



Development of the Great Lakes Ice-circulation Model (GLIM): Application to Lake Erie in 2003–2004

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ABSTRACT

To simulate ice and water circulation in Lake Erie over a yearly cycle, a Great Lakes Ice-circulation Model (GLIM) was developed by applying a Coupled Ice–Ocean Model (CIOM) with a 2-km resolution grid. The hourly surface wind stress and thermodynamic forcings for input into the GLIM are derived from meteorological measurements interpolated onto the 2-km model grids. The seasonal cycles for ice concentration, thickness, velocity, and other variables are well reproduced in the 2003/04 ice season. Satellite measurements of ice cover were used to validate GLIM with a mean bias deviation (MBD) of 7.4%. The seasonal cycle for lake surface temperature is well reproduced in comparison to the satellite measurements with a MBD of 1.5%. Additional sensitivity experiments further confirm the important impacts of ice cover on lake water temperature and water level variations. Furthermore, a period including an extreme cooling (due to a cold air outbreak) and an extreme warming event in February 2004 was examined to test GLIM's response to rapidly-changing synoptic forcing.

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Introduction

Lake ice cover in the Great Lakes region can have an important impact on the regional weather and climate: two examples are lake-effect snow in winter and modulation of regional surface air temperature (SAT). Lake ice cover can also modify the lake circulation patterns and thermal structure because: 1) momentum transfer into the water column from wind stress drag is considerably greater over the water surface than over the ice surface; 2) the albedo over ice differs from that over water, and 3) heat and moisture exchange between the atmosphere and the lake water can differ significantly (as much as orders of magnitude different) with and without lake ice (Walter et al., 2006), thus leading to a striking difference in evaporation in wintertime due to strong cooling and wind mixing. Prediction of the lake's ice extent (i.e., ice cover) is crucial for predicting the lake's mixed layer, circulation, temperature, and water level, and thus for predicting primary and secondary productivity. In addition, the timing of ice melt, determined by SAT that is controlled by climate variability, will determine the timing of spring phyto-

plankton and zooplankton blooms (Vanderploeg et al., 1992). As a result, lake ice cover, although thin, is an important physical parameter for other ice-associated systems such as ecosystems and habitats for fisheries. This is in part because lake ice dynamics and thermodynamics significantly modify the water temperature, heat flux, mixing intensity, and water column stratification, which are important factors controlling phytoplankton blooms.

The Great Lakes are usually at least partially covered with ice from December to April. Initially, ice begins to form in shallow bays and then gradually grows offshore. Maximum ice extent is normally observed in late January to early February, when ice typically covers from 24% of Lake Ontario to 90% of Lake Erie (Assel et al., 1983). Naturally-formed ice thickness can vary from a few centimeters to a meter or more (Rondy, 1976). Ice decay and breakup usually begin in March as solar radiation increases, and the thinner ice can then be more easily broken up by the action of wind and waves. Recent observations of sensible and latent heat fluxes over Lake Erie (Gerbush et al., 2008) show a rapid decrease in flux magnitude as ice concentration approaches 100%.

The presence of ice cover also affects momentum transfer between the atmosphere and the water column, which determines waves and circulation patterns in a large lake. Momentum transfer is generally reduced by the presence of ice. Measurements of ice movement in Lake Erie using drifting buoys in winter 1984 show that wind is the

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major forcing to ice transport in the Great Lakes (Campbell et al., 1987). They reported that the mean observed speed of the buoys in ice is about 8 cm s^{-1} , half the mean speed observed in open water. An experiment to obtain under-ice currents in Lake Erie was conducted in 1979–80 (Saylor and Miller, 1983), but no specific analyses for the impact of ice on the lake circulation were made.

Lake Erie ice is first year ice, with ice thickness being typically a few centimeters to 1 m or more due to ice ridging or rafting caused by wind and waves. Synoptic weather patterns and cyclone passage (Lofgren and Bieniek, 2008) can significantly affect lake ice distribution. Thus, since the predictability of lake ice using statistical methods is poor due to the complexity of the climate patterns (Assel and Rodionov, 1998; Mysak et al., 1996; Wang et al., 2010) and highly dynamic regional weather patterns, numerical ice modeling is an important tool to help understand lake ice thermodynamic and dynamic features on synoptic time scales.

Wake and Rumer (1979, 1983); Rumer et al., 1981 developed a numerical model of ice transport in Lake Erie based on Hibler's (1979) dynamic–thermodynamic sea ice model, but no further progress has been made since then, perhaps due to a lack of resources and initiative. At present, there exists no viable ice model for use as a research and operational forecast tool in the Great Lakes, which is long overdue. However there have been some successful efforts in coupled ice–ocean modeling in many subpolar seas and bays, such as in Hudson Bay (Wang et al., 1994; Saucier and Dionne, 1998; Saucier et al., 2004), in the Gulf of St. Lawrence (Saucier et al., 2003), in the Baltic Sea (Meier, 2002a,b; Haapala, 2000; Haapala et al., 2001), and in the Labrador Sea (Yao et al., 2000; Tang, 2008). These areas are similar (except for salinity) to the Great Lakes because they do not have perennial ice cover.

The Great Lakes Coastal Forecasting System (GLCFS) presently predicts lake water circulation, temperature, and surface waves (<http://www.glerl.noaa.gov/GLCFS>). Since it currently does not have a lake ice component, empirical methods have been used to keep the system running over the winter. Wave forecasts also must be modified, as ice cover dampens surface waves significantly during winter. Thus, it is inadequate to use only a circulation model to investigate hydrodynamics and thermodynamics when lake ice is present. The increasing need for predicting lake ice for navigation, weather forecasting, rescue efforts, and ecosystem studies motivated us to develop a coupled ice–circulation model.

The next section briefly describes the model, forcings, and data used to validate the model. The section of Simulation results presents physical explanations of lake ice dynamics and thermodynamics,

and the model validation using satellite and in situ measurements, followed by the Summary and conclusions.

Description of GLIM, atmospheric forcings, and validation data

The GLIM is a combination of the Coupled Ice Ocean Model (CIOM) developed and applied to the Arctic Ocean and subpolar seas (Yao et al., 2000; Wang et al., 2002, 2003, 2005, 2009) and the Great Lakes version of the Princeton Ocean Model (POM, Schwab and Bedford, 1999; Beletsky and Schwab, 2001; Beletsky et al., 2003, 2006). The CIOM is based on a thermodynamic and a dynamic model with a viscous–plastic sea ice constitutive law (Hibler, 1979) and a multi-category ice thickness distribution function (Thorndike et al., 1975; Hibler, 1980) coupled to the Princeton Ocean Model. The coupling is governed by the boundary processes as discussed by Mellor and Kantha (1989).

The principal difference between the GLIM and the CIOM is the adaptation of heat and momentum flux submodels from the POM-based Great Lakes Coastal Forecasting System (Schwab and Bedford, 1999) so that during the ice-free season, the model is identical to the Great Lakes version of POM. Heat and momentum flux over the lake are calculated using a bulk aerodynamic approach using estimates of wind speed, air temperature, dew point, and cloud cover, which are interpolated to each grid point from hourly surface observations at a network of stations (Fig. 1) in and around the lake (Beletsky et al., 2003). Measurements are adjusted to a common 10 m anemometer height above the water surface using the profile method developed by Schwab (1978) and described more fully by Liu and Schwab (1987). The profile method employs the Charnock relation for increasing surface roughness with increasing wind speed and profile similarity theory presented by Businger et al. (1971) to describe the dependence of the profile on atmospheric stability.

Over open water, the profile theory is used at each grid square at each time step to estimate surface stress using the surface water temperature from the circulation model. This procedure provides estimates of bulk aerodynamic transfer coefficients for momentum and heat. Surface heat flux H is calculated by

$$H = H_{sr} + H_s + H_l + H_{lr}, \quad (1)$$

where H_{sr} is the short-wave radiation from the sun, H_s is the sensible heat transfer, H_l is the latent heat transfer, and H_{lr} is the long-wave radiation. The heat flux procedure follows the methods described by McCormick and Meadows (1988) for mixed layer

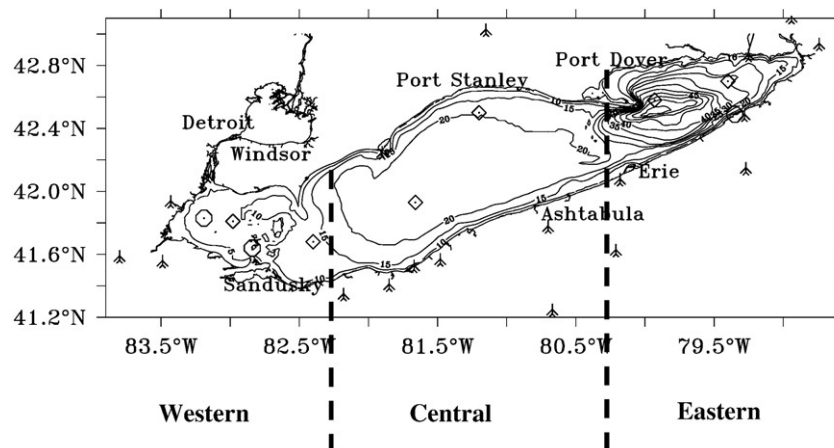


Fig. 1. Lake Erie bathymetry (depths are in meters) and the model domain with 2-km resolution. The meteorological forcing of the model is derived from the NDBC (National Data Buoy Center) buoys ($\diamond$), C-MAN (Coastal Marine Automatic Network) stations ($\circ$), and local airports. The vertical dashed lines (82.4°W and 80.4°W) divide Lake Erie into the western, central, and eastern basins.

modeling in the Great Lakes. H_{sr} is calculated on the basis of latitude and longitude of the grid square, time of day, day of year, and cloud cover (CL).

$$H_{sr} = H_{cs}F_3(CL) \quad (2)$$

where H_{cs} is a clear-sky value, and F_3 is a cubic function of cloud cover that ranges from 1.0 for clear sky to 0.36 for total cloud cover. H_s and H_l are calculated using the bulk aerodynamic transfer formulas:

$$H_s = C_h C_p \rho_a u_w \Delta T \quad (3)$$

$$H_l = C_d q_1 \rho_a u_w (h_a - h_w) \quad (4)$$

where C_h is the bulk heat coefficient, C_p is the specific heat of air at constant pressure, ΔT is the water–air temperature difference, C_d is the drag coefficient, q_1 is the latent heat of vaporization, h_a is the specific humidity of air, and h_w is specific humidity at the water surface. H_l is calculated as a function of T_a , T , and cloud cover according to Wyrski (1965). McCormick and Meadows (1988) showed that this procedure works quite well for modeling mixed layer depth in the Great Lakes and it has been used with very good success in the Great Lakes version of POM (Schwab and Bedford, 1999; Beletsky and Schwab, 2001; Beletsky et al., 2006).

When ice is present in a grid square, heat and momentum fluxes are calculated as described in Wang et al. (2005) with the following exceptions:

- 1) Wind stress on the water squares where ice is present is reduced by 0.5 times the ice concentration in that square (note that this empirical method will be discussed later).
- 2) Short-wave radiation into grid squares where ice is present is calculated using surface albedo equal to 0.28 times the ice concentration.
- 3) Long-wave radiation in grid squares where ice is present is calculated the same as for open water.

The domain of the Lake Erie model (Fig. 1) used the same 2-km grids as the circulation only model (Schwab et al., 2009). There are 21 sigma levels in the vertical. In the present study, 10 ice thickness categories (0, 0.10, 0.20, 0.3, 0.4, 0.6, 0.8, 1.0, 1.2, and 1.4 m) were used, and the amount of ice in each category is calculated at each grid point. Thus, ice thickness at each grid was calculated from the sum of the 10 ice categories. The external and internal time steps for the lake model are 20 s and 300 s, respectively. The ice model uses the internal time step of the lake model. Other parameters are listed in Wang et al. (2002, 2005).

The GLIM was initialized with a uniform water temperature of 2 °C on April 1, 2003, while salinity was set to zero. Initial ice concentration, thickness, and velocity were set to zero. The model was run from April 1, 2003 to December 31, 2004 using measured hourly atmospheric forcings derived from the coastal stations.

The hourly atmospheric forcings (Fig. 2) were taken from in situ measurements at 29 stations (Fig. 1) around the lake including the National Data Buoy Center (NDBC) buoys and the Coastal Marine Automated Network (C-MAN) stations. The station data were interpolated into the model domain using the objective analysis technique widely used in meteorology, which has been used in the present Great Lakes Coastal Forecast System (GLCFS), developed by Schwab and Bedford (1999). Seasonal variations of SAT and lake surface temperature (LST) during 2004 (Fig. 2a) indicate that the SAT fluctuation is larger than LST, particularly in the winter season. Cold air outbreaks occurred during December–February. SAT warming events occurred in spring, while cooling events occurred in autumn. The hourly wind vectors (Fig. 2b) indicate that wind direction varies in time, due to strong synoptic weather events, with high wind in winter. However, the monthly average wind shows that the dominant wind is westerly from November to March.

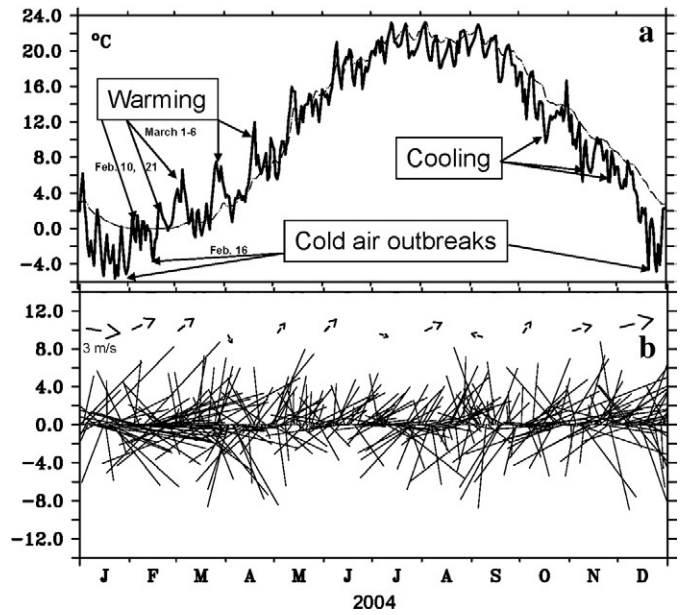


Fig. 2. a) Hourly observed domain-averaged air (Thick line) and surface water (Thin line) temperature in 2004. b) Plot of daily mean wind speed vectors derived from the hourly data during 2004 at (80 W, 42 N), with monthly mean wind vectors.

To describe wind direction in detail, lake-wide average wind roses for four seasons were constructed (not shown). In winter, southwesterly winds dominate (16%), and northwesterly places second (12%). Occasionally, easterly winds also appear (6%), possibly due to the passage of winter storms. In spring, southwesterly winds dominate (12%), while there are also northerlies and northeasterlies (~6%). In summer, the dominating winds are still westerly and southwesterly (17%), while easterly places second (13%). In autumn, wind directions are more variable, with equal distribution, indicating that winds blow from all directions. In summary, the wind directions vary profoundly in each season, although with more frequent westerly events, indicating active synoptic storms in the Great Lakes region year round.

Satellite-measured LST was derived from AVHRR (Advanced Very High Resolution Radiometer). Satellite-retrieved ice concentration was derived from the National Ice Center (NIC) Great Lakes Ice Analysis Charts, which are based on Radarsat-2, Envisat, AVHRR, GOES (Geostationary Operational and Environmental Satellites) and MODIS (Moderate Resolution Imaging Spectroradiometer). These satellite measurements were used to validate the seasonal variations of the simulated ice cover and LST. Spatial variability of the satellite-measured ice cover and LST were also used to validate the modeled spatial patterns of lake ice and LST. In situ water level measurements were also used to validate the simulated water level with and without ice.

To measure the GLIM's skill for reproducing the measurements, two statistical measures or skills are introduced to conduct the model-data comparison. Mean bias deviation (MBD) is defined as

$$MBD = 100 \frac{\frac{1}{N} \sum_{i=1}^N (x_i - y_i)}{\frac{1}{N} \sum_{i=1}^N y_i} = 100 \frac{\bar{X} - \bar{Y}}{\bar{Y}}, \quad (5)$$

and root mean square deviation (RMSD) is defined as

$$RMSD = \left[\frac{\sum_{i=1}^N (x_i - y_i)^2}{N} \right]^{1/2}, \quad (6)$$

where x_i and y_i ($i = 1, 2, 3, \dots, N$) are the modeled and observed time series of any variable such as ice area, LST, etc., N is the total sampling number, and the overbars denote the average of the time series. MBD directly measures the relative bias or error of the modeled time series from the observed in percentage. RMSD measures the absolute error of the modeled time series against observation.

Simulation results

Seasonal variation of Lake Erie ice

Temporal variations

Fig. 3a shows the 2003/04 seasonal cycle of ice area, which is defined as the product of the grid area and ice concentration, from December 1, 2003 to April 30, 2004. The model simulation of the seasonal cycle compares reasonably well to the satellite-derived ice area. Lake ice started to form in December 2003, and grew slowly. Lake ice grew rapidly in January and its area reached a maximum around January 20, 2004 and persisted until mid-February. Lake ice suddenly retreated on February 22, because of a warming event due to a cyclone passage (see Fig. 2a), which will be discussed in detail later. Beginning in March, lake ice rapidly decreased because two warming events persisted throughout March. Lake ice was completely melted by mid-April. There is some discrepancy between the simulation and the measurements with MBD being 7.4% and RMSD being 1840 km². For example, the model produces more ice than was observed from late January to early February, and less ice in late March (see Fig. 3a). The simulated basin-averaged ice thickness was computed from the ice-covered area for the 2003/04 ice season (Fig. 3b). The maximum domain-averaged ice thickness in late January and early February was about 9 cm, and increased to ~10.5 cm on February 17 due to a cold air outbreak on February 16 (Fig. 2a). In late February, the thickness rapidly reduced to ~7.5 cm due to a warming event on February 21. Although there were no ice thickness measurements available for comparison in 2004, we conducted a field measurement of ice thickness at nine locations on Lake Erie with U.S. Coast Guard helicopter support on February 27, 2008 (Fig. 4). The 2007/08 ice season had an average maximum ice cover of 55%, comparable to the

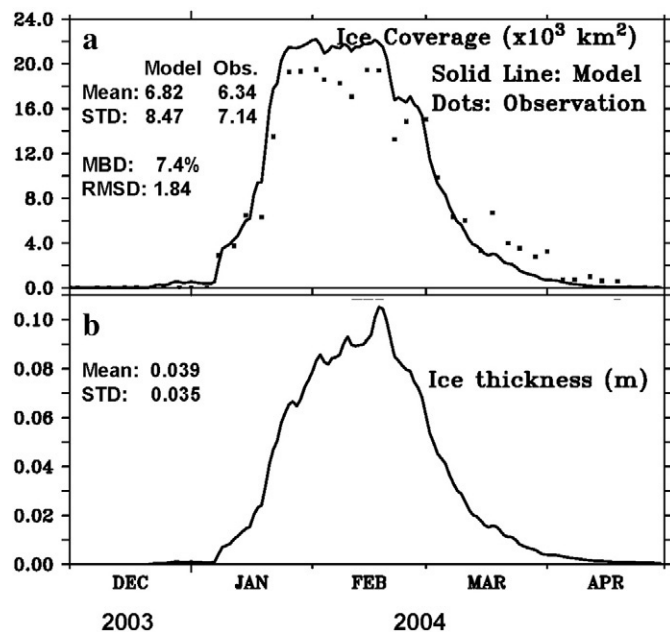


Fig. 3. a) The GLIM-simulated ice area (solid, in km²) and satellite-measured ice area (dots). b) Simulated domain-averaged ice thickness (in meters) from December 1, 2003 to April 30, 2004. The simulated and observed means and standard deviations, and MBD and RMSD were provided.

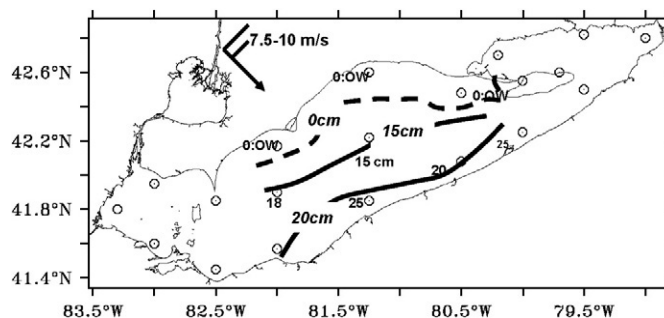


Fig. 4. Ice thickness measurement on February 27, 2008 conducted by using a USCG helicopter. The average wind speed and direction for this day are also shown. The squares with numbers were the planned measurement stations (1–20), and the numbers are the thickness measured in centimeters. OW denotes open water. The italic numbers denote the thickness contour lines of 0, 15, and 20 cm, respectively. Daily wind was shown to be northwesterly at 7.5–10 m s^{−1}.

2003/04 ice season average of 50% coverage, over the Great Lakes. Ice thickness at the stations ranged from open water (0 cm) to 25 cm. The average thickness of these nine stations is 9 cm (Table 1). The model-simulated domain-averaged ice thickness is 7.5 cm on February 27, 2004, which is comparable to the measurement. Since the overall maximum ice coverage for these two years was similar, it can be assumed that ice growth and decay during the 2007/08 ice season was similar to that in 2003/04.

Spatial variability

Fig. 5 (left column) shows the composite satellite-measured ice concentration during the 2003/04 ice season. On January 9 (not shown), 2004, the observation shows that landfast ice formed in the western basin, while in the eastern basin, there was little ice along either the south or north shores. On January 16 (Fig. 5), the landfast ice further expanded eastward and formed along the entire coast. Ice forms more rapidly along the shores because water temperature reaches the freezing point in shallow water before it does in deeper water. On January 23 (not shown), lake ice completely covered the entire lake, although with variable concentration. By January 30 (Fig. 5), Lake Erie was completely ice-covered with >90% ice concentration, except for part of the eastern lake that had 70–80% ice concentration. Complete ice coverage persisted until February 20 (not shown). Lake ice started to break up in late February (Fig. 5c) along the north shore and south shore, and along the shoal between the central and western basins due to the sudden decrease in water depth from 5 m to 10 m. The landfast ice west of the islands in the western lake was still 100%, because of its attachment to the shores. To the east of the islands, pack ice broke up early due to water depth (~10 m), because pack ice in deep water was more mobile than in shallow water in response to the same atmospheric forcing. Under solar warming and wind forcing, breakup of landfast ice and pack ice near the islands was where breakup occurs first, mainly due to the discontinuity in the 10-m isobath. This phenomenon is similar to the Beaufort Sea where landfast ice is confined within the 20-m isobath (Mahoney et al., 2007), and breakup occurs first along the 20-m isobath, rather than along the coast. During March (Fig. 5), rapid melting continued from the western to the eastern lake.

Table 1

Measurement of ice thickness (in cm) at nine stations in Lake Erie on February 27, 2008. The maximum ice coverage was 55% in the 2007/08 ice season, similar to 50% in the 2003/04 ice season. The unit for thickness is cm.

Station	6	7	8	9	10	11	12	15	16	Mean
Lon (°N)	82.0	82.0	82.0	81.25	81.25	81.25	80.5	80.0	80.0	
Lat (°W)	42.17	41.90	41.57	41.85	42.22	42.6	42.08	42.55	42.25	
Thickness	0	18	9	25	15	0	20	0	25	9

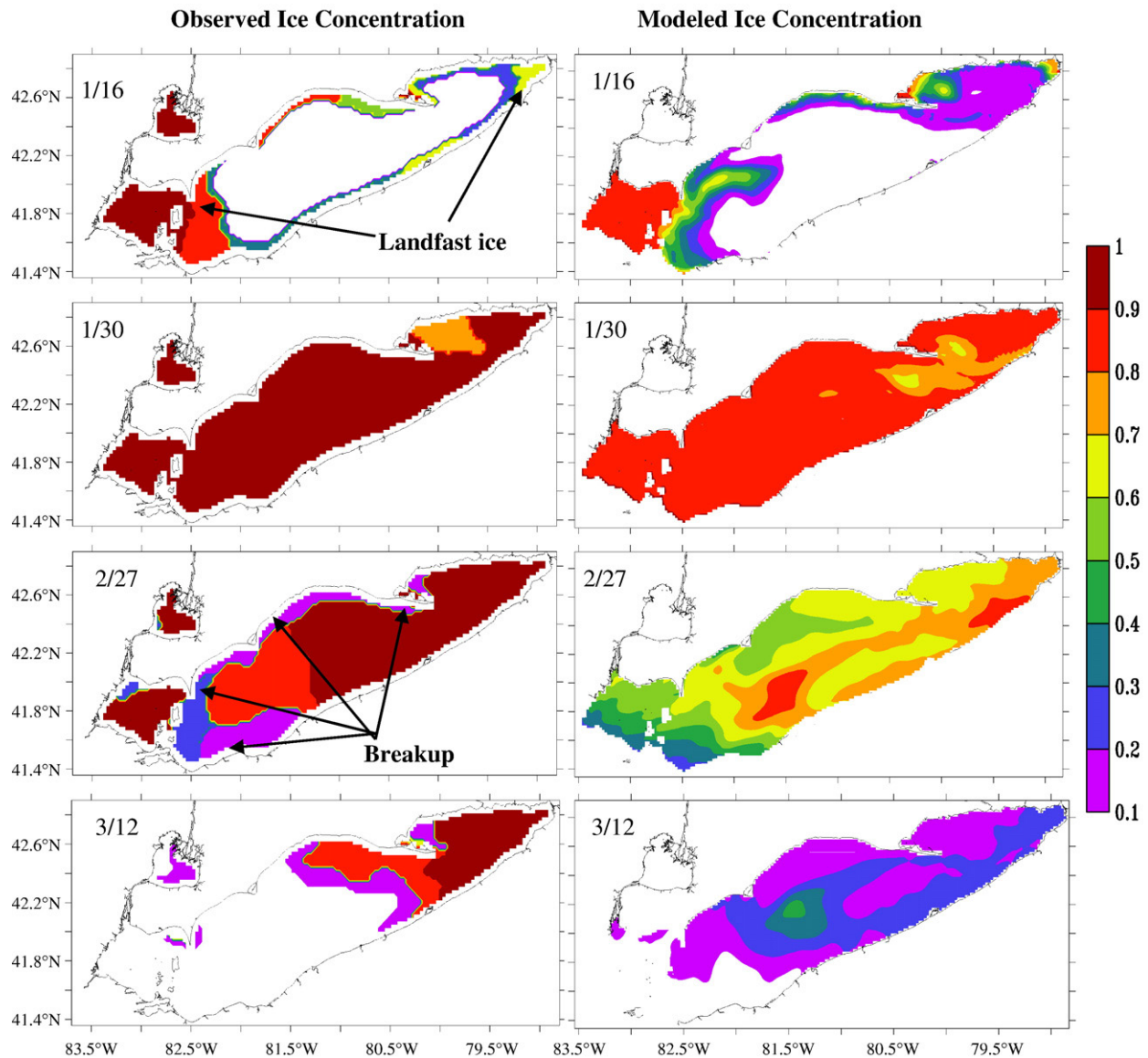


Fig. 5. (Left column): Spatial distribution of the satellite-measured, 3-day composited ice concentration in 2004 on January 16 and 30, February 27, and March 12. (Right column): Spatial distribution of the model-simulated daily ice concentration in 2004 on January 16 and 30, February 27, and March 12.

Fig. 5 (right column) also shows the modeled spatial variability of lake ice concentration of the 2003/04 ice cycle. Landfast ice formed from the western lake because the water depth is only 5–10 m. The simulated landfast ice formed along the north coast (Fig. 5) due to faster depletion of heat storage in the shallower water, similar to landfast ice in coastal seas such as the Beaufort Sea (Mahoney et al., 2007) and the Baltic Sea (Meier et al., 2002a,b; Haapala, 2000). In the Beaufort Sea, the 20-m isobath is the key connection between landfast ice and pack ice (Eicken et al., 2005; Mahoney et al., 2007). However, the model produced less landfast ice along the south coast (Fig. 5) compared to the observation (Fig. 5). Lake ice expanded from the north shore to the south shore on January 23 (not shown), which has a large discrepancy from the observation. On January 30 (Fig. 5), the model simulates complete ice coverage, which compares favorably to the measurement. Ice eventually formed in early February in the deep eastern basin (~60 m).

Note that the observed maps (Fig. 5, left column) are 3-day (twice weekly) composite average maps, while the simulated ice concentration maps (Fig. 5, right column) are daily averages, which were driven by hourly measured atmospheric forcing. Therefore, it is

not surprising that the spatial discrepancy is more obvious than the domain-average time series (Fig. 3a).

Breakup began in the shallow areas of the western basin in late February (Fig. 5), and rapid melting occurred in early March (Fig. 5), again because shallow water is heated faster by solar radiation than deep water. An important mechanism, the so-called ice/water albedo feedback, comes into play to accelerate the melting process once the breakup begins because of the reduction in surface albedo and strong mixing due to waves and wind. Thus, water temperature rises faster in shallow water than in deep water due to this positive ice/water albedo feedback. Lake ice melted quickly from the western to the eastern lake (Fig. 5), and eventually in the deep basin, qualitatively consistent with observations.

Nevertheless, the ice breakup process, including the landfast ice breakup in particular, was not well simulated by the model compared to the timing and spatial distribution of the ice measurements (Fig. 5, left column). This difficulty was also encountered in the Arctic seas (Wang et al., 2008). The major problem might be that the ice/water albedo feedback was not accurately parameterized to include phenomena such as melting ponds on the ice. Mixing by wind waves

and tides/seiches are not included in the model, although these mixing mechanisms accelerate the melting process, in particular in the shallow area and around islands. Therefore, more research is needed on this area in both the Great Lakes and the Arctic seas.

Fig. 6 (left column) shows the simulated lake ice thickness superimposed with ice velocity (black arrows) and wind velocity (green arrows). Seasonal variations of spatial distribution in ice thickness have similar patterns to ice concentration (Fig. 5, right column). Landfast ice first formed in the shallow western basin on January 9 (not shown) and expanded rapidly in mid-January (Fig. 6). Ice continued to grow from the north shore to the south shore and completely covered the whole lake on January 30 (Fig. 6b). Lake ice thickness reached its maximum of about 10.5 cm on February 17 (Fig. 3b) due to a cold air outbreak (see Fig. 2a), but immediately broke up and melted following an intensive warming (see Fig. 2a). On February 27 (Fig. 6), lake ice with an average thickness of 7.5 cm broke up and melted first in the western basin and along the coast. Note that on February 27, 2004 (Figs. 6 and 5), there was a cross-lake gradient in thickness and concentration with more ice near the south shore and less ice near the north shore, consistent with the thickness measurement on February 27, 2008 (Fig. 4), because wind directions on these two days were from the northwest in 2008 and from northeast in 2004. The key is that the component of the northerly winds pushed lake ice to the south shore, forming a cross-lake thickness gradient.

The GLIM also simulates the full 3D hydrodynamic system in the lake, including circulation and temperature structures. The surface circulation and LST in the 2003/04 ice season are shown (Fig. 6, right column). Surface circulation was transient and basically wind-driven

because wind was the most important forcing. It is difficult to draw any conclusions about circulation patterns with no wintertime measurements for comparison. However, the LST is quite consistent with atmospheric seasonal forcing and consistent with ice cover. On January 9, the LST dropped to freezing point in the western lake, leading to the formation of landfast ice (not shown), with the high LST in the deep basin in the central and eastern lake. The freezing temperatures extended from west to east, and from the coast to the center of the basin (Fig. 6). Note that even with complete ice cover on January 30 (Figs. 5 and 6), the LST in the deepest basin was still above the freezing point ($\sim 1^\circ\text{C}$) (Fig. 6). This indicates that heat storage in the deep basin is not completely depleted at the end of January. The remaining heat is continuously advected upward to melt the ice until LST reaches the freezing point. On February 20–27, the LST started to rise over the entire lake (Fig. 6). The LST warmed up from the west to the east, and from the coast to the interior (not shown).

Regional and seasonal characteristics

Basin-scale lake ice formation and melting processes in the three basins (western, central, and eastern, divided by the 82.4°W and 80.4°W meridians, see Fig. 1) were investigated separately. Western lake ice formed in mid December and grew rapidly, reaching a peak of 85% coverage in early February (not shown). Although ice in the central and eastern basins formed slowly in early January, ice formed rapidly in mid-January, and reached the maximum at the same time in late January as the western basin. The high concentration persisted until late February. The decay of lake ice started from the western lake, progressed toward the central and the eastern lakes, consistent with

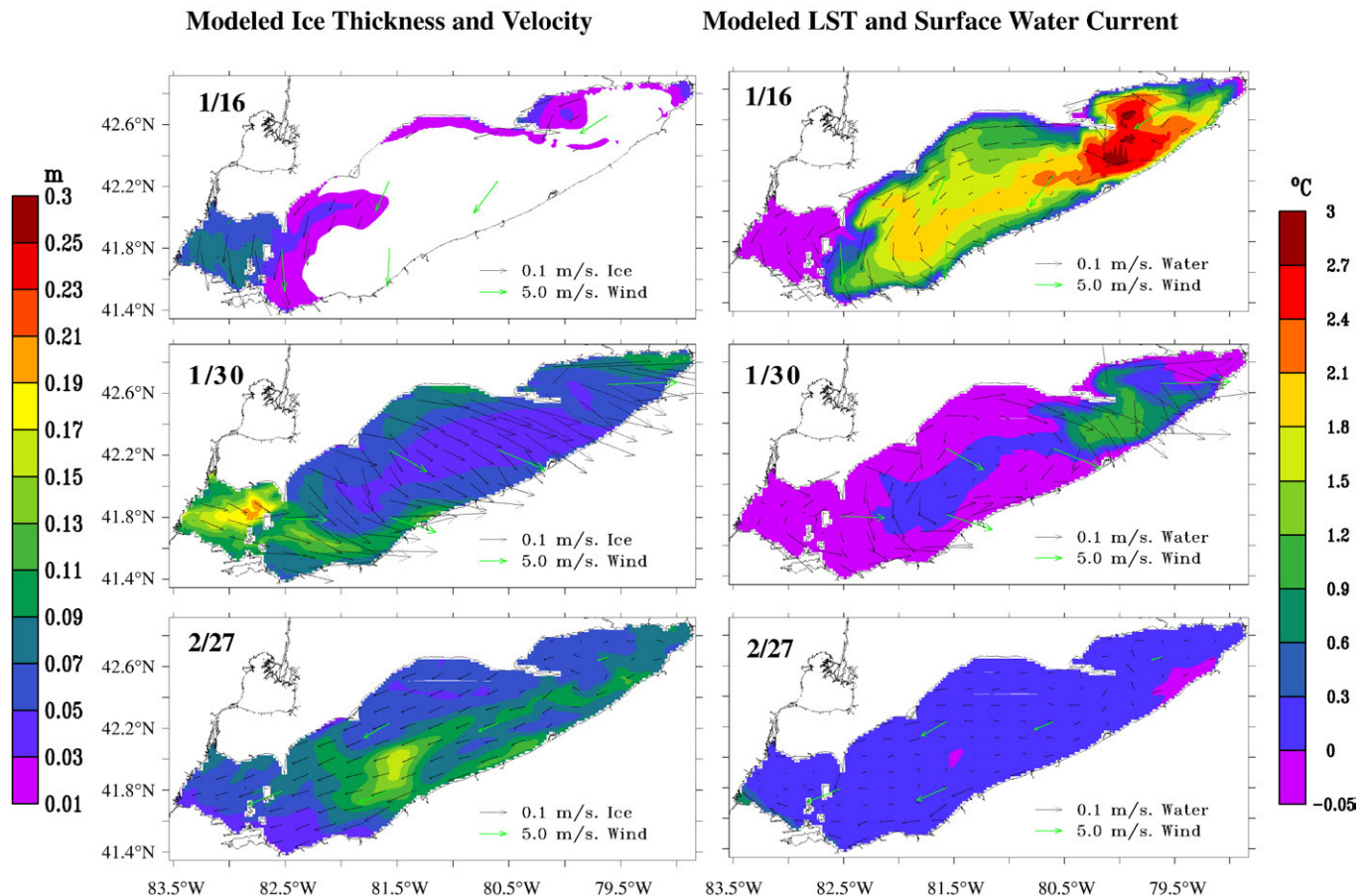


Fig. 6. (Left column): GLIM-simulated daily ice thickness in 2004 on January 16 and 30, and February 27. Superimposed are the ice velocity vectors (black) and wind velocity vectors (green). (Right column): GLIM-simulated daily lake surface temperature in 2004 on January 16 and 30, and February 27. Superimposed are the surface water velocity (black) and wind velocity vectors (green).

the spatial distribution. The basin-averaged ice thickness also shows ice thickness progression from the western to the eastern lake. The maximum ice thickness in the western basin was ~14 cm, while it was ~10.5 cm and 9 cm in the central and eastern basins, respectively.

To further examine the overall ice thickness distribution in Lake Erie, the thickness histograms were constructed using the December–March simulations (Table 2). In December, ~82% ice thickness was less than 3 cm, and ~16% ice thickness ranged from 3 to 6 cm. Only 2% of the ice thickness was over 6 cm. In January, the ice moved toward thicker ice categories. In February, thick ice categories (10–20 cm and 20–30 cm) appeared. The most frequent ice thickness ranged from 3 to 10 cm, totaling 60%. In March, thin ice dominated with 89% ice being less than 3 cm, similar to December. The total (December to March) ice thickness also shows that Lake Erie is thin ice dominant. These simulated ice thickness distributions should be validated by field measurements in the future.

A histogram based on GLIM-simulated ice speed was also constructed from December to March (Table 2, lower panel). In December, 98% of ice speed was less than 5 cm s⁻¹ due to the nature of the landfast ice that is attached to shore or formed in the shallow western lake (5–10 m depth). In January, high ice speed as large as 40–60 cm s⁻¹ appeared. In February, the most frequent ice speed ranged from 5–10 cm s⁻¹ (38%) to 10–20 cm s⁻¹ (32%). In March, lake ice moved back to low speed categories with <5 cm s⁻¹ being dominant (91%), due to the reduction in wind speed. The March dominant wind is southwesterly (see Fig. 2b), which advects lake ice toward the eastern lake with small wind fetch. The total ice speed ranged from <5 cm s⁻¹ (64%), 5 to 10 cm s⁻¹ (15.5%), 10 to 20 cm s⁻¹ (14.5%), 20 to 40 cm s⁻¹ (5%), and 40 to 60 cm s⁻¹ (1%).

Campbell et al. (1987) conducted a measurement study of ice drift using four ice buoys in central and eastern Lake Erie in the winter 1983/84, which provides some useful information for our model comparison. Three of the four buoys provided ice speed values to construct a histogram for January–March 1984 (Table 3, left column), which compares well to the modeled total ice speed histogram (Table 3, right), except for the low speed category (<5 cm s⁻¹), because December was dominated by low ice speed. Thus, when we re-construct the histogram using the simulated ice speed from January to March, 2004 only (Table 3, right column), the comparison is much better. This indicates that the GLIM reproduces the lake ice speed very well, although there are differences between the model simulations and the measurements due to year-to-year variability in the wind field.

To further investigate the dynamic responses of lake ice and surface water velocities, domain- and time-averaged wind, ice velocity, and water velocity vectors were constructed at ice-covered grids only from December 2003 to April 2004 (not shown). Ice drift

Table 3

The histograms for observed ice drift speed during January–March, 1984 (left column; taken from Campbell et al. (1987)) and GLIM-simulated ice speed during January–March 2004 (right). Units are in percentage.

Speed (m s ⁻¹)	Observed	Modeled
0–0.05	46	53
0.05–0.1	30	22
0.1–0.2	17	19
0.2–0.4	6	5
>0.4	1	1

direction is about 28° to the right of the wind vector, while the surface water vector is ~17° to the right of the ice drift. Therefore, the classical Ekman drift theory is also valid in an ice-covered lake, that is the surface water drift is ~45° to the right of the wind vector in the northern hemisphere. In terms of magnitude, ice and surface water velocities are about 1.9% and 1.7% of the wind speed, respectively. This is very close to the empirical threshold value of 2% in the ocean Lagrangian drift simulations using surface wind forcing.

LST temporal and spatial variability

To demonstrate GLIM's capability to simulate the seasonal cycle of LST with no data assimilation, bi-monthly spatial distribution of LST in 2004 was constructed (Fig. 7, left column). The LST map indicates heat storage was not completely depleted in January, because the LST in the deep basin was greater than 0 °C. However, February was the month with the lowest LST (not shown), being near the freezing point, indicating possible depletion of heat storage in the upper layer of the lake. Warming gradually started in March in the western lake, followed by a rapid warming from April to July. LST in August also remained warm, but started to cool down in September. Rapid cooling occurred from October to December. In December, the cooling is from west (shallow) to east (deep) and from coast to basin, with visible high LST in the deep basin.

To validate the model performance, the same bi-monthly mean LST maps (Fig. 7, right column) were derived from the AVHRR data. The Great Lakes Surface Environmental Analysis (GLSEA) charts are daily cloud-free composite surface temperature charts that are derived from daily geometrically corrected and cloud masked AVHRR images. All valid overwater temperatures for the day are combined with previous composite charts to produce overwater temperature at each pixel position. Digital ice concentration values derived from the National Ice Center Great Lakes Ice Analysis charts are then overlaid on the composite chart. The AVHRR-measured bi-monthly LST compares favorably to the modeled LST (Fig. 7) in general in other seasons, except for late spring. For example, in May, the measured LST is 11–14 °C, while the modeled LST is 12–15 °C. The difference is ~1.5 °C.

The temporal variation between the measured and modeled LST (Fig. 8) shows that the seasonal cycle is reasonably well reproduced by the model. The modeled LST captures short-term synoptic variability, while the AVHRR does not. The reason is that the model is forced by hourly atmospheric forcing, while the AVHRR measurement is the average composite maps using daily cloud-free images only. A systematic error occurred in both spring and autumn with the largest discrepancy in May. The most likely factor is heavy cloud cover in spring during the rapid warming season. The AVHRR measurements likely underestimate the LST because the composite average spans a 20-day window. As spring progressed, the cloud cover becomes lighter, and the LST becomes warmer. Thus, heavily-used images in the earlier period of the window (spring) with cloud-free conditions are heavily weighted towards the colder conditions in early spring, leading to the underestimate of LST. Similarly, the AVHRR measurements likely overestimate the LST in autumn also because the composite average spans a 20-day window. As autumn progressed, the cloud cover becomes heavier, and the LST becomes colder. Thus,

Table 2

The histograms for GLIM-simulated ice thickness (upper panel) and ice speed (lower) in December, January, February, March, and the total in the 2003/04 ice season. Units are in percentage.

	Dec	Jan	Feb	Mar	Total
Thickness (cm)					
0–3	82	51	19	89	65
3–6	16	30	30	8	15
6–10	2	12	30	2	11
10–20	2	12	30	2	11
>20			1.5		
Speed (m s ⁻¹)					
0–0.05	98	49	19	91	64
0.05–0.1	2	19	38	8	15.5
0.1–0.2		22	32	1	14.5
0.2–0.4		8	9		5
>0.4		2	2		1

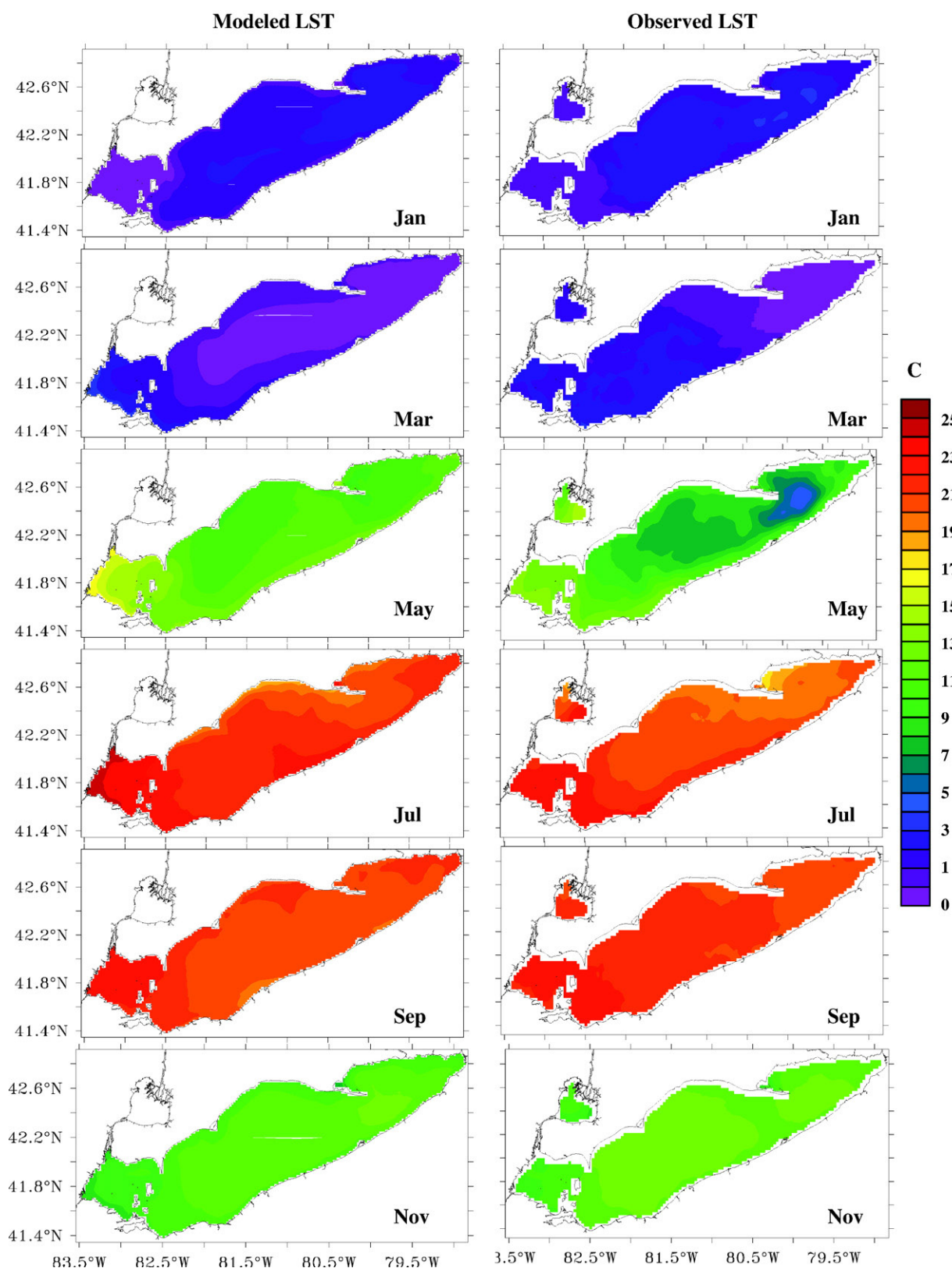


Fig. 7. The GLIM-simulated (left column) and AVHRR-measured (right column) monthly average LST in January, March, May, July, September, and November in 2004.

heavily-used images in the earlier period of the window (autumn) with cloud-free conditions are heavily weighted towards the warmer conditions of early autumn, leading to the overestimate of LST (see Fig. 8). The second possible reason for this may result mainly from a strong stratification due to insufficient surface wind-wave mixing, which leads to a warmer upper mixed layer due to a strong thermal

structure (Hu and Wang, 2010). The weak vertical thermal structure caused by mixing of the surface waves can allow more heat to penetrate into the lower layer of the water column, leading to the lower LST. Thus, the mixing caused by surface wind waves should be taken into account for better understanding of the thermal structure of the lake.

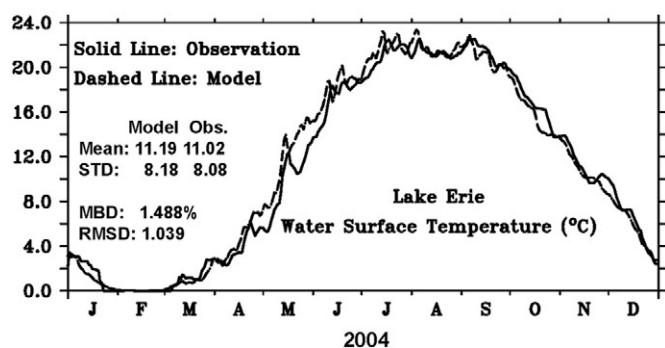


Fig. 8. Temporal (daily) comparison of the domain-averaged LST between the GLIM simulation (dashed) and AVHRR measurement (solid) in 2004 (units: °C). The simulated and observed means and standard deviations, and MBD and RMSD were provided.

Sensitivity studies: impacts of ice cover on water temperature and level

To confirm that lake ice cover affects the seasonal cycle of water temperature, we conducted a sensitivity study from the same initial conditions on December 8, 2003. Fig. 9 shows the comparison of basin-averaged water temperature with no ice (green) and with ice (red). Without ice, the water temperature cools faster than with ice, starting at day 30 and lasting until day 60, indicating that lake ice acts as an insulator. Without ice, water temperature increases faster than with ice due to an underestimate of surface albedo to incoming solar radiation, which is used to heat up the water, rather than to melt the ice. In the no ice case, an empirical constraint that water temperature is not allowed to go below zero must be applied. Nevertheless, the heat, which should be used to melt the ice, is incorrectly used to warm the water. Three months later, on February 8, 2004, the difference is more than 1.2 °C. At the end of five months (April 8), the difference is about 1 °C. Therefore, a lake ice model is essential for hydrodynamics and ecosystem modeling in the Great Lakes.

Lake ice not only affects the seasonal cycle of water temperature, but also wave and lake water level variations. In previous modeling studies using ocean-only models and in the GLCFS, empirical methods have to be used to dampen wave heights and water levels based only on observed ice charts. Fig. 10 shows the comparison of water levels between observations (black) and model simulations from the GLIM with ice (red) and with no ice (green) at Buffalo (Fig. 10b) and Toledo (Fig. 10d). When ice (Fig. 10a, black) forms, the water temperature (red) drops to freezing, and stays as long as lake ice concentration is

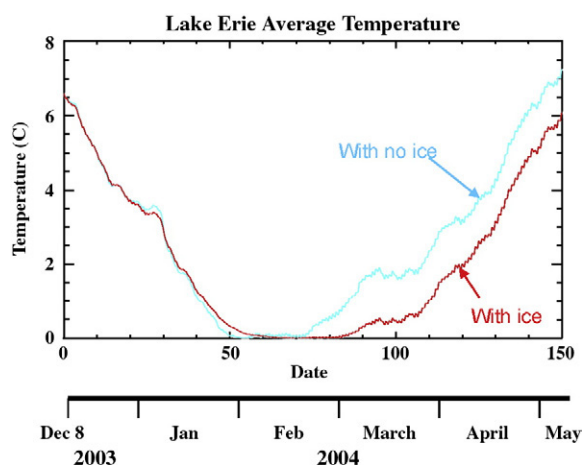


Fig. 9. The GLIM-simulated domain-averaged water temperature with ice (red) and with no ice (green). The start date is December 8, 2003.

over 70%. However, without lake ice, the water temperature (green, Fig. 10a) increases to above freezing, and drastically increases to more than 1 °C by day 60. The comparison between the observed water level in Buffalo (Fig. 10b, black) and modeled water level (red) is reasonably good; however, the modeled water level without ice (green) is overestimated. Another example at Toledo (Figs. 10c and d) further confirms that lake ice not only modulates water temperature, but also dampens the water level including waves and seiches. In other words, without a lake ice model, the water temperature, water level, and wave heights cannot be accurately reproduced and predicted for the winter season in the Great Lakes. The existence of lake ice, even thin, is very important for the protection of fisheries habitats and shoreline from erosion caused by waves and seiches.

A process study: cold air outbreak vs. warming event

To investigate the rapid response of the GLIM to strong cold air outbreaks from the north and warming events due to a cyclone passage from the south, we chose a period from February 10 to 23, 2004, which covers both cooling and warming events (Figs. 2a and 11). On February 10, lake-averaged SAT was about 0.6–0.8 °C (Fig. 11) with a southwesterly wind, and gradually decreased to ~0.2 °C on February 11 (Fig. 11). Lake ice in the western lake started to break up and melt (Fig. 11). Within the next four days, SAT suddenly dropped to –3.6 °C on February 16 due to the northeasterly cold air outbreak (Fig. 11) with the eastern lake being 6–12 °C below freezing (Fig. 11). On February 17, the entire lake was ice-covered with ice concentration being >80% (Fig. 11). Within the next five days from February 16 to 21, an extreme warming event occurred with lake-averaged SAT rising from –3.6 °C to +2.8 °C (Fig. 11) due to the advection of warm air from the south by the southeasterly winds (Fig. 11). Lake ice drastically melted in the western lake with only 50–60% ice cover (Fig. 11).

The model simulation and the 3-day averaged (twice weekly products) ice concentration (Fig. 3a) compare reasonably well. So far, there are no daily average ice measurements for a detailed comparison. In summary, based on this case study, it is apparent that the GLIM responds sensitively and quickly to both the warming and cooling events in terms of ice melting and re-freezing processes in the middle of the winter. This indicates that the GLIM has potential for use in a nowcast/forecast system in the Great Lakes.

Summary and conclusions

The Great Lakes Ice-circulation Model (GLIM) was developed and applied to Lake Erie. An application to the 2003/04 ice season was conducted to test the model performance that was validated by satellite measurements and in situ observations. Both temporal and spatial variation of lake ice concentration, thickness, velocity, and LST were investigated in depth. The impacts of lake ice cover on LST and water level were examined using sensitivity experiments, which are also validated by in situ gauge station measurements. A process study including both extreme cold air outbreak and warming events on synoptic time scales was conducted to test GLIM's response and sensitivity to the fast-changing weather for a future application to forecasting lake ice. Based on the above investigations, major conclusions can be drawn as follows:

- 1) The ice model with simple viscous-plastic ice rheology with one-layer ice works reasonably well in Lake Erie because the maximum domain-averaged ice thickness is about 10 cm. Thus, a simple one-layer ice model (Hibler, 1979) is probably suitable for the Great Lakes.
- 2) The lake ice seasonal cycle was reproduced by GLIM using hourly (high-frequency) atmospheric forcing, and compares reasonably well with measurements. The MBD and RMSD are 7.4% and $1.84 \cdot 10^3 \text{ km}^2$, respectively. Lake ice reached its maximum at the

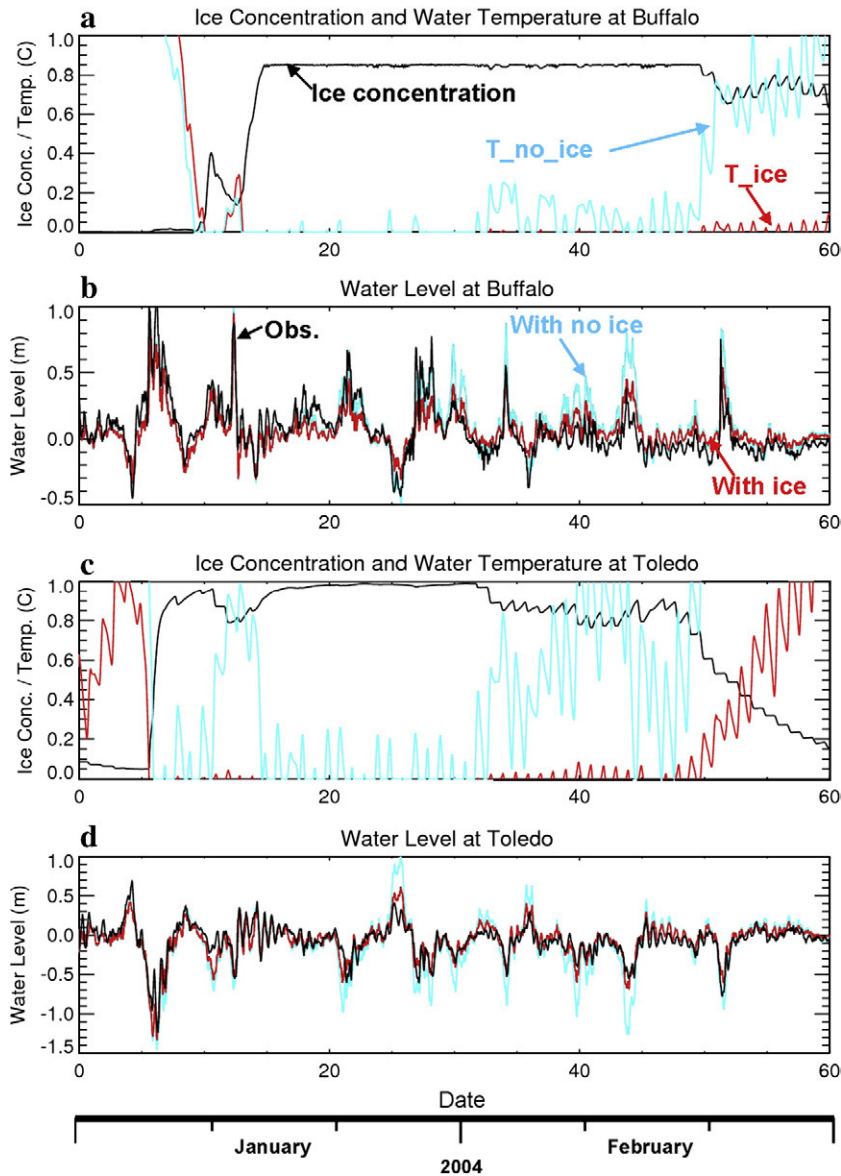


Fig. 10. a) The GLIM-simulated ice concentration (black), surface water temperature with ice (red), and with no ice (green) at Buffalo; b) GLIM-simulated water level variations with ice (red) and with no ice (green) with comparison to the measurement (black); c) same as a), except at Toledo; and d) same as b) except at Toledo. The start date is January 1, 2004. The means of the observed water levels were removed in the comparison.

end of January or early February with a domain-averaged thickness of 10.5 cm. The domain-averaged ice thickness from the model simulation is 7.5 cm on February 27, 2004, close to the observed ice thickness on February 27, 2008, with similar ice severity.

- 3) Based on measurement, landfast ice formed early in the shallow western lake and along the shallow coast. Then, ice expanded from the coast to the deep basin. The deepest part of the eastern basin is the last area for ice formation due to its larger water heat storage capacity. In late February, landfast ice breaks up first in the shallow western basin and along the coast. Then melting occurs from west to east. However, the GLIM did not capture the observed spatial breakup pattern of landfast ice in the western basin.
- 4) The GLIM also reproduces the seasonal cycle of LST, which compares well with the AVHRR-measured spatial and temporal LST. The MBD and RMSD are 1.5% and $\sim 1^\circ\text{C}$, respectively. The modeled LST spatial distribution is consistent with the lake ice

formation. However, a large discrepancy occurs in May, possibly due to cloud cover that leads to an underestimate of LST.

- 5) Sensitivity studies show that with no ice, domain-averaged water temperature can be overestimated by about 1.2°C . Furthermore, without ice cover, water level can be overestimated by over 0.5 m at Buffalo and Toledo.
- 6) The process study indicates that GLIM responds sensitively to extreme warming and cooling events in the synoptic weather patterns. In other words, in the middle of a winter, GLIM is capable of simulating lake ice breakup, re-freezing, and melting.

The weaknesses of the present version of GLIM include:

- 1) Wind stress applied to ice/water interface, i.e., ice–water stress, is not fully coupled. Thus further studies are needed to identify what causes the failure in the coupling in lake ice modeling. The empirical ice–water stress was applied since the conventional coupling between ice and water stresses failed. The possible reasons may be: (i) ice is too thin, strong non-linear effect in the

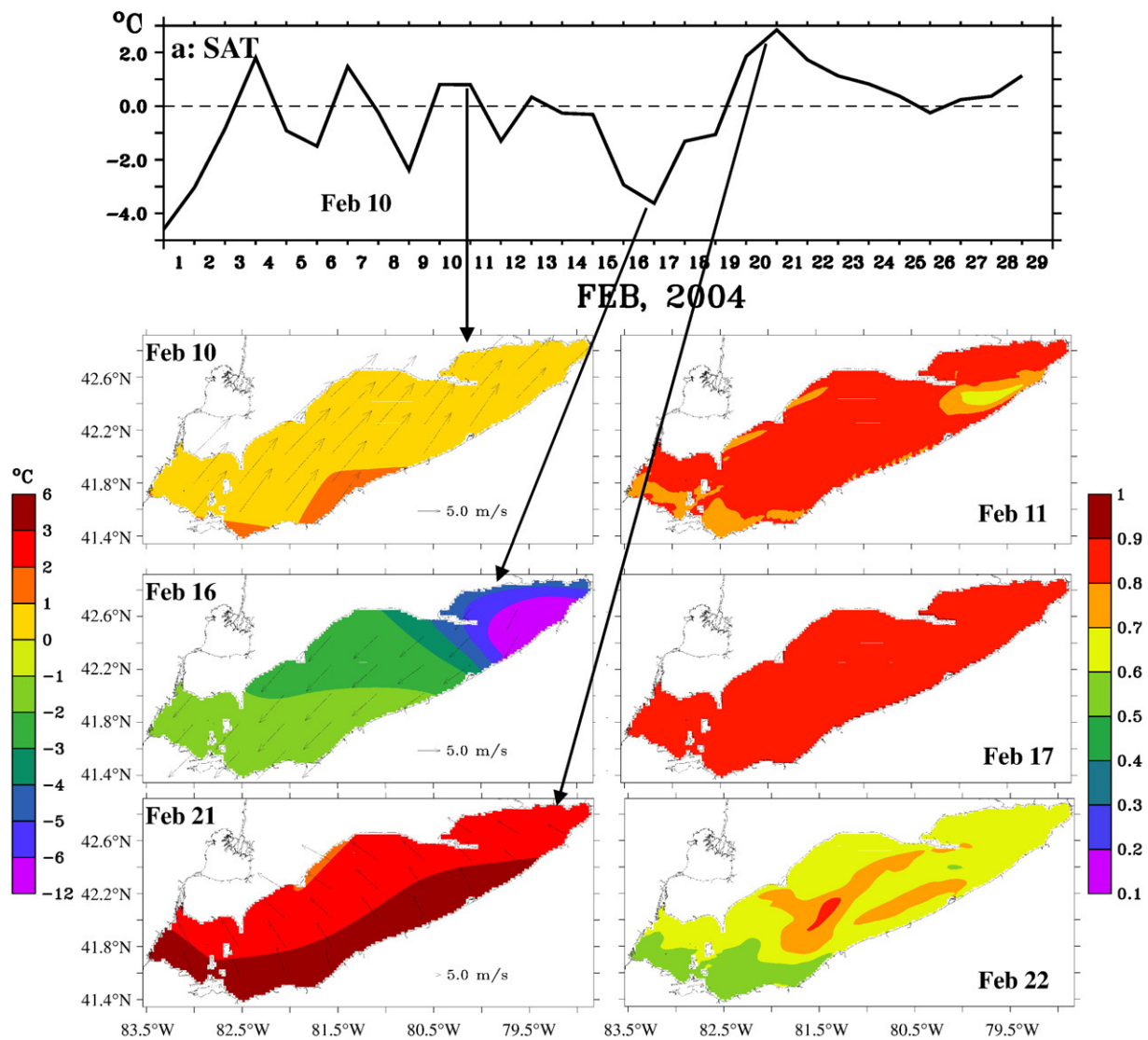


Fig. 11. Time series of SAT in February 2004 (a), spatial SAT and wind field on February 10, 16, and 21 (left column), and the corresponding spatial distribution of ice concentration on February 11, 17, and 22 (right column).

internal ice stress, which indicates the present isotropic model may not be suitable for such thin ice that behaves in an anisotropic manner, and (ii) it is speculated that in a lake with such thin ice (3–10 cm), the plastic-viscous (VP) rheology works well. An elastic feature may be needed. To answer these questions is beyond the scope of the GLIM. Therefore, research on anisotropic features and EPV (elastic-plastic-viscous) rheology of lake ice is needed.

- 2) Spatial patterns of the melting process were not well simulated, in particular, the landfast ice breakup process, when compared to its overall domain-averaged values and the freezing process.
- 3) There is a large error between the simulation of LST and satellite measurements in late spring, compared to other seasons, possibly due to: (i) cloud cover that leads to an underestimate of LST, and (ii) insufficient surface wind-wave mixing in GLIM that may lead to higher surface temperature due to stronger thermal stratification.

It should be pointed out that the GLIM was not fully validated due to lack of measured water current data, although the measured vertical temperature profile time series are being used for validation in another study. Therefore, further improvement

and validation of GLIM will be carried out for year-to-year simulations.

The importance of the development of GLIM has significant and broad impacts on interdisciplinary research in the Great Lakes. The further application of GLIM includes the update of the present GLCFS using GLIM to meet the increasing needs for winter ice forecast for safe navigation and rescue efforts, etc. Any ecosystem models should be coupled to an ice model such as GLIM in order to simulate the dynamics of lower trophic levels ecosystem (e.g., phytoplankton, and zooplankton) on both seasonal (yearly) cycle (over a winter) and interannual time scales. Coupling GLIM to a regional climate model is also an urgent task to accurately estimate the energy (heat and freshwater/moisture) budget over the Great Lakes for better understanding of long-term lake ice variability in response to a changing climate (Wang et al., 2010).

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