The general circulation of Lake Superior: mean, variability, and trends from 1979-2006

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Abstract

Previous observations and modeling studies of Lake Superior have only partly elucidated its large-scale circulation, in terms of both the climatological state and interannual variability. We use an eddy resolving, three-dimensional hydrodynamic model to bridge this gap. We simulate Lake Superior circulation and thermal structure from 1979 to 2006 and consider the mechanisms responsible for the flow. Model results are compared to available direct observations of temperature and currents. Circulation in the lake is primarily cyclonic during all seasons, and a two gyre structure is sometimes present. Surface circulation patterns in winter mimic wind directions, but become organized in summer by the presence of thermal gradients. On the annual mean, nearshore currents are controlled by thermal gradients, while offshore flow is primarily determined by the wind. From a uniform bathymetry simulation, we determine that topographic variations cause small-scale structures in the open lake flow and are critical to the development of nearshore-offshore temperature gradients. The lake exhibits significant variability in current speed and direction on synoptic timescales, but coherent patterns of interannual variability are not found. Long-term trends due to changing meteorological forcing are found. Model results suggest the increase in lake surface temperature (0.37°C/decade) is significantly correlated to increases in wind speed above the lake (0.18 m/s/decade), increased current speeds (0.37 cm/s/decade), and declining ice coverage (-886 km²/yr).
Introduction

Understanding of the mean and variability of the large-scale circulation in Lake Superior is of critical importance to environmental problems such as contaminant and invasive species transport, water quality, and ecosystem analysis and prediction. Mixing, circulation, and temperature determine the availability of light and nutrients for primary productivity, and changes in the thermal structure of the lake should alter locations and timing of phytoplankton blooms. Observations of surface lake partial pressure of carbon dioxide ($p_{\text{CO}_2}$) also suggest the carbon cycle cannot be understood without accounting for spatial heterogeneity in the Lake Superior’s physical system [Atilla et al., submitted 2010; Urban et al., 2005]. Lakes are subject to the atmospheric conditions and anthropogenic changes of their larger watersheds, and lake ecosystems may respond more rapidly to changes in the climate system than their terrestrial counterparts [Williamson et al., 2009]. Lake Superior is no exception. Its summer heat content is increasing twice as rapidly as regional atmospheric temperatures [Austin and Colman, 2007]. Wind speeds above the lake are also increasing and are expected to continue to change in warming conditions [Desai et al., 2009].

Lake Superior is the largest lake in the world by surface area and its volume is large enough to hold all of the other Great Lakes combined. The deepest part of the lake exceeds 400 m [Schwab and Seller, 1996], its mean depth is ~150 m, and its water residence time is 178 yrs [Quinn, 1992]. The lake turns over twice yearly, and the entire water column must warm (or cool in winter) to 3.98°C before stratification, because
freshwater is densest at 3.98°C. Due to its northern location and great depth, Lake Superior is the coldest of the Great Lakes and often does not stratify until June near shore and later in the open waters. The lake is completely ice covered about once every ten years [Assel, 2003, 2005], but each year ice forms throughout the near shore zone. The presence of near shore ice and otherwise harsh winters have generally precluded wintertime observations of temperature and currents.

There have been a few observations and modeling studies of the mean circulation and interannual variability of circulation in Lake Superior, but overall the system is poorly understood. Published studies on its circulation are based on data that is restricted in space and/or time. The first map of Lake Superior’s summer circulation was put together by Harrington [1895] from a bottle drift experiment. He released bottles within the lake and inferred a curved path between the bottle’s starting point and where it washed ashore. Harrington’s [1895] results exhibited a cyclonic flow around the basin during summer with many bottle paths following the bathymetry. The Federal Water Pollution Control Administration (FWPCA) collected current meter data at point locations throughout the lake during 1966 and 1967. These observations are the most recent lake-wide current observations for Lake Superior. Sloss and Saylor [1976] analyzed this data and organized the results into monthly maps of currents at the sixteen summer and nine winter locations. Beletsky et al. [1999] summarized the analysis of Sloss and Saylor [1976] into two maps, one of mean integrated summer circulation in the lake and one of the mean integrated winter circulation. Neither map is intended to represent a long-term mean. Large areas of these maps are empty, particularly in winter, because no current observations exist.
Intensive current observations were also taken during the Keweenaw Interdisciplinary Transect Experiments (KITES). Spring through fall currents and temperatures were measured along the Keweenaw Peninsula during the late 1990s using Acoustic Doppler Current Profilers (ADCP). Current measurements from the KITES project have high temporal and depth resolution but are limited in space to three locations within 21 km (8 km for one transect) of the southern shore.

Given the limited long-term, lake-wide observations, numerical models are valuable tools to fill in the gaps of both time and space. Hydrodynamic modeling of Lake Superior began in the 1970s when Lam [1978] simulated the lake-wide currents from June through September of 1973 using a 10 km horizontal grid resolution, prescribed model temperatures, and meteorological winds as forcing. The integrated large-scale flow of Lam [1978] largely agrees with the summer map presented in Beletsky et al. [1999], suggesting that integrated summer flow during 1973 was similar to the summer circulation of 1966 and 1967. Zhu et al. [2001] and Chen et al. [2001] developed a hydrodynamic model to study the development of the currents along the Keweenaw Peninsula and the effects of heat fluxes on the jet intensification. Their model was lake-wide, but the simulations were for summer periods and their analyses focused on currents along the Keweenaw Peninsula. NOAA’s Great Lakes Coastal Forecasting System (GLCFS) [Schwab and Bedford, 1999] simulates the daily circulation and temperature for each of the Great Lakes, Lake Superior at 10km resolution. The GLCFS is an excellent tool for water quality predictions. Plots of lake surface temperatures, surface currents, temperature transects, temperature profiles, and water levels for the current date are
available online at: http://www.glerl.noaa.gov/res/glcfs/. It is unknown by these authors if
the many years of hydrodynamic modeling done at GLCFS have been done with a single,
internally consistent model. The accumulated years of simulated currents have not been
analyzed in the literature. Although hydrodynamic models have been utilized for Lake
Superior, studies have not considered the climatological large-scale circulation of the lake
or its long-term variability.

In the other Great Lakes, there have been numerical studies that have focused primarily
on describing the mean circulation. Lake Michigan’s large-scale circulation is best
studied. Beletsky and Schwab [2008] simulated ten years of lake hydrodynamics in Lake
Michigan during 1998-2007 and found a stable cyclonic depth-integrated circulation
during both winter and summer. Weak anticyclonic circulations exist in the far northern
and southern regions of the lake. Schwab and Beletsky [2003] analyzed the generation of
vorticity in Lake Michigan and determined that wind stress curl during winter and heat
fluxes during summer are primarily responsible for the pattern of large-scale circulation
in the lake. Bathymetry also has an impact on current patterns, but its effect is not as
important as wind curl or heat fluxes. Recently, models have been developed for other
Great Lakes. Sheng and Rao [2006] simulated the circulation of Lake Huron for 1974-
1975 using a high resolution, nested grid hydrodynamic model and presented monthly
mean circulation and thermal structure of the lake during that time. Schwab et al. [2009]
present simulated summer currents in Lake Erie for 1994, and Prakash et al. [2007] used
a hydrodynamic model to simulate mean circulation patterns and pollutant transport in
Lake Ontario.
In this paper, we simulate the thermal structure and circulation of Lake Superior from 1979 to 2006. This is the longest single Great Lake simulation to date. Modeled temperatures and currents are compared to available observational data. Seasonal maps of the climatological circulation are created for surface, integrated, and below 50 m depths. Variability in integrated currents at nine open lake stations is examined and summarized in current rose plots. We address the following questions. What are the climatological patterns of the lake’s circulation? How much does the current direction change from year to year? Do winds and heat fluxes control Lake Superior’s circulation, as in Lake Michigan, or does the bathymetry have a greater control on flow? How has the lake circulation and thermal structure changed over the last three decades?

The paper is organized as follows. Section 2 includes a description of the hydrodynamic model and a comparison of model results to observations. The climatology of Lake Superior’s circulation is presented in Section 3.1, and mechanisms of the lake circulation are examined in Section 3.2. We discuss lake-wide trends in Section 3.3. Discussion and conclusions are presented in Section 4.

Section 2: Model

Section 2.1: Physical Model

Lake Superior’s large volume and long residence time (178 years) [Quinn, 1992] make it an appropriate water body to study with a fixed volume numerical model. We use the MIT general ocean circulation model [Marshall et al., 1997a, 1997b] configured to the
bathymetry of Lake Superior [Schwab and Seller, 1996]. The model resolves eddies and has a uniform horizontal resolution of 2 km x 2 km. The model uses a z-coordinate system of 28 vertical layers. The top 50 meters have finest vertical resolution, with layer thicknesses of 5 meters. Vertical resolution gradually becomes coarser with depth to a thickness of 33.7 meters at 322 meters depth. The model setup uses the hydrostatic approximation. The Smagorinsky [1963] horizontal diffusivity scheme and the K Profile Parameterization (KPP) vertical mixing scheme [Large et al., 1994] simulate the effects of sub-grid scale processes.

A bulk formula atmospheric module is used to calculate momentum exchange and heat fluxes between the atmosphere and lake, dependent on atmospheric stability. Model evaporation is a function of local winds, lake surface temperature, and atmospheric humidity. The rate of evaporation is modified by the atmospheric stability. We do not include precipitation in the model. Ice cover data from NOAA [Assel, 2003; 2005] is applied as a fractional mask to each grid cell at daily resolution. The presence of ice alters lake albedo and prohibits a fraction of the evaporation, heat and momentum exchange. This fraction is equal to the fraction of the grid cell that is ice-covered. Temporal increases/decreases in fractional ice coverage create a heat flux to/from the lake. For this purpose, lake ice is assumed to have a constant thickness of 0.25 meters when present.

The model is forced with 3-hourly winds, downward shortwave and longwave radiation at the surface, air temperature, and specific humidity from the North American Regional
Reanalysis (NARR) [Mesinger et al., 2006] between 1979 and 2006, which has a uniform horizontal resolution of 32 km. NARR is the only dynamically consistent, historical, and freely available public dataset that provides meteorological conditions over all of North America at 3-hourly and 32 km spatial resolution between 1979 and the present (Supplementary Section 1).

The time step of numerical integration is 200 seconds. The physical model is spun up for five years using 1979 forcing before the model is run for 1979-2006. For the entirety of the analysis, summer is June through September (JJAS), to be consistent with the circulation maps presented in Beletsky et al. [1999].

Section 2.2: Model - Data Comparisons

To evaluate the model simulation, we utilize the available temperature and current data throughout the lake.

Section 2.2.1: Model-Data Temperature Comparisons

The National Data Buoy Center (NDBC) observes surface water and meteorological conditions on Lake Superior between April and November at the three open lake buoy stations shown in Figure 1. Buoys 45001, 45004, and 45006 became operable in 1979, 1980, and 1981 respectively. In Figure 2, we show model surface temperature at the three buoy locations during 2000 and 2004. Modeled surface temperatures capture both the seasonal cycle and synoptic variability in surface temperatures observed at the three buoys during 2000. In 2000, root mean square error (RMSE) at buoys 45001, 45004, and
45006 are 1.14, 1.14, and 1.54 °C respectively. During 2004, modeled surface temperatures are higher than observed because the model stratifies earlier than in the data. In 2004, the RMSE at buoys 45001, 45004, and 45006 are 2.78, 3.05, and 2.9°C, respectively. The model generally warms too early in spring and does not cool as rapidly as observed; however, the shapes of the seasonal cycles at the three buoys are consistent with the spatial and temporal heterogeneity seen in the observations. The model captures cooling and warming events on synoptic time scales recorded by the buoys, suggesting the model adequately represents physical processes but has a warm bias. During 2004, temperatures during August and September were below the 28-year mean of 12.1, 12, and 14.3°C at buoys 45001, 45004, and 45006, respectively. The summer of 2000 was warmer than average.

For the entire 28-year period, model RMSE at buoys 45001, 45004, and 45006 are 3.2, 3.0, and 3.4°C. During years with above average temperatures during August and September (warm years), model RMSE are 2.7, 2.4, and 3.1°C; RMSE during colder years (August and September temperatures below mean) are 3.7, 3.7, and 3.8°C. The RMSE for all years at the buoy locations is provided in Supplementary Table 1, and summer average temperatures at buoy locations in the model and data are shown in Supplementary Figure 1.

The Environmental Protection Agency (EPA) has been sampling the physical and chemical characteristics at ~8 depths within the water column at nineteen open lake stations in Lake Superior (Figure 1) twice each year since 1992, usually in April and
August. The sampling depths are not consistent from year to year or between stations, so we compare model temperatures to EPA data (Figure 3) at the EPA depths closest to two depths (surface, 75m) in April and three depths (10m, 20m, 75m) in summer, using all available years on GLENDAGA, the EPA’s searchable data archive website: (http://www.epa.gov/glnpo/monitoring/data_proj/glenda/glenda_query_index.html).

During April, the model overestimates surface temperatures with a root mean square error (RMSE) of 1.07°C at the surface and 1.03°C at 75m. The model is better able to capture observations above 2°C, so the scatter plot is overall flatter than the one-to-one reference line. Summer temperatures with depth are significantly warmer in the model than observed, with RMSE of 4.5, 5.4, and 3.4°C at depths of 10, 20, and 75 m, respectively. Interannual variability in the bias is present, with colder observed years being colder modeled years, just not as cold as observed. The model’s surface warm bias illustrated in these two comparisons can be partially explained by a warm bias in the 10m height air temperature forcing described below.

Section 2.2.2: Warm Bias in NARR Forcing

The North American Regional Reanalysis Project is a long-term, 32 km horizontal resolution climate product for the North American region. Air temperatures above Lake Superior have a warm bias in NARR. When compared to air temperatures observed at the NDBC buoy locations, NARR forcing has a RMSE of 2.4°C during April. NARR 10m air temperatures at the NDBC buoy locations on Lake Superior have large RMSE (4-7°C).
and a warm bias during May through July (Supplementary Figure 2). NARR air temperatures above the lake are biased warm except during the warmest years.

To determine atmospheric conditions above the Great Lakes, the North American Regional Reanalysis Project sets lake surface temperatures to a spatially heterogeneous climatology of its seasonal cycle (personal communication F. Mesinger, D. Schwab). Interannual variability in observed lake surface temperatures are not utilized. Once lake temperature reaches 3.98ºC, the lake stratifies and surface temperatures rapidly increase. Due to averaging, the climatology of lake surface temperatures is significantly warmer than observed during the spring and early summer of the cooler years; thus, boundary conditions for the atmospheric product are too warm during cooler years. A lake that is warmer than observed will cause air temperature above the lake to be warmer than observed in the atmospheric product. The warmer waters may also decrease the atmospheric stability above the lake in spring, when lake temperatures are cooler than air above the lake. Boundary layer parameterizations in the NARR atmospheric model may also cause spring and summer air temperatures to be too warm.

The warm bias in NARR air temperatures causes the model to be too warm in spring and summer. The larger summer heat content means that the heat loss from the lake during winter must be greater than observed for the lake to cool to observed temperatures. This effect and the lack of an explicit lake-ice model contribute to model error during April. If the model does not cool to zero degrees in all years before the ice mask is applied, the
presence of ice impedes lake cooling and traps some heat over winter. This can modify local surface temperatures by up to one degree during spring.

Section 2.2.3: Model-Data Current Comparisons

No lake-wide multi-year current observations exist in Lake Superior. Modeled surface circulation during summer is qualitatively similar to the lake-wide circulation deduced by Harrington’s [1895] bottle drifter experiment. Lake-wide integrated currents were observed by the FWPCA during the late 1960s. We determine the speed and direction of depth-integrated flow at nine locations around the lake, consistent with current observation locations depicted in Sloss and Saylor [1976]. We integrate to 150m, because FWPCA instrumentation did not exceed this depth. Figure 4 depicts the daily speed and direction of integrated flow at the nine stations during summers of 1979-2006. Subplots A through I correspond to the locations shown in Figure 1. Daily depth-average current direction and speed are depicted in the current roses for each location A-I. Bar direction indicates direction toward which current flows. Length of bar indicates percentage of summer days (1979-2006) that daily average flow is in that direction. Color of bar indicates speed of flow. Flow in one direction may have many colors, indicating percentage of days that daily average flow was both in that direction and with that speed. Separate arrow indicates direction toward which current flow was observed by the FWPCA, as summarized by Beletsky et al. [1999]. Color and number at end of arrow indicate mean speed of the observed flows [Sloss and Saylor, 1976]. Due to instrumentation failures, FWPCA data does not span the entire depth and season at all nine selected locations. Exact latitude and longitude coordinates are also not available for
the direct observations, so we selected the model grid cell of the observations by matching the bathymetry to published figures in Sloss and Saylor [1976].

We first note that modeled flows are generally in good agreement with current speed and direction observed by the FWPCA in summer. At locations A, B, C, E, F, and H, modeled currents are in the same direction as observed. At location D, modeled current direction is highly variable and mean observed flow suggests currents toward the northwest. At location G, the model shows a bimodal distribution of integrated direction, and so only agrees with the observed mean for about half of the days of summer. Just east of Isle Royale, at location I, simulated currents are generally more southward than those observed.

Current observations with high temporal resolution are only available near shore during the late 1990s. In spring through winter of 1998 and 1999 during the Keweenaw Interdisciplinary Transport Experiments (KITES), Acoustic Doppler Current Profilers (ADCP) were stationed offshore along the Keweenaw Peninsula (Figure 1,5). The instruments recorded zonal and meridional velocities and temperature throughout the water column. In Figure 5, we compare modeled and observed current velocities and temperature through the water column at station H3 from May to December 1998. The model is able to capture the seasonal cycle of temperature, cooling events, and the magnitude and direction of horizontal velocities unique to the station. The model is able to replicate the observed spatial and temporal heterogeneity, and captures the temperature and velocity patterns. It is also able to capture station to station and year to year
variations (Supplementary Figures 3,4). The model over predicts maximum summer
surface temperatures and begins to stratify prematurely in spring, consistent with
NARR’s warm bias (Section 2.2.2). Heat also penetrates deeper in the model than in the
data, likely due to lack of colored dissolved material in the model, which has been found
to increase the attenuation of light within the water column in the global ocean [Anderson
et al., 2007].

In summary, although the model has a warm bias due a biased meteorological forcing
product, the model captures the spatial and heterogeneity and interannual variability in
the thermal structure and circulation seen in observations. Warming and cooling events
driven by the passing weather systems occur in the correct locations and at the correct
time, as seen in the buoy measurements of lake surface temperature. The model is unable
to capture the coldest temperatures during the coldest years but replicates warmer years
well. Under influence of a warming climate, as experienced between 1979 and 2006,
trends should be dampened in our results given the enhanced warm bias during cold
years. The model should under-predict year-to-year variability and trends. Nevertheless,
the directions and magnitudes of observed currents are captured by the model, and we
believe the model is an adequate tool to analyze the circulation and long term changes in
lake thermal structure.

Section 3: Results

Section 3.1: General Circulation
Figure 6 shows model results for the general circulation averaged over 1979-2006 during winter (DJFM) and summer (JJAS) in the top 15m, below 50m, and integrated throughout the entire water column. Colors depict the average water temperature within the same columns.

During winter, winds are primarily from the north, and horizontal gradients in water temperature are small. Temperature increases with depth to just under four degrees below 50m. The water column can stratify, with coldest water at the surface, but currents are primarily barotropic during winter, indicated by the similarity of surface currents to those at depth. Currents are southward in the western arm of the lake. There is a return flow toward the northeast along the Keweenaw Peninsula. A temperature gradient exists from near to offshore along the peninsula with colder waters near shore. This alone would support a flow toward the southwest if the resulting pressure gradient force was balanced by the Coriolis force (thermal wind). Thus, the northeast flow indicates dominance by the wind. Some northward flow exists in the eastern and central parts of the lake during winter, where there are two rough cyclonic cells just east and north of the Keweenaw Peninsula. These cells are more easily seen when the model is run at coarser (10km) horizontal resolution (Figure 7).

During summer, there are significant near to offshore temperature gradients and winds from the southwest. Warmer waters are along the coastlines, since shallower water heats more rapidly. We choose to separate the flow above 15m and below 50m so that flow in the variable depth of the thermocline zone does not confuse the interpretation. In contrast
to winter surface circulation, summer flows above 15m do not mimic winds. The circulation is largely cyclonic, i.e. the flow pattern is counterclockwise along the eastern coastlines. Flow diverges when it reaches Isle Royale, and there is a southward flow to the Keweenaw Peninsula. The flow east of Isle Royale may be viewed as a single cyclonic surface gyre. The western arm is warmer than in the eastern arm. Circulation is still primarily counterclockwise along the coastlines, but the southwestern-most region of the lake, by the mouth of the St. Louis River, is anticyclonic. Summer circulation is baroclinic, and small-scale structures are omnipresent below the thermocline. Two cyclonic cells follow along the isobaths. A cell exists just east of the Keweenaw Peninsula, and a second, smaller, cell exists north of the Keweenaw Peninsula and east of Isle Royale. Summer integrated flows are strong and counterclockwise along the coastlines. Flow is cyclonic everywhere except in the far southwestern arm, and there is weaker flow along the 200m isobaths in the two cyclonic gyres.

**Current Speed**

Lake-wide mean current speeds peak in October and November (2.8 cm/s) and decay over winter, reaching a minimum during late May (0.7 cm/s). Current speeds are relatively constant between March and May. Although wind speeds are greatest in winter, thermal gradients within the lake decrease and increased ice coverage prohibits the transfer of momentum to the lake. The annual average of lake-wide integrated current speed is 1.5 cm/s. Fast coastal currents begin to develop in June after slowing throughout winter and spring, and coastal currents reach maximum mean velocities of over 5 cm/s during October and November (>15 cm/s in top 15 m). Open-lake current speeds are
weakest during May and June and strongest in October and November (2.5 cm/s), when an ice-free lake begins to experience increasing winds. Offshore flows remain above 2 cm/s until January, when they begin to significantly slow because of ice coverage. (Supplementary Figure 5). Surface velocities (top 15 m) show the same seasonal cycle and structure as integrated velocities throughout the lake, but reach a maximum lake-wide speed of 7.4 cm/s in October and weakest velocities (2-3 cm/s) from March through May.

Variability

In Figure 4, we depict direction and speed of summer currents for all years on a single plot. Selecting any individual year during the model simulation results in current roses with nearly identical distributions of current direction at the nine locations; thus interannual variability in summer integrated currents appears to be minimal. However, variability on synoptic time scales is present, as all current roses exhibit spread in current speed and direction. These distributions are also robust to alternate choices of “summer” months. We conclude that the mean currents depicted in Beletsky et al. [1999] are representative of the flow for nearly all days of summer at locations A, B, F, and H. Currents at locations C, D, E, and G are best understood as probability distributions. While flow at location C is most often toward the northeast, twenty percent of the time, currents are weaker and toward the southwest. Here, flows may enter the cyclonic cell when going toward the northeast. Near the eastern part of the same cell, integrated currents at location D are best described as weak and frequently westward. Significant spread in direction near Sault Saint Marie at location E is also present. In the far western
arm, at location G, currents are bimodal and stronger than averaging observations would suggest. Flows are toward the southwest just as often as toward the northeast, so mean currents only tell part of the story. Even currents along the Keweenaw Peninsula (location A) can weakly reverse direction when winds are able to overcome the temperature gradient [Zhu et al., 2001].

Section 3.2: Current Mechanisms

What controls the current speed and direction? In Lake Michigan, Schwab and Beletsky [2003] find that lake-wide vorticity is primarily generated by vorticity in wind stress during winter and baroclinic effects during summer. Vorticity generated by the bottom topography is a second order effect. Are the mechanisms similar in Lake Superior?

For wind-driven currents we expect that transport within the Ekman layer will be 90° from wind stress. For thermally driven flows, we expect currents to be geostrophic and thus along isotherms, with colder water to the left. To determine how much thermal gradients contribute to both the direction and speed of currents, we correlate daily anomalies of local temperature gradients in the zonal direction with daily anomalies in current speed in the meridional direction ($r_{zt}$). We correlate anomalies of the local temperature gradient in the meridional direction with anomalies in the current speed in the zonal direction ($r_{mt}$). We present the spatial distribution of the geometric mean ($r_g$) of these correlations (Equation 1) in the top of Figure 8.

$$r_g = \sqrt{|r_{zt} \times r_{mt}|}$$

Equation (1)
A geometric mean of 1 indicates a perfect correlation of wind stress anomalies and currents in both zonal and meridional directions. Correlations between temperature gradients and currents may be underestimated due to the use of local temperature gradients.

Similarly, we would like to know how significant an effect wind stress has on local currents. We compute the geometric mean of correlations between anomalies in wind stress in the zonal direction with currents in the meridional direction and correlations between anomalies in wind stress in the meridional direction with current anomalies in the zonal direction. The resulting geometric mean of these correlations is shown in the bottom of Figure 8. We integrate to 25 m, because the Ekman depth is dependent on the eddy flux of momentum and varies in both time and space. Integrating to 50 m does not significantly alter the results or conclusions.

Figure 8 clearly shows that near surface flows in the open lake and immediately onshore are significantly correlated with the wind stress, but currents within the coastal jets are steered along isotherms and are faster flowing when temperature gradients increase. A change in winds is accompanied by an immediate change in surface flows in the open lake and within a couple kilometers of the shore. Similarly, if a temperature gradient from near to offshore is observed, the pattern and speed of the coastal jets may be inferred. We show summer correlations, because they are strongest, but annual correlations show the same pattern. The correlations were computed for lags up to one week, but correlations
were strongest for both wind and temperature gradients with a lag of zero. Thus, changes in wind or temperature gradient translate into a change in currents on the same day.

We cannot decompose integrated flows in the same manner, since we do not expect a change in winds to result in currents to change at a 90° angle below the Ekman Layer. However, the pattern is identical when we correlate daily anomalies in wind speed and integrated current speeds and the magnitude of the thermal gradient and integrated current speeds. Winds are the dominant current mechanism in the open lake and immediately onshore, and temperature gradients control flows near shore. This finding is consistent with sensitivity studies in a model of 10km horizontal resolution (Supplementary Section 2, Supplementary Figure 6). Thus, wind is a first order control of circulation patterns all year, but temperature gradients in summer are the first order control on near-coastal flows.

Although it is apparent winds are important to the circulation pattern, the cyclonic gyres are also coincident with lake bottom topography. We would like to know if the wind-induced pattern would differ without the topographic gradients. To consider this effect, we run the model with a uniform, flat bathymetry of the mean lake depth, such that lake volume is the same as in real-bathymetry simulations. Lake coastlines remain unchanged. In Figure 9, we present monthly average flows and vorticity for January and June of 2006 from the flat bottom and realistic topography runs. Bathymetry significantly changes the horizontal temperature structure of the lake. Shallower water near shore cools to lower temperatures during winter and heats more rapidly during spring and summer. While
temperature gradients control lake flow patterns near the coast in summer, these
temperature gradients are primarily created by variations in lake bathymetry. In the open
lake, wind is the largest driver of lake flow patterns and speeds. However, bathymetry
creates small-scale structures in the open lake flow pattern as a result of a barotropic
response to changes in depth, and local currents are not homogeneous as in the uniform
bathymetry model run. The large gyre that encompasses the central and eastern basin in
the uniform bathymetry run is split into two cyclonic gyres by the presence of bottom
topography (Figure 9). In summary, winds are an important mechanism to circulation
patterns year round, and baroclinic effects alter nearshore currents during summer.
Topographic gradients enhance horizontal temperature gradients during summer, split a
single cyclonic gyre into two cyclonic gyres, and create small-scale structures in local
flows throughout the year.

Section 3.3: Trends

Figure 10 shows annual average modeled integrated temperature (analogous to heat
content from zero degrees), surface temperature, integrated and surface current speeds,
and modeled summer mixed layer depths. The figure also shows model forcing of air
temperature, ice cover, and wind speed (10m) over the lake during the same period.
Slopes of trend lines and their p-values are presented in Table 2. Modeled annual average
integrated temperature increased from 5.2°C (mean 1979-2003) to 5.5°C (mean 2002-
2006). Lake surface temperatures increased from 7.4 to 8.1°C, forced by an increase in
annual atmospheric temperature above the lake of 3.7 to 5.7°C. Modeled temperature
trends likely underestimate realistic trends, given the warm bias in the forcing that causes
the model to fail to capture the coldest observed temperatures but capture temperatures
during the warmest years.

As air temperatures have increased, so have wind speeds above the open lake [Desai et
al., 2009]. Wind speeds increased from 4.9 (1979-1983) to 5.5 m/s (2002-2006) in the
NARR forcing over the 28-year period. The increase in wind speed and lake surface
temperature drives an increase in the wind stress on the lake, resulting in increased
current speeds. Surface current speeds (15m) increased from 4.1 to 5.2 m/s. Mixed layer
depths are determined by surface heating and turbulent mixing, often driven by the
winds. The increase in lake temperatures could result in a shallowing of the thermocline,
but an increase in wind stress should work to deepen the mixed layer. These two drivers
appear to work against each other during the 28-year period; although wind speed is
increasing over the lake, so is lake surface temperature, and the modeled result is no
statistically significant trend in mixed layer depth.

Decreases in ice cover are significantly correlated to increases in air temperature (r = -
0.71) and modeled evaporation (r = -0.75). Lake ice coverage is decreasing by 886
km²/yr, but evaporation has no significant trend. Increases in wind speed also appear to
contribute to increases in evaporation (r = 0.55).

Section 4: Discussion and Conclusions

In this study, we use an eddy-resolving model to simulate and analyze the circulation of
Lake Superior for 1979-2006. We find that summer surface circulation is highly
organized with strong counterclockwise flows around the coastlines. This pattern is in agreement with the first map of Lake Superior circulation created by Harrington [1895] after a bottle drifter experiment. Flows are baroclinic during summer, and open-lake subsurface flows follow along the isobaths, causing integrated currents to consist of two cyclonic cells, strong coastal currents, and weak anticyclonic rotation near the St. Louis River. Circulation during winter is more barotropic, and integrated flows consist of two rough cyclonic cells in the central and eastern basin and weaker coastal currents.

By correlating anomalies in wind stress to anomalies in lake currents, we show that winter lake-wide circulation is primarily controlled by wind. Temperature gradients control currents near shore in summer, while changes in open lake currents are wind-driven all year. It has long been conjectured why medium to large-sized lakes are consistently cyclonic. This study agrees with the findings in Lake Michigan of Schwab and Beletsky [2003] that curl in the wind stress field is the dominant source of lake-wide positive vorticity. Lake Superior is warmer than the surrounding land during winter, causing a mesoscale low pressure system over the lake [Petterssen and Calabrese, 1959], which creates positive vorticity in the wind stress and promotes the cyclonic lake circulation. Due to the effect of lake surface temperatures on atmospheric stability and the presence of horizontal temperature gradients in the lake, even a spatially uniform wind field exerts wind stress with positive vorticity during summer. This effect, proposed by Emery and Csanady [1973], is sufficient during summer to enhance thermally driven currents and capture lake-wide circulation (Supplementary Section 2 and Supplementary
Figure 6). The winter pattern of the circulation cannot be generated without spatially heterogeneous winds.

From a uniform bathymetry run, we determine that lake bathymetry causes significant near to offshore temperature gradients that drive coastal currents during summer. Lake topography creates local heterogeneities in flow and confuses the integrated large-scale circulation pattern year-round. East of Isle Royal, lake topography divides the single cyclonic gyre of the uniform bathymetry simulation into two, more complex cyclonic cells. While side-by-side gyres have been observed in Lake Ontario [Pickett and Richards, 1975] and many small lakes, the gyres were wind-driven and counter-rotating, not both cyclonic. Such counter-rotating cells are sometimes present in the southwestern arm of Lake Superior. Both cells in central and eastern Lake Superior are cyclonic and are the result of a single cyclonic gyre divided by the spatially variable topography.

We have shown that integrated currents vary significantly during the summer season in both direction and speed, but that year-to-year changes in circulation at nine locations throughout the lake are minimal. Waples and Klump [2002] determined that wind directions have shifted over all of the Great Lakes except for Superior, so the lack of significant year-to-year variability may not hold true in other Great Lakes. Still, Lake Superior is undergoing significant changes. Modeled annual lake surface temperatures increase 1°C during the 28-year simulation, despite a known underestimate due to warm biases in the model forcing that are pronounced during cold years. Austin and Colman [2007] analyzed spring through summer buoy temperatures and determined that Lake
Superior surface temperatures during this time of year are warming twice as rapidly as regional air temperatures. For the lake to freeze, the entire water column must first cool to 3.98°C, and then the surface layer may begin to cool. Strong winter winds often make this surface layer very deep. Winter air temperatures determine lake ice coverage and spring lake heat content. A warmer winter leads to a larger spring heat content and an earlier onset of stratification, because less heat needs to be added to warm the entire water column to 3.98°C. This earlier onset of stratification leads to warmer lake surface temperatures throughout summer, causing a significant correlation between winter ice coverage and summer surface temperatures. Our findings are in agreement with Austin and Colman [2007].

While Hanrahan et al. [2010] find an increasing trend in evaporation from GLERL modeling results in Lake Michigan and Lake Huron beginning in the 1980s, we do not find a statistically significant increase in evaporation in Lake Superior. The model indicates that ice cover is the dominant mechanism controlling annual evaporation. Lake surface temperatures and wind speeds are also increasing above the lake, but these changes are not large enough to cause a significant change in annual evaporation during this period.

The warmer lake surface temperatures are causing decreased atmospheric stability above the lake, causing an increase in wind speeds above the lake and an increase in lake current speeds [Desai et al., 2009]. This mechanism should drive lake surface temperatures to continue to increase in warmer conditions due to anthropogenic climate change.
warming in the coming decades. This should lead to faster lake currents and a decreased lake mixing time. These changes could spread pollutants and invasive species (mussels) more rapidly around the lake. Increased lake temperatures may also allow sea lampreys to begin feeding earlier in spring and thus to grow larger and more deadly to host fish \cite{Kitchell and Breck, 1980} and may require updated management techniques. Increased and systematic year-round monitoring of lake thermal structure throughout the lake began in 2009 by investigators at the University of Minnesota-Duluth and the Large Lakes Observatory (J. Austin, personal communication) and will increase our understanding of the current, and likely future, physical state of the lake. These data and further numerical simulations can shed light on Lake Superior’s physical and ecological response to climate change.
Acknowledgments

The authors would like to thank NSF for funding (grant number) and David Schwab at GLERL for providing meteorological observations over the lake for the five years (1997-2001) alternative forcing.
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Prognostics modeling studies of the Keweenaw current in Lake Superior. Part II:
Figure Captions

Figure 1. Locations of observations in Lake Superior used for model evaluation. Solid black circles are 19 open lake EPA stations. Three black crosses are locations of three surface buoys operated by NOAA. A through I are locations of current measurements from Sloss and Saylor [1976]. E3, E4, and H4 are ADCP locations during the KITES experiment. Isobaths every 100 m. Inset of boxed region along the Keweenaw Peninsula. Isobaths within inset every 50 m.

Figure 2. Daily average modeled and observed lake surface temperatures at the three NDBC buoy locations during 2000 (left) and 2004 (right). 2000 was a year of above-average lake temperatures, and 2004 temperatures were below the 28-year average.

Figure 3. Model daily average temperature plotted against instantaneous temperature measured at the 19 EPA stations between 1992 and 2006 during April (a) and summer (b). Summer sampling primarily done during the month of August. Model temperatures compared to sample temperature at depths nearest the surface and 75m in April. In summer, model temperatures compared to sampled temperatures nearest the depths of 10m, 20m, and 75m. One-to-one line shown on both (a) and (b) for ease of comparison.

Figure 4. Subplots A through I correspond to the locations shown in Figure 1. Daily depth-average current direction and speed are depicted in the current roses for each location A-I. Bar direction indicates direction toward which current flows. Length of bar indicates percentage of summer days (1979-2006) that daily average flow was in that
direction. Color of bar indicates speed of flow. Flow in one direction may have many
colors, indicating percentage of days that daily average flow was both in that direction
and with that speed. Separate arrow indicates direction toward which current flow was
observed at that location during two years during the 1960s, as summarized by Beletsky
et al. [1999]. Color and number at end of arrow indicates mean speed of the observed
flows [Sloss and Saylor, 1976].

Figure 5. Zonal velocity (U), meridional velocity (V), and temperature (T) at ADCP
location H3 during the summer of 1998. Data in left column, model on right. Data and
model plotted versus depth (m) and time (day). Month denoted on first day of the month.
Data averaged within model layers. For ease of comparison, model layers with no ADCP
measurements are left blank for both data and model.

Figure 6. (a,b) Mean integrated currents and column temperature during winter (DJFM)
(1979-2006) (a) and summer (JJAS) (1979-2006) (b). (c,d) Mean currents and column
temperature within the top 15 m of the water column during winter (c) and summer (d).
(e,f) Mean currents and temperature of the water column below 30 m during winter (e)
and summer (f). Note change in current scale for surface flows.

Figure 7. Mean integrated current direction during winter (DJFM) and summer (JJAS) of
1997-2001 from a model simulation with a 10km horizontal grid resolution. Current
speed is not indicated by arrow length.
Figure 8. (a) Geometric mean of: 1) correlation between daily summer zonal temperature gradient anomalies and meridional current anomalies (integrated to 25 m) and 2) correlation between daily meridional temperature gradient anomalies and zonal current (integrated to 25 m) anomalies during summer. A geometric mean of 1 corresponds to a correlation of 1 for both 1 and 2. (b) Geometric mean of: 1) correlation between daily summer zonal wind stress anomalies and meridional current (integrated to 25 m) anomalies and 2) correlation between daily summer meridional wind stress anomalies and zonal current (integrated to 25 m) anomalies.

Figure 9. (a,b) Lake-wide integrated currents and temperature in January 2006 for model simulations with and without realistic bathymetry. Color indicates temperature. (c,d) Integrated vorticity during January for the uniform and realistic topography runs. (e,f) June integrated currents and temperature and (g,h) vorticity for the uniform and realistic topography simulations.

Figure 10. Lake-wide annual average modeled lake surface temperature, integrated temperature, surface current speed, integrated current speed, and NARR 10 m above-lake air temperature and wind speed from 1979-2006.
## Tables

Table 1. $R^2$ Values of lake variables of interest

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Table 2. Lake Trends and p-values

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Figures

Figure 1.
Figure 2.
Figure 3.
Figure 4.
Figure 5.

DATA

U (cm/s)

MODEL

U (cm/s)

V (cm/s)

T (°C)

V (cm/s)

T (°C)
Figure 6.
Figure 7.

Mean Integrated Winter Currents

Mean Integrated Summer Currents
Figure 8.
Figure 9.
Figure 10.

Lake Superior Trends

(a) Temperature (°C)

(b) Ice cover, evaporation, mixed layer depth
Supplementary

Section 1: Model Forcing

The model was run for 1997-2001 using both the NARR atmospheric conditions and over-lake conditions interpolated from meteorological observations from over and around the lake using an empirical method [Schwab, 1978; Hsu, 1986]. Model results from these simulations were compared to observed lake surface temperatures at the three open-lake buoys, thermal profiles at the 19 EPA stations, and ADCP measurements off the Keweenaw Peninsula. In Supplementary Table 1, we show root mean square error (RMSE) at the three buoy locations for model simulations with both types of forcing. The model simulation using NARR forcing has lower RMSE at all buoy locations during 1998, a warm El Niño year, and 2000. During 1999, running the model using NARR forcing results in a lower RMSE at one of the three buoys. During the five years, warmer years were better simulated using NARR forcing, and colder years were better captured using Schwab forcing. Both forcing products are reasonable.

The North American Regional Reanalysis Project uses well-documented, internally-consistent methods and a uniform horizontal resolution of 32km to create their product. Meteorological observations above the lake are often more than 100 km apart, and as one goes back in time, fewer and fewer direct atmospheric observations over the lake exist. We opted to use NARR atmospheric conditions to force the model for 1979-2006.

Supplementary Section 2: Circulation Sensitivity Studies
To assess the importance of heat fluxes, winds, and vorticity in the wind field, we simulated lake circulation for 1997-2001 in three sensitivity runs using the model at 10km horizontal resolution. The model was run without any wind (test 1), without heat fluxes (test 2), and with a spatially uniform wind field (test 3). In test 1, winter flows are extremely weak compared to the simulation with wind. Only one gyre is present in the central and eastern basins, and the western arm is anticyclonic instead of cyclonic. During the summer, the lack of wind results in negative vorticity in the western arm and weakened flow in the far eastern basin. When the model was run without heat fluxes, winter circulation patterns are similar to the control. Summer circulation still has a cyclonic gyre in the eastern basin, but the north-central basin is now anticyclonic. Heat fluxes are crucial to the development of the coastal jet. A uniform wind results in many small rotating cells without any organized pattern during winter. A spatially varying wind is less important during summer, because horizontal lake temperature gradients generate vorticity in the wind stress field. Thus, wind is a first order control of circulation patterns all year. Vorticity in the wind field is crucial for winter circulation patterns, but not summer patterns. Temperature gradients in summer are a first order control of near-coastal flows.
Supplementary References


Supplementary Figure 1. April through November average lake surface temperatures at the three NDBC buoy locations in the model and data for 1979-2006. We use only modeled temperatures on dates with available observational data in each yearly mean.

Supplementary Figure 2. Daily average 10m air temperature from the North American Regional Reanalysis Project is plotted against observed air temperature at the three NDBC buoy locations in Lake Superior for all days of available observations between April and September of 1979-2006. The one-to-one line is shown for ease of comparison. Correlation between NARR and observed air temperatures and RMSE of NARR air temperatures are provided for each month.

Supplementary Figure 3. Zonal velocity (U), meridional velocity (V), and temperature (T) at ADCP location E4 during the summer of 1998. Data in left column, model on right. Data and model plotted versus depth (m) and time (day). Month denoted on first day of the month. Data averaged within model layers. For ease of comparison, model layers with no ADCP measurements left blank for both data and model.

Supplementary Figure 4. Zonal velocity (U), meridional velocity (V), and temperature (T) at ADCP location E4 during the summer of 1999. Data in left column, model on right. Data and model plotted versus depth (m) and time (day). Month denoted on first day of the month. Data averaged within model layers. For ease of comparison, model layers with no ADCP measurements left blank for both data and model.
Supplementary Figure 5. Mean integrated current speeds for January through December 1979-2006. Contours of lake bathymetry every 100m in dark blue.

Supplementary Figure 6. Mean integrated current direction and speed during winter (DJFM) and summer (JJAS) of 1997-2001 when the model with 10km horizontal resolution is run (a) without wind, (b) without heat fluxes, and (c) with a spatially uniform wind field.
Supplementary Tables

Table 1.

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Supplementary Figures

Supplementary Figure 1.

Buoy Temperatures

- --- model buoy SST (AMJJASON)
- - - data buoy SST (AMJJASON)
Supplementary Figure 2.
Supplementary Figure 3.
Supplementary Figure 4.
Supplementary Figure 5.

[Images of monthly flow graphs from January to December]
Supplementary Figure 6.