distinct climate data sets were developed, each consisting of forty-five years of daily meteorology at about 1000 sites. These are physically plausible and coherent climates, characterized by realistic temporal and spatial variabilities. Lake effect corrections were estimated from the present climate and applied to one of the climates, creating a fifth climate. The Great Lakes hydrology of each transposed climate is estimated here by applying GLERL's system of models to these data and comparing the outputs to a base case derived from historical data.

The sections that follow describe Great Lakes dynamics and modeling, present the climate change assessment methodology, outline Great lakes hydrological and lake thermodynamic responses to climate change, discuss hydrological sensitivities, and summarize the findings. Representative annual means and standard deviations are selected here for only four Great Lakes (Superior, Michigan, Erie, and Ontario), two transposed climates, and 13 hydrological variables. Some additional annual results are given also for Lake Huron and another transposed climate to illustrate the most pronounced impacts of considering lake effects in the analysis. Likewise, seasonal means and standard deviations are given here for only one Great Lake, one transposed climate, and 13 variables. This partial presentation of results avoids the confusion of many numbers while illustrating the primary points; all details are available elsewhere (Croley et al., 1996). The methodology is extended to estimate impacts on net basin supplies and lake levels elsewhere (Croley et al., 1995, 1996). They will be discussed further in an upcoming paper.

2. Great Lakes Dynamics and Modeling

The behavior of the Laurentian Great Lakes system (Kunkel et al., 1998, Figure 1) is governed by its huge capacities for water and energy storage. Lakes Superior, Michigan, Huron, and Ontario are very deep, while Lakes Erie and St. Clair are very shallow. Table I contains descriptive statistics on the sizes of the Great Lakes, Lake St. Clair, and their basins. Mass balance on the basins comprises the primary process determining lake levels, via the hydrological cycle of the Great Lakes Basin. GLERL applied their daily rainfall-runoff models to the 121 watersheds about the Great Lakes (Croley et al., 1983a,b). Daily precipitation, air temperature, and insolation are used to determine snow pack accumulations and net surface supply based on degree-day determinations of snow melt. The net surface supply

Table I

Laurentian Great Lakes size statistics[a]

Characteristic	Superior	Michigan	Huron[c]	St. Clair	Erie	Ontario
Basin area,[b] km^2	128,000	118,000	131,000	12,400	58,800	60,600
Lake area, km^2	82,100	57,800	59,600	1,114	25,700	18,960
Volume, km^3	12,100	4,920	3,540	3	484	1,640
Average depth, m	147	85	59	3	19	86
Maximum depth, m	405	281	229	6	64	244

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

is divided into infiltration to soil zones and surface runoff by taking infiltration proportional to the net surface supply rate and to the areal extent of the unsaturated portion of the soil zone. Outflow from each storage within the watershed is taken as proportional to the moisture in storage. The evapotranspiration rate from the soil zones is proportional to available moistures there and to the heat rate available for evapotranspiration; it also reduces the heat available for subsequent evapotranspiration. The total amount of heat in a day is split between that used and that still available for evapotranspiration by empirical functions of air temperature based on a long-term heat balance. Mass continuity yields a first-order linear differential equation for each of the moisture storages, which are tractable analytically; they are solved simultaneously to determine daily moisture storage, evapotranspiration, and basin runoff. Because of the buffering effect of the large snow pack and moisture storages, runoff from rivers into a lake can remain high for many months 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. The outflows consist of evaporation from the lake surface, flow to downstream lakes, and diversions. 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 is also important and must be considered together with a lake's mass balance. Current Great Lakes evaporation studies use mass transfer formulations that include atmospheric stability effects on the bulk transfer coefficients, applied to monthly data for water surface temperatures, wind speed, humidity, and air temperatures. GLERL used that approach here applied to daily data but combined with models for over-water meteorology, surface heat fluxes, ice cover, lake heat storage, and lake heat balance (Croley, 1989, 1992). As overwater data are not generally available, over-land data are used by adjusting for over-water conditions. Air temperatures and specific humidities over ice are used for over-ice evaporation calculations and over water for the over-water calculations. Empirical functions are used for surface heat fluxes. Water and ice pack heat balances (Croley and Assel, 1994) were used to relate ice cover extent to meteorology, heat storage, and surface fluxes between the atmosphere, the water body, and the ice pack. The effects of past additions or losses are superimposed to determine the surface temperature on any day as a function of heat in storage; each past addition or loss is parameterized by its age. Heat in storage in the lake at the end of each day is given by a simple conservation of energy. The lake thermodynamic model was applied to the seven Great Lakes water bodies, illuminating representative behavior. 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. Normal ice cover reaches a maximum in February of 75% (Superior), 45% (Michigan), 68% (Huron and Georgian Bay), 90% (Erie), and 24% (Ontario) (Assel et al., 1983); the annual averages are, respectively, 4.6%, 2.6%, 3.4%, 14.3%, and 0.9%.

The large basin and lake storages of water and ice and the large lake and ice storages of energy represent an 'intrinsic memory', that is captured in GLERL's models, which allows forecasting of basin moisture storage and runoff (basin storage buffering) in the face of uncertain meteorology. These models also allow prediction of evaporation (heat storage buffering) and lake levels (lake storage buffering) of up to about six months of low-frequency changes. They further enable estimation of ice formation and all related hydrological variables.

The magnitude of the hydrological variables vary with season; monthly precipitation is fairly uniform throughout the year while runoff peaks during the spring, resulting primarily from snow melt. Runoff is minimum in late summer and early fall due to large evapotranspiration from the land basin. Lake evaporation reaches a minimum during spring and gradually increases to 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. Over lake precipitation, basin runoff, and lake evaporation are all of the same order of magnitude for each lake. Annual precipitation ranges from 82 cm for Superior to 93 cm for Ontario; annual runoff to the lake ranges from 62 cm for Superior to 169 cm for Ontario; and annual lake evaporation ranges from 56 cm for Superior to 90 cm for Erie.

3. Water Impact Assessment

The Great Lakes Environmental Research Laboratory integrated their models into a system to estimate net basin supplies to the lakes, lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1993, 1995; Croley and Hartmann, 1987, 1989; Croley and Lee, 1993; Hartmann, 1990). They partially assessed their models by computing net basin supplies to the lakes (basin runoff plus over lake precipitation minus over lake evaporation) with historical meteorological data for 1951–1980 and comparing to historical net basin supplies, derived as a residual from lake water balances. The absolute average annual difference ranged from 1.6% to 2.7% in 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 snow melt and evapotranspiration model discrepancies for the other lake basins. Also, the runoff model does not have explicit vegetation components; it only reflects vegetation effects through calibration to observed runoff. While monthly differences were generally small, a few were significant. The low annual residuals were felt to be acceptable for model use 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 by comparing to historical net basin supplies is difficult since water budgets incorporate all budget term errors in the derived net basin supplies.

The system of hydrological models was first applied to the (untransposed) historical daily meteorological time series of air temperature, precipitation, wind speed, humidity, and cloud cover within the Great Lakes basin. The 43 years of available data from 1948–1990 were used as follows. The data from 1948–1950 were used in model simulations with arbitrary initial conditions (for soil moisture, snow pack, groundwater storage, and lake heat storage and surface temperature) to yield preliminary conditions initial to 1951. Then the 40 years between 1951 and 1990 were simulated by using 1 January 1951 modeled values as initial conditions. The simulation was repeated with end conditions used as initial conditions until there was no change (to arrive at a 'steady-state' condition); this became the estimates of the 'base case' hydrology. (This required only one iteration for all watersheds and lakes.) Later data, available at the time of the study from 1991–1992, were not used as they were of poorer quality since fewer stations were then available. This same procedure was then used to estimate the Great Lakes hydrology of each transposed climate by directly applying the system of hydrological models to all transposed climate 40-year daily meteorological time series. The impacts were estimated by comparing the outputs for each transposed climate to the base case. These are climates 1 through 4 and climate 5, which is climate 3 corrected for lake effects, as detailed in the companion paper by Kunkel et al. (1998). Climate 1 is the climate of 6° S and 10° W of the Great Lakes basin; climate 2 is 6° S × 0° W;

3 is 10° S × 11° W; 4 is 10° S × 5° W; and 5 is climate 3 corrected for lake effects. Correction of the other climates for lake effects was discontinued when we found that the changes in annual hydrology resulting from the consideration of lake effects was inconsequential.

4. Great Lakes Climate Change Hydrological Responses

The results of the model runs are characterized by comparing the mean annual and seasonal values of each hydrological variable, for each of the five climates tested, with each other and with 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. The sections that follow give annual average and seasonal estimates for both means and standard deviations for selected variables. In the interests of clarity, seasonal steady-state behavior is exemplified here in figures for only the Lake Superior basin and climate 2, and summarized for the entire period in annual tables only for Lakes Superior, Michigan, Erie, and Ontario and only climates 2 and 3. In addition, some annual tables also include Lake Huron (without Georgian Bay) and climate 5 to illustrate the most pronounced impacts of considering lake effect corrections to one of the transposed climates. (Lake Huron without Georgian Bay was treated separately from Georgian Bay for modeling of runoff and water body thermodynamics.) These examples are sufficient to illustrate the primary observations. The entire daily time series and the average annual cycles (daily values) for the means and standard deviations for each of these (and other) variables are available separately (Croley et al., 1996).

4.1. BASIN HYDROLOGY

The increased air temperatures, consequent in all transposed climates (Kunkel et al., 1998), significantly alter the heat balance of the surface hydrology. As seen in Table II, the snow pack is almost completely eliminated as the relative change varies among climates and lake basins from a 86% to a 99% drop in accumulated snow moisture. Furthermore, evapotranspiration increased significantly (19% to 88% for the lakes shown in Table II). For the eastern basins (represented by Erie and Ontario in Table II), the greatest increases in evapotranspiration and the greatest decreases in snow pack correspond to the southern-most transposed climates (3, 4, and 5), which have the greatest increases in average annual air temperatures. For the western basins (Superior and Michigan), the greatest increases in evapotranspiration come from the eastern-most climates (2 and 4), which have the greatest increases in precipitation for those basins. This implies that the western basins, under the current climate, have moisture-limiting evapotranspiration where the effects of relatively less water availability (as reflected by precipitation) limit evapotranspiration more than in the eastern basins. This is due primarily to the hydrology of the soils and the

Table II

Average annual basin hydrology means and relative changes[a]

Basin	Snow water equivalent (mm)			Total moisture storage (mm)			Overland evapotranspiration (mm)			Runoff as an overland depth (mm)		
	BASE	#2	#3	BASE	#2	#3	BASE	#2	#3	BASE	#2	#3
Superior	50.6	–86%	–98%	297	–41%	–71%	423	37%	19%	394	–27%	–62%
Michigan	11.7	–87%	–98%	115	–3%	–58%	506	48%	22%	322	23%	–34%
Erie	5.7	–88%	–98%	24	–20%	–40%	569	41%	49%	344	48%	17%
Ontario	15.7	–96%	–99%	61	–42%	–36%	473	48%	88%	461	–14%	9%

[a] Climate #2 is 6° S × 0° W and climate #3 is 10° S × 11° W.

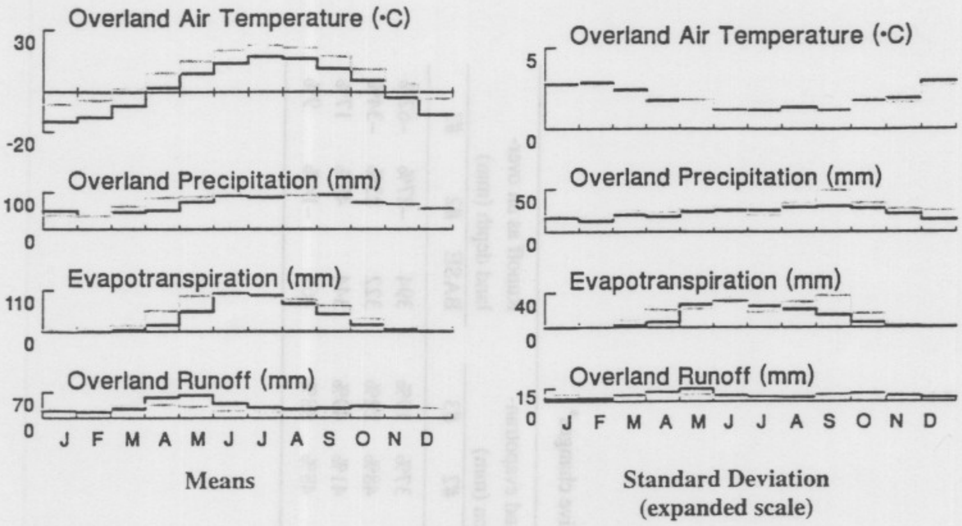

Figure 1. Seasonal Lake Superior basin hydrology (black is base case and gray is climate 2, 6° S × 0° W).

watersheds in those basins. The increased evapotranspiration and decreased snow pack result in a general reduction of total moisture (snow water, soil moisture, groundwater, and surface storage) that is greatest for the western-most climates.

By itself, the net effect of the increased air temperatures, through increased evapotranspiration and decreased moisture storage in the basins, would be decreased runoff. While Table II does indeed show decreased runoff in some cases, others show runoff increases. In these cases, increased evaporative demand is offset by precipitation increases since precipitation is generally higher under all of the transposed climates (Kunkel et al., 1998). Again, the greatest runoff decreases are observed for the northern-most basins and the western-most transposed climates.

Figure 1 depicts seasonal behavior in mean evapotranspiration and runoff for Lake Superior under climate 2. For all lake basins for the western-most climates (1, 3, and 5), evapotranspiration has shifted earlier in the seasonal cycle. A much larger percentage of the annual evapotranspiration and the peak evapotranspiration occur earlier. This also appears true for the northernmost basins for climates 2 and 4; for the other basins in climates 2 and 4, while evapotranspiration increases, the seasonal pattern is not significantly changed. These shifts are due to the earlier loss of snow moisture and increased spring temperatures which allow direct evaporation of liquid soil water and transpiration from developing vegetation to begin earlier in the season. This affects the seasonal distribution of runoff as well. The shift to an earlier seasonal peak in mean runoff (Figure 1) is typical of the behavior on all basins for all climates. The shift in runoff is therefore far more consistent across all basins and all climates than is the shift in evapotranspiration. Again, this results

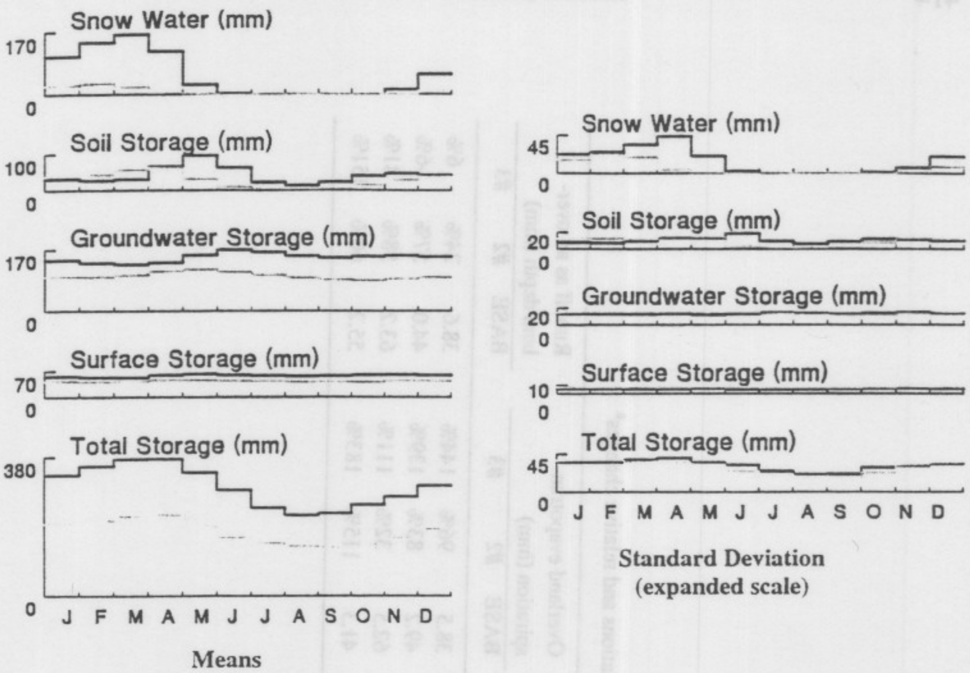

Figure 2. Seasonal Lake Superior basin moisture storage (black is base case and gray is climate 2, 6° S × 0° W).

from the loss of snow moisture; basins have more winter runoff than the base case, contributing to the runoff shift.

Table III and Figure 2 (standard deviations) 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 total moisture are generally less than 40% and of mixed sign, with respect to both annual and seasonal values. Evapotranspiration variability is considerably greater than the base case, particularly for the southernmost climates (Table III; Figure 1). This corresponds to the greatest and most variable precipitation; see Tables II and IV in Kunkel et al. (1998), 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 as much relative change in basin runoff variability as in evapotranspiration variability in Table III and Figure 1 (standard deviations), because of the unchanging variability of water storage in the basin; see Figure 2 (standard deviations). There does appear to be a slight increase in runoff variability during the winter for almost all basins and climates, as in Figure 1 for climate 2 standard deviations on the Superior basin, corresponding to the absence of the snow pack and consequent runoff during the wintertime. The

Table III

Average annual basin hydrology standard deviations and relative changes[a]

Basin	Snow water equivalent (mm)			Total moisture storage (mm)			Overland evapotranspiration (mm)			Runoff as an overland depth (mm)		
	BASE	#2	#3	BASE	#2	#3	BASE	#2	#3	BASE	#2	#3
Superior	11.0	–57%	–91%	22.3	15%	–6%	38.5	96%	144%	38.6	24%	6%
Michigan	5.4	–70%	–94%	13.8	11%	–36%	49.2	83%	139%	44.0	57%	–6%
Erie	3.7	–73%	–97%	5.0	–36%	–24%	62.5	32%	111%	63.2	58%	81%
Ontario	10.3	–93%	–97%	11.2	–35%	–12%	41.3	115%	183%	55.2	60%	151%

[a] Climate #2 is 6° S × 0° W and climate #3 is 10° S × 11° W.

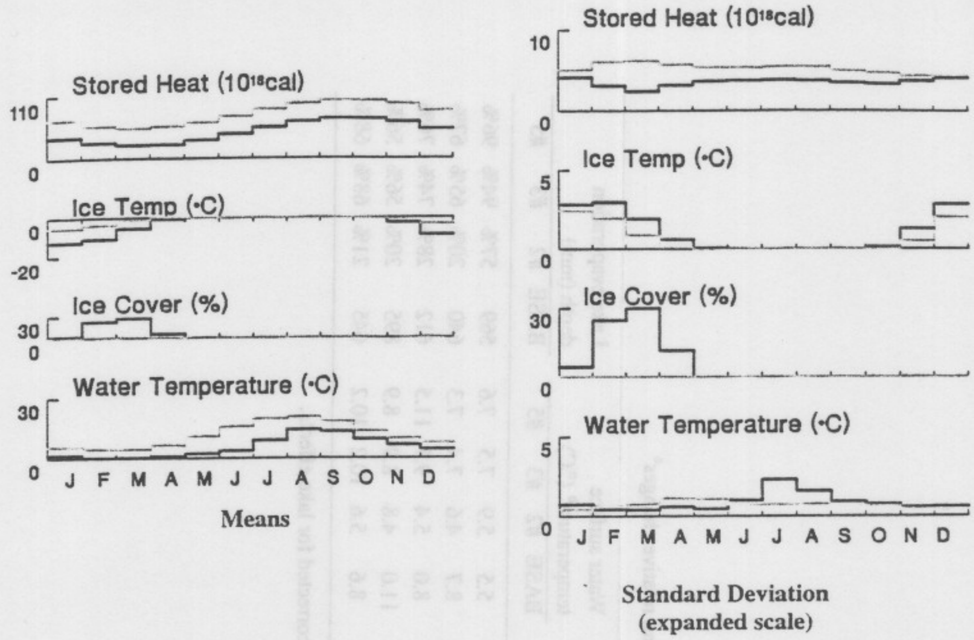

Figure 3. Seasonal Lake Superior lake heat storage characteristics (black is base case and gray is climate 2, 6° S × 0° W).

greatest consistent change in variability across all climates in both evapotranspiration and runoff, occur on Ontario, exposed to the most-eastern part of each climate with its most variable precipitation.

4.2. LAKE HEAT BALANCE

Insolation changes reflect largely the cloud cover changes reported by Kunkel et al. (1998, Table III); 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 climates. 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. Net long wave and sensible heat exchange increase, implying more heat stays in the lakes; the long-term average increases in both are more pronounced on all lakes for the southern-most climates (3, 4, and 5). These overall increases in heat storage in the lakes are balanced by increases in evaporation, which are rather large compared to the base case.

Average stored heat increases over the Great Lakes, depending on the transposed climate considered; Table IV shows increases between 36% to 180%. The largest relative heat increases are seen to occur for the southern-most climates (3, 4, and 5), but are substantial in all cases. The consideration of lake effects (compare climates 3 and 5) makes a difference, particularly on Lake Huron (Lake Huron,

Table IV
Average annual lake heat balance means and relative changes[a]

Basin	Stored heat (10^{17} cal)				Ice cover (%)				Water surface temperature[b] (°C)				Lake evaporation depth (mm)			
	BASE	#2	#3	#5	BASE	#2	#3	#5	BASE	#2	#3	#5	BASE	#2	#3	#5
Superior	47.1	66%	112%	118%	4.6	−100%	−100%	−100%	5.5	5.9	7.5	7.6	569	57%	94%	96%
Michigan	23.9	36%	98%	111%	2.6	−100%	−100%	−100%	8.7	4.6	7.4	7.3	640	20%	65%	67%
Huron[c]	15.9	65%	138%	180%	3.4	−100%	−100%	−100%	8.0	5.4	9.7	11.5	612	28%	74%	76%
Erie	4.9	41%	82%	82%	14.3	−99%	−100%	−100%	11.0	4.8	8.9	8.9	895	20%	56%	56%
Ontario	8.4	62%	157%	157%	0.9	−100%	−100%	−100%	8.6	5.6	10.2	10.2	645	21%	68%	68%

[a] Climate #2 is 6° S × 0° W, climate #3 is 10° S × 11° W, and climate #5 is #3 corrected for lake effects.
[b] Absolute difference instead of relative change.
[c] Not including Georgian Bay.

sans Georgian Bay, and climate 5 are included in Tables IV, V, and VII because of this). The stored heat appears as a constant amount higher throughout the seasonal cycle, since we are looking at steady-state conditions; see Figure 3 (means). The increased heat in storage means that ice cover is practically eliminated under all transposed climates on all lakes. The heat budget gives rise to increased water surface temperatures as seen in Figure 3 (means) and summarized in Table IV. The average increase in water surface temperatures are seen to rise from 4.6 °C on Lake Michigan (climate 2) to 11.5 °C on Lake Huron (climate 5) in Table IV. Again, the consideration of lake effects on climate 3 (climate 5) results in a much larger difference on Huron than on any other lake. The heat storage capacity of a lake influences the increase in water surface temperatures that can be seen in the Figure 3 means. 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 climates (3, 4, and 5) result in the greatest water surface temperature increases. The increased water temperatures are responsible for the increases in evaporation and sensible heat flux. Table IV shows increases in annual lake evaporation of 20% to 96%, depending on the lake and the climate. Again, the southern-most climates increase lake evaporation over the base case the most.

The variabilities associated with the lake heat balance variables are summarized in Table V and depicted for Lake Superior, climate 2, standard deviations in Figure 3. The stored heat exhibits some increase in variability for all climates, in Table V, for the deep lakes only, and notably different variability for Lake Huron's stored heat, water surface temperature, and lake evaporation between climates 3 and 5 (representing the impact of considering lake effects). The variability appears to be spread more uniformly across the seasonal cycle in every transposed climate 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 3 standard deviations are typical in this regard. Also, since the ice pack is not present anymore (see Figure 3 means), the variability associated with the ice pack is zero (ice pack stays at a constant zero value); see Figure 3 (standard deviations). This means relative change of 100% in the standard deviation of ice cover in Table V.

Water surface temperature is more constant (variability is reduced); see Table V. Figure 3 standard deviations show that for Lake Superior under climate 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 climates and other lakes; it reflects, again, the absence of the ice cover under the transposed climates. Seen in Figure 3 (standard deviations), and more pronounced on other lakes and under other climates, 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.

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

Table V

Average annual lake heat balance standard deviations and relative changes[a]

Basin	Stored heat (10^{17} cal)				Ice cover (%)				Water surface temperature[b] (°C)				Lake evaporation depth (mm)			
	BASE	#2	#3	#5	BASE	#2	#3	#5	BASE	#2	#3	#5	BASE	#2	#3	#5
Superior	3.10	61%	55%	58%	5.50	−100%	−100%	−100%	0.70	−29%	−43%	−43%	59.4	12%	20%	24%
Michigan	1.50	67%	47%	40%	3.40	−100%	−100%	−100%	0.80	0%	−50%	−50%	67.3	1%	8%	11%
Huron[b]	1.10	73%	27%	18%	5.9	−100%	−100%	−100%	0.60	0%	−17%	−33%	64.1	15%	7%	1%
Erie	0.30	0%	0%	0%	8.00	−93%	−100%	−100%	0.70	0%	−29%	−29%	78.5	−25%	−33%	−33%
Ontario	0.80	50%	25%	25%	1.80	−100%	−100%	−100%	1.00	−40%	−60%	−60%	60.2	15%	−3%	−3%

[a] Climate #2 is 6° S × 0° W, climate #3 is 10° S × 11° W, and climate #5 is #3 corrected for lake effects.
[b] Not including Georgian Bay.

in which a majority of annual evaporation occurs within a few days. By estimating the evaporation time series on each lake under each climate, it is possible to address to what extent evaporation is 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 through December as humidity drops and water temperatures (and hence vapor pressure deficits) rise. It then generally decreases as dropping water temperatures lower the vapor pressure deficit over water 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 synoptic systems (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 a tenth of the annual cycle (not necessarily continuously); see Table VI. 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 VI shows that it 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% to 88.9% of the annual cycle, depending on the lake, it occurs under the transposed climates in 92.3% to 98.2% of the annual cycle, depending on the lake and climate.

4.3. LAKE THERMAL STRUCTURE

The deep lakes (Superior, Michigan, Huron, Georgian Bay, and Ontario) show water surface temperatures that stay above 4 °C throughout the annual cycle in some years. Figure 4 illustrates this for 1961 for Lake Superior under both the base case climate and under transposed climate 2. This means that buoyancy-driven turnovers of the water column do not occur in the same way as they do at present. In some years, the deep lakes are changed from dimictic lakes (turnovers occur twice a year as water temperatures pass through the point of maximum density, 4 °C) to

Table VI
Minimum fractions of the year for occurrence of annual lake evaporation fractions,[a] %

Annual evap. (%)	Lake Superior Base	Lake Superior #2	Lake Superior #3	Lake Michigan Base	Lake Michigan #2	Lake Michigan #3	Lake Erie Base	Lake Erie #2	Lake Erie #3	Lake Ontario Base	Lake Ontario #2	Lake Ontario #3
10	2.4	2.7	2.7	2.2	2.3	2.5	2.1	2.6	2.7	2.2	2.6	2.8
20	5.4	6.2	6.2	5.2	5.5	5.9	4.9	6.2	6.6	5.1	6.0	6.5
33.3	10.2	11.8	11.9	9.9	10.6	11.8	9.3	12.2	13.1	9.8	11.5	12.8
50	17.6	20.1	20.6	17.2	18.6	21.2	16.4	21.7	23.2	16.9	20.2	22.6
66.7	26.7	30.5	31.8	26.5	29.1	33.8	25.9	33.7	35.7	26.0	31.4	35.4
80	36.0	41.7	44.5	36.4	40.5	47.3	36.7	46.0	48.0	35.7	43.5	48.8
90	45.7	54.0	59.5	47.2	53.2	61.1	48.8	58.3	60.2	46.3	56.4	62.1
100	79.8	94.4	98.2	82.9	92.3	96.0	88.9	96.1	94.2	84.2	96.9	96.8

[a] Climate #2 is 6° S × 0° W and climate #3 is 10° S × 11° W.

Table VII
Average characteristics of turnovers/reversals[a]

Basin	Fraction dimictic (%)				Monomictic reversal water temperature (°C)			
	Base	#2	#3	#5	Base	#2	#3	#5
Superior	100	24	0	0	–	4.5	6.1	6.3
Michigan	100	50	0	0	–	4.7	6.9	7.6
Huron[b]	100	2	0	0		5.7	9.5	12.4
Erie	100	63	0	0	–	5.5	7.9	7.9
Ontario	100	0	0	0	–	5.9	10.7	10.7

[a] Climate #2 is 6° S × 0° W, climate #3 is 10° S × 11° W, and climate #5 is #3 corrected for lake effects.
[b] Not including Georgian Bay.

monomictic lakes (maximum turnover occurs at the temperature 'reversal' where temperatures stop declining and start rising again and the minimum temperature is greater than 4 °C). Figure 4 shows that the base case temperature profile for Lake Superior passed through 4 °C in June 1961 and approached, in December, the January 1962 transition. Under transposed climate 2, temperatures remain above 4 °C but have a nearly vertical profile throughout most of February and March. This represents a change from dimictic to monomictic, and makes lake turnovers more dependent on factors other than temperature profiles, e.g., winds and water chemistry.

Table VII shows that the deep lakes remain dimictic under the transposed climates only part of the time or not at all, depending on the lake and the climate. The largest change is associated with Lake Ontario which is the furthest south of the deep lakes. The southern-most transposed climates (3, 4, and 5) show the largest shifts, with all deep lakes becoming 100% monomictic. Table VII 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. Table VII also shows notable differences in monomictic reversal temperatures for Lakes Michigan and Huron for climates 3 and 5 (reflecting the impact of considering lake effects).

The timing of maximum turnovers or temperature reversal shifts under the transposed climates. Table VIII 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 maximum turnover occurs even earlier in the year than the dimictic turnovers.

Temperature-depth profiles, as in Figure 4, for every day of a single model year can be combined and depicted as depth-time plots of temperature isolines; see Figure 5 for an example on Lake Superior for the base case and climate 2 for 1961.

Table VIII

Average dates and depths of maximum turnover or temperature reversal[a]

Basin	Dimictic dates						Dimictic depths					
	Base		Climate #2		Climate #3		Base		Climate #2		Climate #3	
	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall	Spring	Fall
Superior	27 Jun	24 Dec	10 Apr	22 Feb	–	–	238 m	156 m	99 m	275 m	–	–
Michigan	23 May	31 Dec	6 Apr	6 Feb	–	–	123 m	105 m	47 m	145 m	–	–
Erie	25 Apr	24 Dec	10 Mar	29 Jan	–	–	63 m	64 m[b]	29 m	60 m	–	–
Ontario	21 May	14 Jan	–	–	–	–	231 m	202 m	–	–	–	–

Basin	Monomictic dates			Monomictic depths		
	Base	#2	#3	Base	#2	#3
Superior	–	19 Mar	14 Mar	–	306 m	304 m
Michigan	–	4 Mar	1 Mar	–	195 m	226 m
Erie	–	17 Feb	9 Feb	–	64 m[b]	60 m
Ontario	–	11 Mar	5 Mar	–	232 m	228 m

[a] Climate #2 is 6° S × 0° W and climate #3 is 10° S × 11° W.
[b] Maximum average depth of the lake.

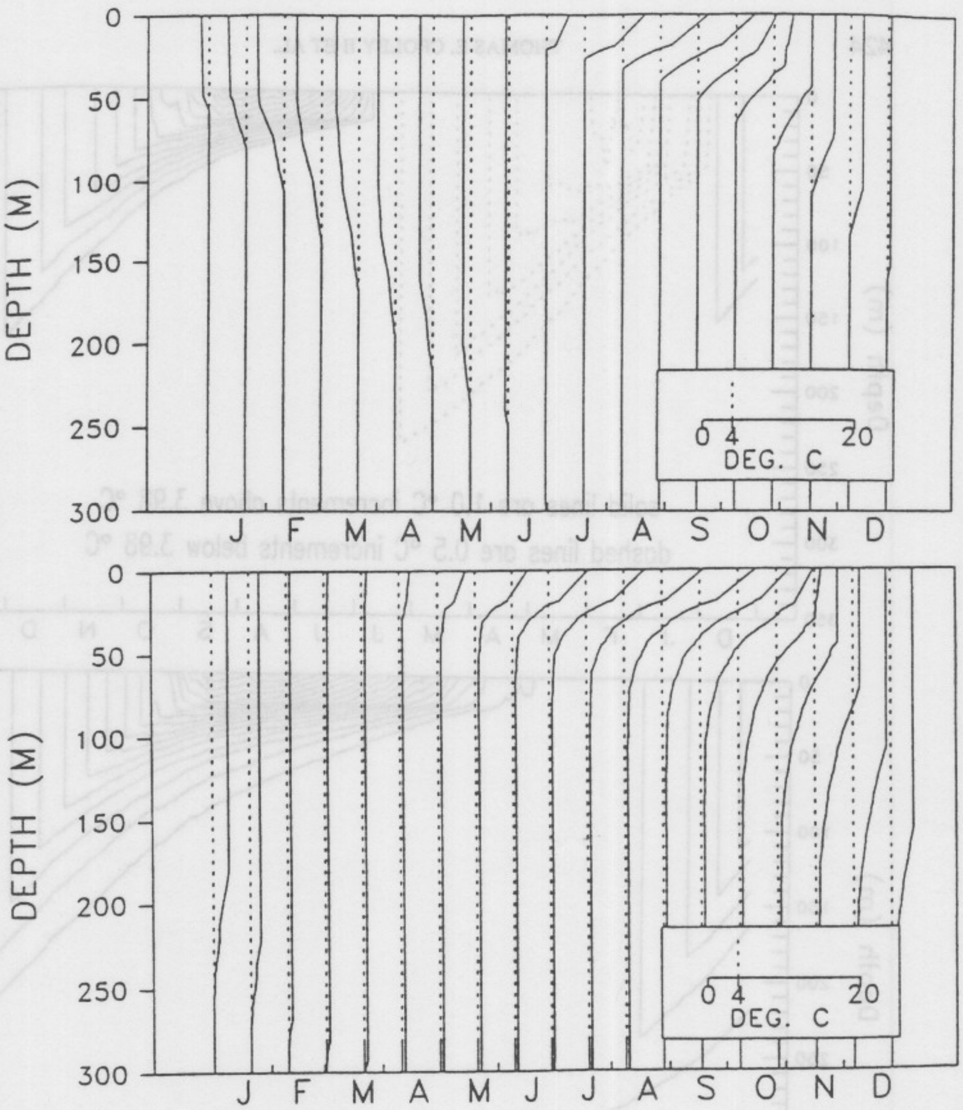

Figure 4. Lake Superior temperature-depth profiles every 20 days for 1961 (top is base case and bottom is climate 2, 6° S × 0° W).

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 VIII also summarizes the maximum depths at turnover in the lakes. Dimictic spring turnovers exhibit shallower average depths under the transposed climates 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

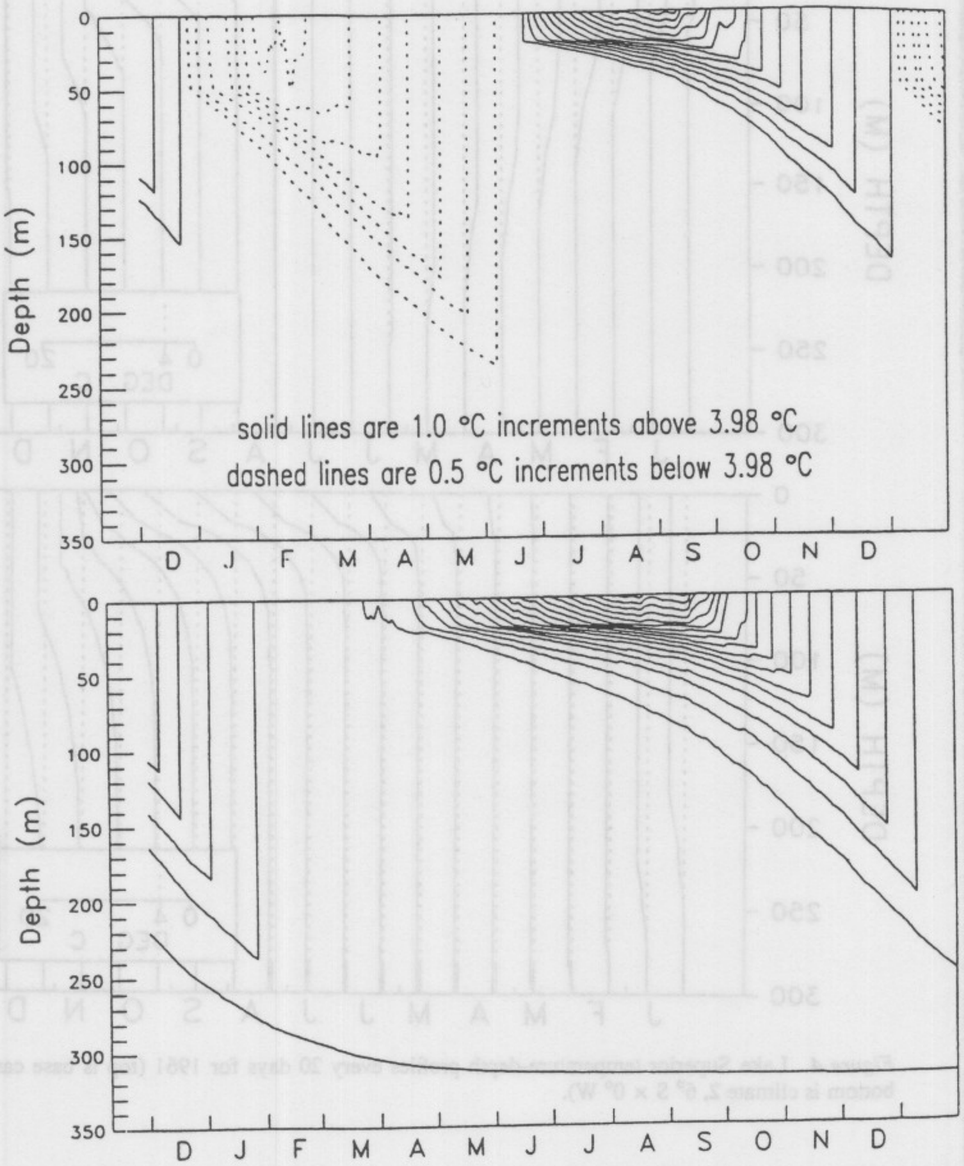

Figure 5. Lake Superior depth-time temperature isolines for 1961 (top is base case and bottom is climate 2, 6° S × 0° W).

Ontario. On Lake 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.

There is a normal hysteresis observed in graphs of lake heat plotted with surface temperature, such as in Figure 6 for Lake Superior under the base case for 1961.

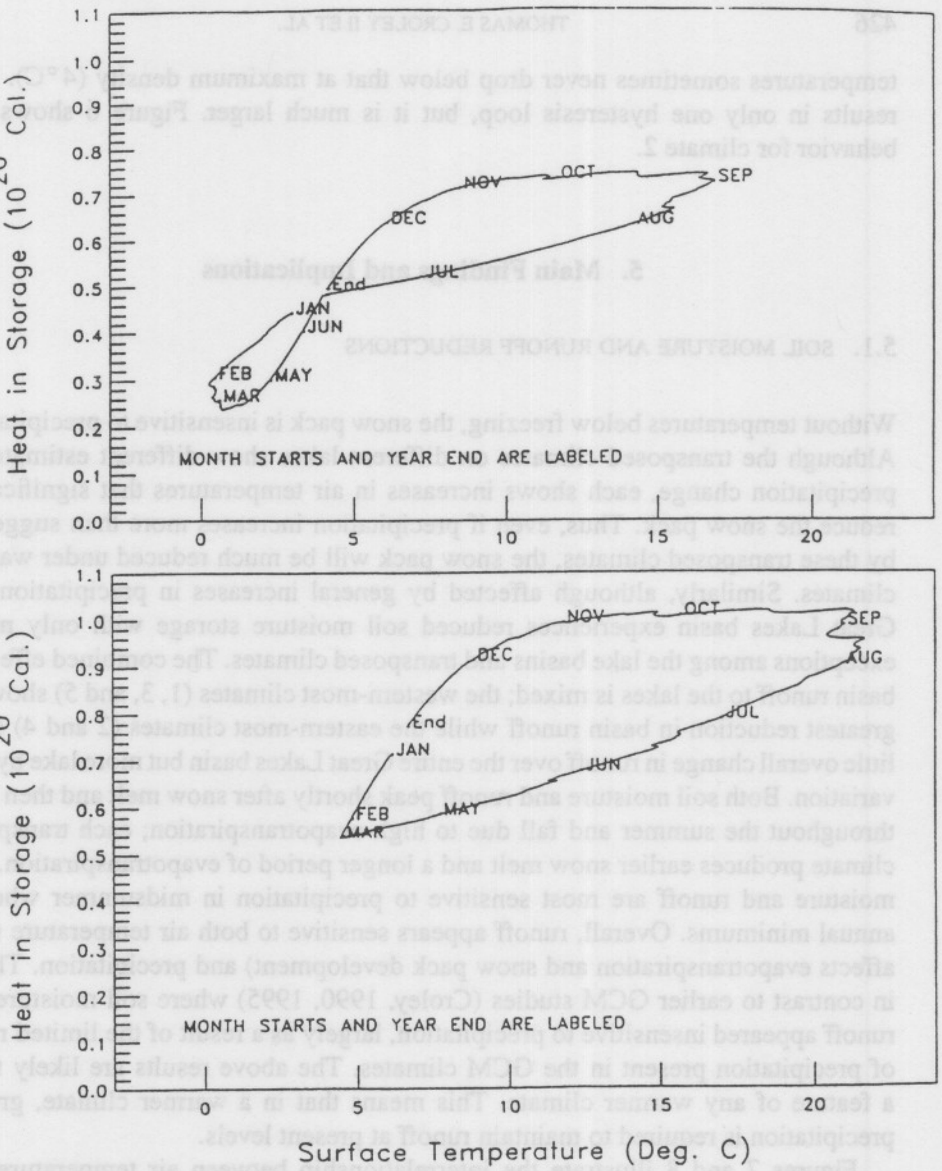

Figure 6. Lake Superior heat-temperature hysteresis for 1961 (top is base case and bottom is climate 2, 6° S × 0° W).

This reflects the mixing of heat at depth. Surface temperatures rise quickly and heat storage follows after the spring turnover. When surface temperatures begin to drop in the fall, stored heat does not initially also fall, but eventually drops more slowly. After the fall turnover, heat storage drops more rapidly than temperatures, resulting in the characteristic double 'loop' in the plot. Under the warmer climate changes, temperatures sometimes never drop below that at maximum density (4 °C). This results in only one hysteresis loop, but it is much larger. Figure 6 shows this behavior for climate 2.

5. Main Findings and Implications

5.1. SOIL MOISTURE AND RUNOFF REDUCTIONS

Without temperatures below freezing, the snow pack is insensitive to precipitation. Although the transposed climates on different lakes show different estimates of precipitation change, each shows increases in air temperatures that significantly reduce the snow pack. Thus, even if precipitation increases more than suggested by these transposed climates, the snow pack will be much reduced under warmer climates. Similarly, although affected by general increases in precipitation, the Great Lakes basin experiences reduced soil moisture storage with only minor exceptions among the lake basins and transposed climates. The combined effect on basin runoff to the lakes is mixed; the western-most climates (1, 3, and 5) show the greatest reduction in basin runoff while the eastern-most climates (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 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 it affects evapotranspiration and snow pack development) and precipitation. This is in contrast to earlier GCM studies (Croley, 1990, 1995) where soil moisture and runoff appeared insensitive to precipitation, largely as a result of the limited range of precipitation present in the GCM climates. 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 7 and 8 illustrate the interrelationship between air temperature and runoff-precipitation ratios and lake evaporation changes for the Lake Superior basin, the basin subjected to the widest range of scenarios in this study (i.e., the largest temperature increases and the most variable precipitation changes). The figures also include data from the previously mentioned (Kunkel et al., 1998) EPA and IJC studies. The relationship among these variables is linear. The response of annual runoff to annual temperature and precipitation changes is similar among the different lakes and their interrelationship can be approximated by the following linear equation:

$$\Delta(RO) = a\Delta P + b\Delta T,$$

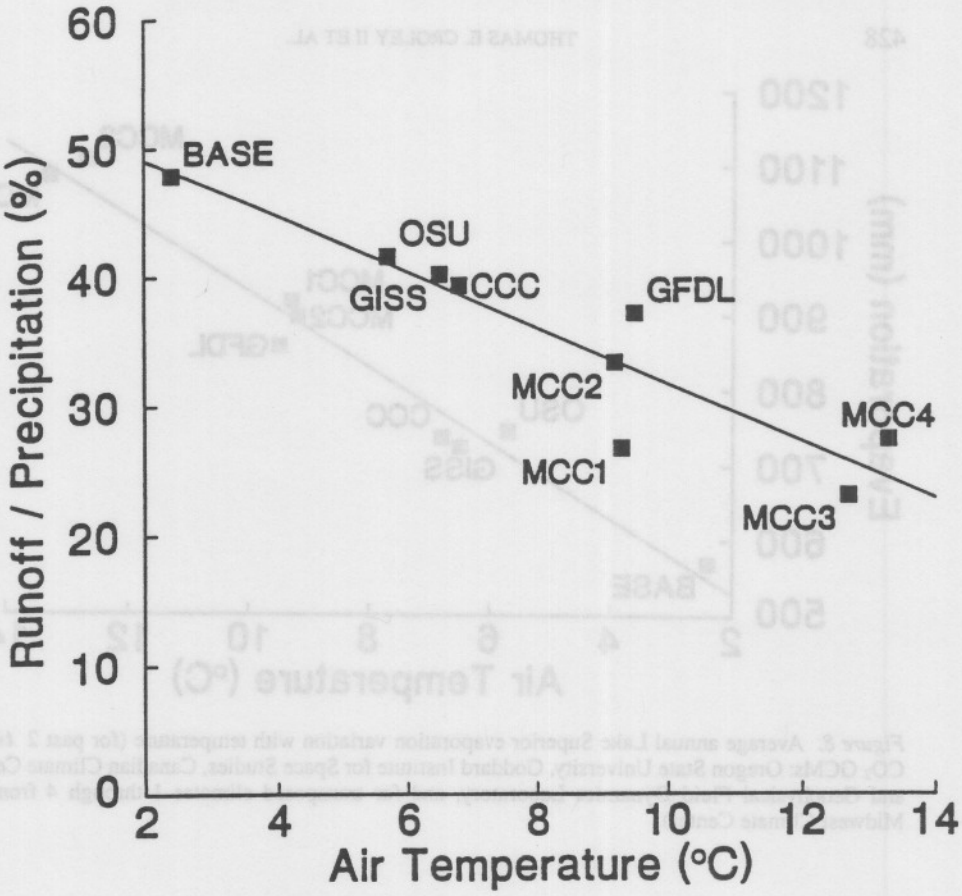

Figure 7. Average annual Lake Superior basin effective-runoff variation with temperature (for past $2 \times CO_2$ GCMs: Oregon State University, Goddard Institute for Space Studies, Canadian Climate Centre, and Geophysical Fluid Dynamics Laboratory, and for transposed climates 1 through 4 from the Midwest Climate Center).

where $\Delta =$ change in the variable compared to the base climate, $RO =$ runoff, $P =$ precipitation, $T =$ temperature, and a and b are coefficients to be determined for each lake. Rearranging,

$$\frac{\Delta(RO)}{\Delta T} = a\frac{\Delta P}{\Delta T} + b,$$

enables calculation of the coefficients through a linear fit for each lake; see Figure 9. This graph summarizes model results regarding runoff response to climate change. The y-intercept represents the sensitivity of runoff to temperature changes alone (no precipitation change), generally in the range of –4%/°C to –8%/°C. The slope of the line gives the sensitivity of runoff to precipitation changes. The x-intercept represents the value of $\Delta P/\Delta T$ for which runoff is maintained at its current value

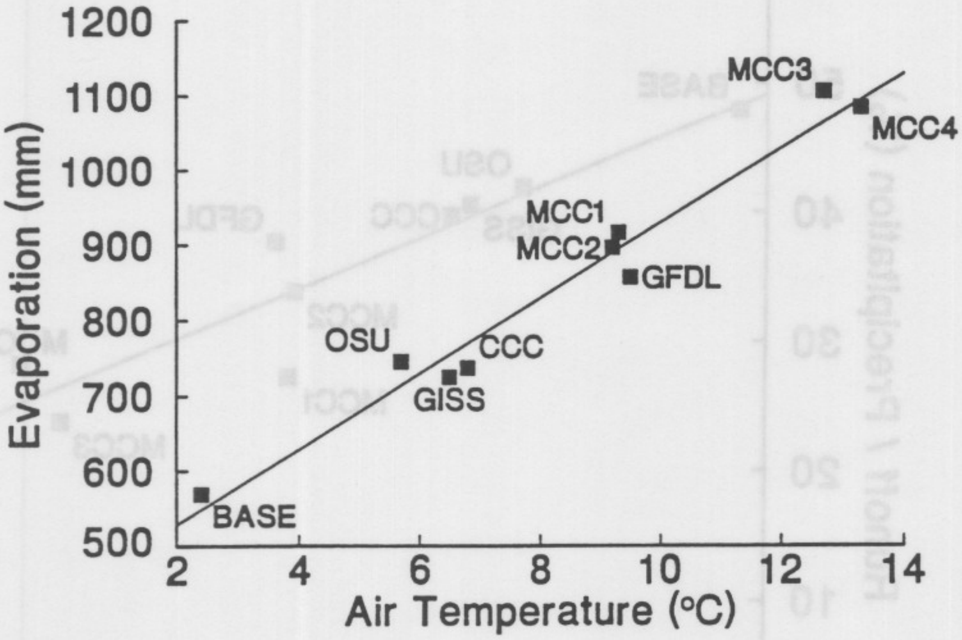

Figure 8. Average annual Lake Superior evaporation variation with temperature (for past 2 *times* CO_2 GCMs: Oregon State University, Goddard Institute for Space Studies, Canadian Climate Centre, and Geophysical Fluid Dynamics Laboratory, and for transposed climates 1 through 4 from the Midwest Climate Center).

(unchanged, generally in the range of +3%/°C to +5%/°C). Thus, for every degree C increase in temperature, a 3% to 5% increase in precipitation is required to maintain runoff at its present climate value.

Figure 10 depicts annual runoff variability as the variability ratio (standard deviation of runoff divided by standard deviation of precipitation) per °C of temperature change. The results for runoff variability (Figure 10) are less well behaved and coherent than for runoff itself (Figure 9). For example, the temperature sensitivity (y-intercept) varies over a rather wide range of –2%/°C to –8%/°C, indicating that for no precipitation change, an annual temperature rise drops the annual runoff variability. In all cases, the slopes of the curves are positive indicating that the annual runoff variability, relative to annual precipitation variability, increases with precipitation.

Summarizing, climatic temperature rises lead to reductions in both runoff and runoff variability. Precipitation increases of 3% to 5% per °C temperature rise would be required to offset runoff reductions caused by temperature changes. It appears unlikely, under all warming scenarios examined to date for the Great Lakes, that precipitation increases will offset temperature effects. This is indicated even though precipitation increases are greater with the transposed climates considered here than with the earlier GCM-generated climates over the Great Lakes. While less

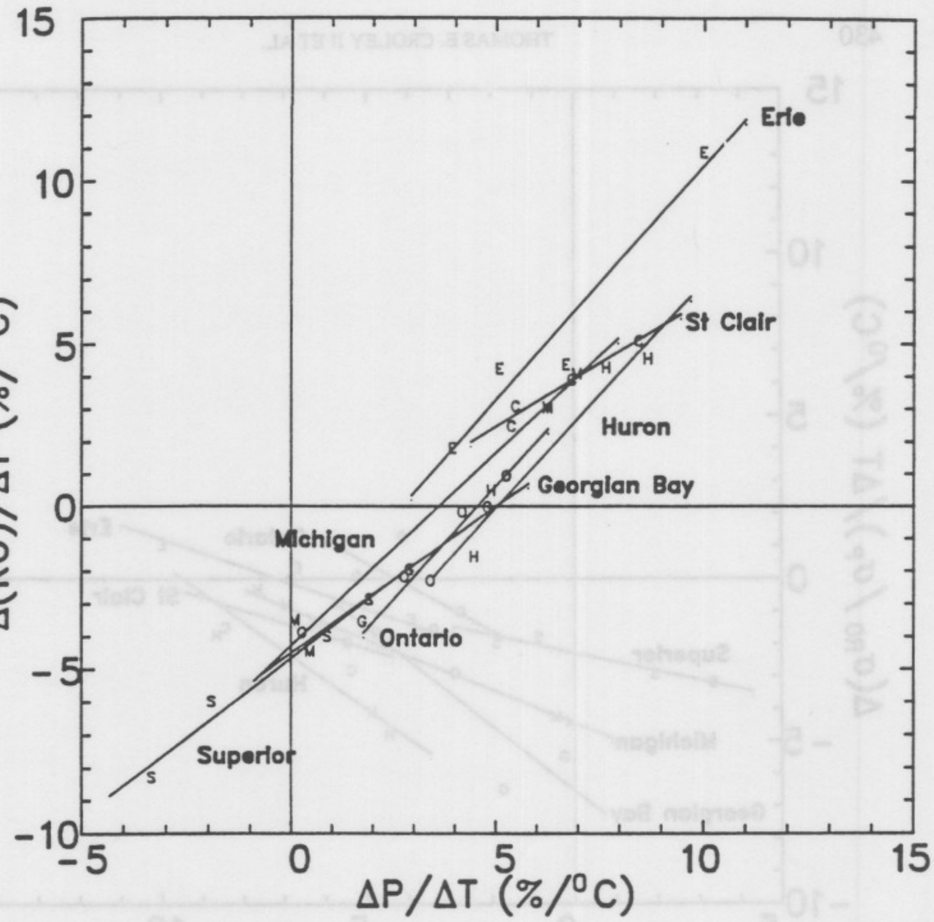

Figure 9. Average annual Great Lake basins unit runoff change variation with unit precipitation change.

pronounced, it appears that runoff variability, relative to precipitation variability, increases with precipitation. This higher variability of reduced runoff suggests low flow conditions will occur more frequently. Therefore, under all Great Lakes warming scenarios (both transposed climate and earlier GCM studies), runoff drops with enhanced potential for water management problems.

5.2. LAKE EVAPORATION INCREASES

All climates produce significant increases in lake evaporation. Interestingly, this occurs in the face of increased humidities on all lakes under all climates and decreased wind speeds on most lakes and climates, which would ordinarily reduce the evaporation. However, there is such a large increase in the heat that is input

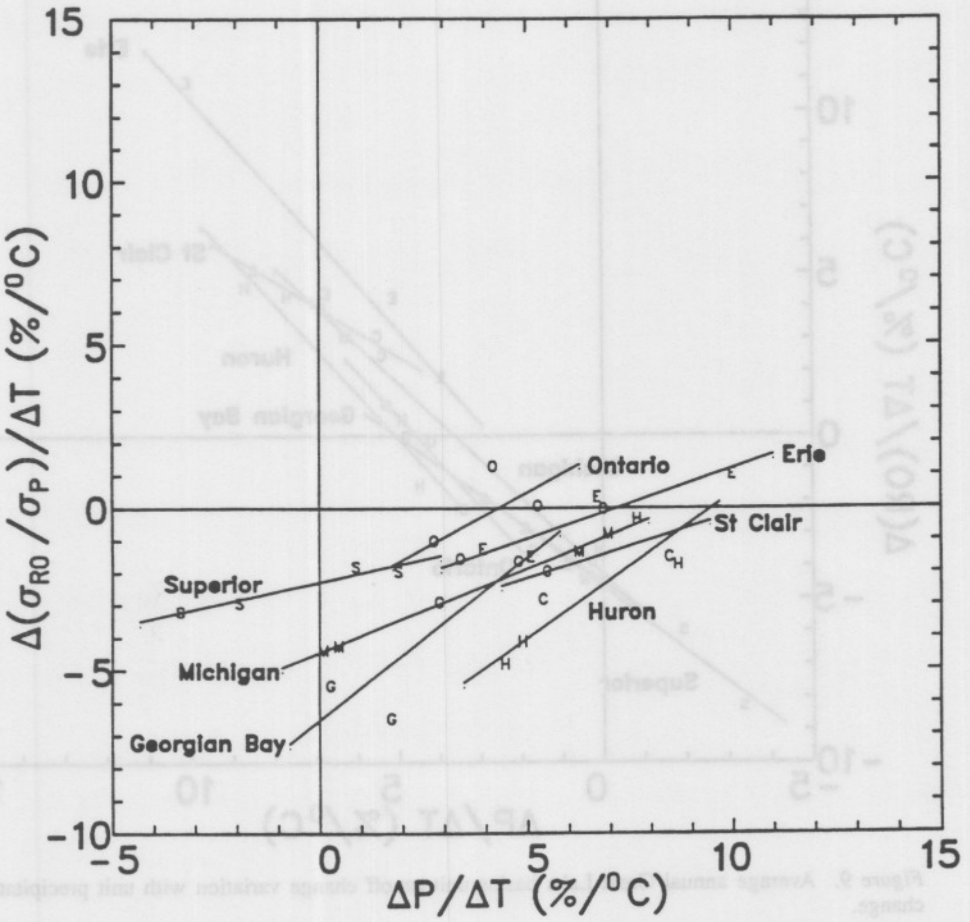

Figure 10. Average annual Great Lake basins unit runoff variability/precipitation variability ratio change variation with unit precipitation change.

and stored in the lakes that water surface temperatures are much higher and peak earlier under the transposed climates than under the base case. This increases the vapor pressure deficit between the water and overlying atmosphere. From an energy standpoint, the energy for increased evaporation is derived from four sources. Most important is the decrease in cloud cover which 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 non-linear relationship between saturation vapor pressure and temperature, the ratio of sensible to latent heat flux (Bowen ratio) decreases in all climates 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, thereby 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. There will thus be a considerable positive pressure on lake evaporation. Most GCM predictions do result in significant increases in lake evaporation (Croley, 1990, 1994, 1995). However, the changes in cloud cover in these climates may not be realized in a future warmer climate over the Great Lakes. The uncertainty associated with cloud cover changes over the Great Lakes and the sensitivity of evaporation to cloud cover suggest that lake evaporation increases may be smaller under a warmer climate than predicted here. This is a very real limitation of this climate transposition methodology.

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 climates for all lakes, the relative contribution of these events to total lake evaporation decreases. However, these events are still important. This indicates that accurate future estimates of lake evaporation will require accurate estimates of the number and severity of cold air outbreaks.

5.3. REDUCED TURNOVER FREQUENCY

Warmer climates can result in reduced frequency of buoyancy-driven water column turnovers. In some of these climates, lake surface water temperatures often do not fall to 4 °C (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/year (dimictic) to once per year (monomictic). 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 cycling, oxygenation of lake water at depth, and metals retention in lake sediments.

5.4. LAKE EFFECTS

Lake effects on regional climate appeared to have negligible hydrological effects on an annual basis for almost all of the Great Lakes. We used existing spatial and quantitative measures of lake effects on various climate conditions to modify data for climate 3. We tested lake effects on basin hydrology by calculating outcomes with and without lake effects present. The differences in basin hydrology for all Great Lake basins were negligible but the differences in lake heat storage and evaporation for Lake Huron were significant. We did not attempt a modeling investigation to ascertain how much lake effects might change in warm-wet climates like 2 or 4, but the lack of differences for climate 3 suggests that huge changes in lake effects would be required to significantly alter the annual basin-wide hydrological results. This is also true for annual lake heat storage and derived quantities in general, but not for Lake Huron. There it appears that lake effects may be significant.

However, on a seasonal basis, lake effects may be important. While the tables show relatively little effect on an annual basis (except.for Lake Huron thermodynamics), Kunkel et al. (1998) mention how seasonal effects were larger but opposite in sign and thus cancel on an annual basis. There are some lake effects worth explaining further where seasonal processes and responses dominate. As evidenced by Lake Huron thermodynamics, heat storage in the lakes may be particularly sensitive to seasonal lake effects, especially since heat storage is so sensitive to cloud cover changes.

5.5. HYDROLOGICAL SENSITIVITIES

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 $1 \times CO_2$ and $2 \times CO_2$ atmospheres. Those studies simply changed the magnitude of meteorological time series without affecting their temporal or spatial structures. Therefore, changes in variabilities 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. The responses of the system to changes in annual variability also are now clear whereas in the GCM studies they were not.

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