The Role of Ice Cover in Heavy Lake-Effect Snowstorms over the Great Lakes Basin as Simulated by RegCM4

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(Manuscript received 6 April 2012, in final form 27 June 2012)

ABSTRACT

A 20-km regional climate model, the Abdus Salam International Centre for Theoretical Physics Regional Climate Model version 4 (ICTP RegCM4), is employed to investigate heavy lake-effect snowfall (HLES) over the Great Lakes Basin and the role of ice cover in regulating these events. When coupled to a lake model and driven with atmospheric reanalysis data between 1976 and 2002, RegCM4 reproduces the major characteristics of HLES. The influence of lake ice cover on HLES is investigated through 10 case studies (2 per Great Lake), in which a simulated heavy lake-effect event is compared with a companion simulation having 100% ice cover imposed on one or all of the Great Lakes. These experiments quantify the impact of ice cover on downstream snowfall and demonstrate that Lake Superior has the strongest, most widespread influence on heavy snowfall and Lake Ontario the least. Ice cover strongly affects a wide range of atmospheric variables above and downstream of lakes during HLES, including snowfall, surface energy fluxes, wind speed, temperature, moisture, clouds, and air pressure. Averaged among the 10 events, complete ice coverage causes major reductions in lake-effect snowfall (>80%) and turbulent heat fluxes over the lakes (>90%), less low cloudiness, lower temperatures, and higher air pressure. Another important consequence is a consistent weakening (30%–40%) of lower-tropospheric winds over the lakes when completely frozen. This momentum reduction further decreases over-lake evaporation and weakens downstream wind convergence, thus mitigating lake-effect snowfall. This finding suggests a secondary, dynamical mechanism by which ice cover affects downstream snowfall during HLES events, in addition to the more widely recognized thermodynamic influence.

1. Introduction

Heavy lake-effect snowstorms are an important meteorological aspect of the Great Lakes region, producing a variety of hydrological and societal impacts (Eichenlaub 1979; Schmidlin 1993). The basic physical mechanisms responsible for heavy lake-effect snowfall (HLES) are well understood and involve the destabilization of a relatively cold, dry air mass by heat and moisture heat fluxes from a comparatively warm lake surface (e.g., Wiggin 1950; Eichenlaub 1970; Kristovich and Laird 1998). Of implicit importance for this process is the existence of open water, in order to generate the large surface–atmosphere temperature gradient needed for HLES development. Lake ice cover is therefore considered a key player in regulating lake-effect snowfall, as implied by the climatological snowfall maximum in the Great Lakes region during December and January (before the peak in ice extent) and as documented by the observed decline in surface sensible and latent heat fluxes as a function of ice cover (Braham and Dungey 1984; Niziol et al. 1995).

Despite the accepted importance of open water as a vital ingredient in HLES, a number of questions still surround the role of ice cover in affecting these phenomena. First, although ice is known to reduce HLES by suppressing sensible and latent heat loss from the lake surface, there have been no long-term quantitative assessments of its effect on downstream snowfall. Second, even though ice cover clearly affects these turbulent energy fluxes, its influence on the *dynamics* of lake-effect

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DOI: 10.1175/MWR-D-12-00107.1

snow is less obvious. For example, do changes in wind velocity induced by ice cover mitigate or enhance HLES? Third, while an impact from open water on snowfall immediately downstream of a particular Great Lake may be apparent, the regional influence of a given lake's ice cover on HLES over the entire basin is less obvious. For instance, by how much can open water over Lake Superior remotely affect precipitation as far away as the snowbelts of Lakes Huron or Ontario? A final motivation for this research is the expectation that a warming climate will lead to significantly less ice cover on the Great Lakes in the future. This change alone should promote more lake-effect snow, consistent with the increase in snowfall over the lake-effect zone of the Great Lakes during the twentieth century, a trend linked to warming lakes and reduced ice cover (Burnett et al. 2003; Kunkel et al. 2009). However, climate models also suggest a substantial decline in Arctic air masses over this region (Vavrus et al. 2006), favoring less lake-effect snowfall. Isolating the role of ice cover in HLES in the present climate serves as a step toward untangling this uncertain interplay.

Previous work on this topic has generally focused on a single HLES event spawned from a particular Great Lake or the general climatological impacts of the Great Lakes on lake-effect snow. A variety of process studies have investigated specific lake-effect storms, employing primitive equation models, mixed layer models, cloud-resolving models, and mesoscale models (Lavoie 1972; Ellenton and Danard 1979; Hjelmfelt 1990; Sousounis and Fritsch 1994; Maesaka et al. 2006). These studies revealed many of the important mechanisms for the generation and distribution of lake-effect snowfall, but they concentrated more on depth than breadth by focusing mainly on a single event. From a climatological perspective, Lofgren (1997) found that the inclusion of the Great Lakes in an atmospheric GCM caused much higher temperatures and much larger sensible and latent heat fluxes during early winter. The effect of the Great Lakes on precipitation is strongest during late autumn–winter and most important downstream of Lake Superior, where precipitation is double the expected amount without the lake, and least impactful downstream of Lakes Ontario and Erie (15%–20% lake enhancement; Scott and Huff 1996). In terms of dynamical effects, the Great Lakes have been identified as a preferred region for wintertime cyclogenesis (Petterssen and Calabrese 1959; Colucci 1976) and can produce mesoscale surface lows during cold-air outbreaks (Sousounis and Fritsch 1994) because of the presence of a relatively warm surface when open water is present.

The current study seeks to bridge these two time and space scales by investigating a collection of HLES case studies for all the Great Lakes and addressing both the local influence of individual lakes on lake-effect snowfall and the regional effect of lakes over the basin. We assume that this set of events represents fairly typical conditions associated with HLES and therefore that the conclusions drawn here also apply to other extreme lake-effect snowstorms over this region. To identify the importance of an open water source during these HLES events, our approach is to suppress all open water by artificially imposing 100% ice cover as a boundary condition to a regional climate model described in section 2. The meteorological impact of these all-ice simulations is described in section 3 in terms of not only snowfall but other related variables such as wind, moisture, temperature, clouds, sea level pressure, and turbulent heat fluxes. An interpretation of these results and a discussion of their representativeness are given in section 4, followed by a summary and suggestions for future work in section 5.

2. Data and methods

a. *Model description and forcings*

In this study we employ the Abdus Salam International Centre for Theoretical Physics Regional Climate Model Version 4 (ICTP RegCM4). A thorough description of the model is given in Pal et al. (2007) and Elguindi et al. (2011), while an evaluation of its simulated lake-effect snowfall characteristics from a companion study is presented in Notaro et al. (2013, hereafter NZV). The atmospheric component of RegCM4 consists of a dynamical core, based on the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5; Grell et al. 1994), and a radiative transfer scheme based on the NCAR Community Climate Model version 3 (CCM3; Kiehl et al. 1996). RegCM4 is a compressible, finite-difference model that is constrained to hydrostatic balance and uses sigma coordinates in the vertical. In this study we run the model at 20-km horizontal resolution and 18 sigma levels over the eastern United States and southeastern Canada (Fig. 1), allowing sufficient representation of the Great Lakes and nearby topographic features. Boundary layer processes are represented by a nonlocal vertical diffusion scheme of Holtslag et al. (1990), while large-scale precipitation and nonconvective clouds are based on a subgrid explicit moisture scheme (SUBEX; Pal et al. 2007). Convective precipitation is parameterized as in Grell (1993), which our analysis deemed preferable to the Kuo scheme (Anthes 1977) over this domain. The phase of cloud condensate is temperature

FIG. 1. Simulation domain with shading for elevation (m) and small dots for the 20-km grid. The inner domain, within the buffer zone, is shown with the thick black box. The pink boxes delineate the primary regions of heavy lake-effect snowfall and are used in this study to compute areally averaged quantities (adapted from NZV).

dependent, as in CCM3 (Kiehl et al. 1996). Cloud water is represented as all ice particles where the ambient temperature is less than −30°C, all liquid droplets above −10°C, and as mixed phase with a linear transition from ice to liquid between −30° and −10°C. Land surface processes are treated by the Biosphere–Atmosphere Transfer Scheme (BATS; Dickinson et al. 1986, 1993) using three soil layers and 20 land cover–vegetation categories, whose surface conditions affect the exchange of energy, momentum, and water vapor with the atmosphere.

RegCM4 is coupled to the one-dimensional energy-balance lake model of Hostetler and Bartlein (1990), which represents the vertical exchange of heat within a lake column via eddy diffusion and convective mixing. The lake model includes a thermodynamic ice parameterization (Patterson and Hamblin 1988) that allows overlying snow cover but no fractional ice cover within a grid box, meaning that every lake cell is either completely ice covered or ice free. Neither the lake model nor ice module treat horizontal heat transfer within the lake or vertical heat exchange with the lake bottom. RegCM4 provides air temperature, moisture, wind speed, surface radiation fluxes, and snowfall to the lake model, which then computes a vertical temperature profile and returns the lake surface (or ice) temperature and ice cover as boundary conditions for the atmosphere. Surface turbulent fluxes of heat and moisture are determined from bulk aerodynamic formulas from BATS. A 1-m vertical lake resolution is used in conjunction with prescribed, spatially varying lake depths based on bathymetric observations.

We drove RegCM4 continuously from May 1975 to December 2002, using initial and lateral boundary conditions from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis (Kalnay et al. 1996) and the Global Sea Ice and Sea Surface Temperature dataset (GISST) from the Met Office (Rayner et al. 1996). The lateral boundary conditions are 6-hourly on a $2.5° \times 2.5°$ grid and transition to the interior domain solution within a 15-gridcell buffer zone using a linear relaxation scheme. We compare the simulated lake ice with observations of ice coverage from the National Oceanic and Atmospheric Administration (NOAA) Great Lakes Ice Atlas (Assel 2003), which provides measurements only through 2002 based on data from the National Ice Center and Canadian Ice Service. Snowfall observations are obtained from the U.S. High Resolution Cooperative Dataset through the National Climatic Data Center (NCDC) and from Environment Canada. After initializing the model for 8 months to spin up the soil moisture and lake temperatures, a continuous solution is generated for the entire 1976–2002 interval to produce 26 autumn–winter periods in our analysis (1976/77 to 2001/02).

b. Model performance

The simulation of lake-effect snowfall and ice cover during this time period is described in detail by NZV. In the current model version, RegCM4 captures the major features of Great Lakes ice and snowfall with respect to climatological means, interannual variability, and spatial distribution. Yearly variations in basinwide ice coverage in the model correlate with observations at 0.91 during these 26 winters, and RegCM4 reproduces maximum snowfall amounts in lake-effect zones downstream of each of the Great Lakes. The seasonal timing of snowfall downstream of each lake is also realistic, exhibiting a December–January maximum ahead of peak ice coverage in February. The simulated interannual variation in downstream snowfall is credible, as correlations with station observations range from 0.72 (Huron) to 0.87 (Michigan). The major model biases of relevance for this study are that ice forms too early in the season along the periphery of the lakes and that deep lake points do not develop enough ice cover. Both of these shortcomings probably stem from the absence of horizontal mixing in the lake model. Simulated seasonal snowfall downwind of Lakes Michigan, Superior, Huron, and Erie are biased low by 19%, 10%, 8%, and 8%, respectively, whereas the model produces 46% more snowfall than observed downstream of Lake Ontario, apparently due to a strong sensitivity to topographically forced ascent from the Adirondack Mountains immediately east of the lake. Similarly, Wilson (1977) reported that the roughly 25% increase in local precipitation attributable to Lake Ontario rises to over 50% in the high terrain to the east.

NZV also proposed a novel definition of heavy lake-effect snowstorms, based on a five-point criteria involving snowfall amount at a location (daily total $\geq$10 cm), proximity to a Great Lake shoreline (within 100 km), the prevailing wind direction (surface wind flow off a Great Lake for at least 6 h), the percentage of upstream open water (at least 30%), and the difference between snowfall amount inside versus outside of the lake-effect snow belt (at least 4 cm). The model simulates heavy lake-effect snowfall most frequently downstream of Lake Superior and secondarily in the elevated region east of Lake Ontario. Ice cover is found to exert a strong influence on heavy lake-effect snowstorms, as their simulated frequency declines sharply on all the lakes from a peak in December–January to February, when ice concentration reaches a maximum. RegCM4 was also shown to generate realistic synoptic patterns of sea level pressure and associated surface winds during heavy lake-effect events.

c. *Experimental design*

In this study, we subsetted 10 of these heavy lake-effect snowstorms (2 for each Great Lake) to investigate the influence of open water on these events. We chose storms that produced especially heavy, widespread snowfall and that the model accurately simulated in comparison with observations. Over the 27-yr simulation, the total number of HLES at a representative point downstream of each lake ranged from a minimum of 26 for Lake Michigan to a maximum of 94 for Superior (NZV). Each event in our analysis is shown in Table 1, which lists the prognostically generated ice concentration on each lake on the date of the HLES. The simulated ice coverage accompanying these events varies greatly, ranging from completely open conditions during early season storms on Lake Erie to a majority of ice cover during one of the Lake Huron storms (in this study, we include Georgian Bay as part of Lake Huron). To test the influence of open water on the characteristics of HLES during these cases, we reran the model starting 1 day prior to each event with an imposed 1-m-thick ice over the entire lake surface (all other boundary conditions were identical to those in the control run). The model was then run for 2 days to identify the effect of the imposed ice cover during the peak snowfall in the control simulation. This approach allowed sufficient time for the model to react to the altered boundary conditions but not long enough to deviate substantially from the synoptic conditions produced in the corresponding control simulation (CONTROL). For each of the 10 case studies, we conducted two experiments. In one experiment, complete ice cover was imposed only on a single lake [individual-lake simulations (IL)], whereas in the all-lake simulations (AL) we prescribe every Great Lake to be completely iced over. This procedure allows us to separate the impact of ice cover locally for each lake (IL) versus remotely over the entire basin (AL) as a result of heat and moisture advection that could cause one lake to influence the lake-effect snowfall of another.

TABLE 1. Dates of the 10 heavy lake-effect snowstorms analyzed (two events per lake). The model-simulated ice coverage on each day is shown next to the observed value (in parentheses) from the NOAA Great Lakes Atlas (Assel 2003). The simulated values represent area averages over the entire lake, whose individual grid cells are represented as either completely ice covered or totally ice free. The observed values are also lakewide area averages, but the fractional ice coverage within each grid cell of nominal 2.5-km resolution is accounted for.

Lake	Date	% ice cover
Superior	7 Dec 1995	16 (0)
	11 Dec 1995	20 (1)
Huron	29 Dec 1985	56 (7)
	26 Dec 1990	33 (37)
Ontario	12 Dec 1995	27 (0)
	18 Jan 1992	39 (3)
Michigan	4 Dec 1991	11 (0)
	5 Jan 1988	27 (12)
Erie	12 Nov 1996	0 (0)
	22 Nov 2000	0 (0)

3. Results

a. *Control simulations*

The synoptic setting on each of the 10 selected HLES days in the CONTROL simulation is shown in Fig. 2 and is very similar to the corresponding maps from the North American Regional Reanalysis (not shown). In every case a midcontinental polar anticyclone promotes a strong northwesterly flow over the Great Lakes (surface wind speeds >8 m s^{-1} above the lakes), allowing cold air to be transported over the partially or totally ice-free lake waters. As expected from these circulation patterns during late autumn–early winter, CONTROL produces relatively high snowfall downstream of each lake (Fig. 3, left panel). The peak amounts occur within favored lake-effect snowfall zones encompassed by black boxes, which are used later in this study to diagnose the downstream response of snowfall and other meteorological variables. Some of the events also exhibit significant

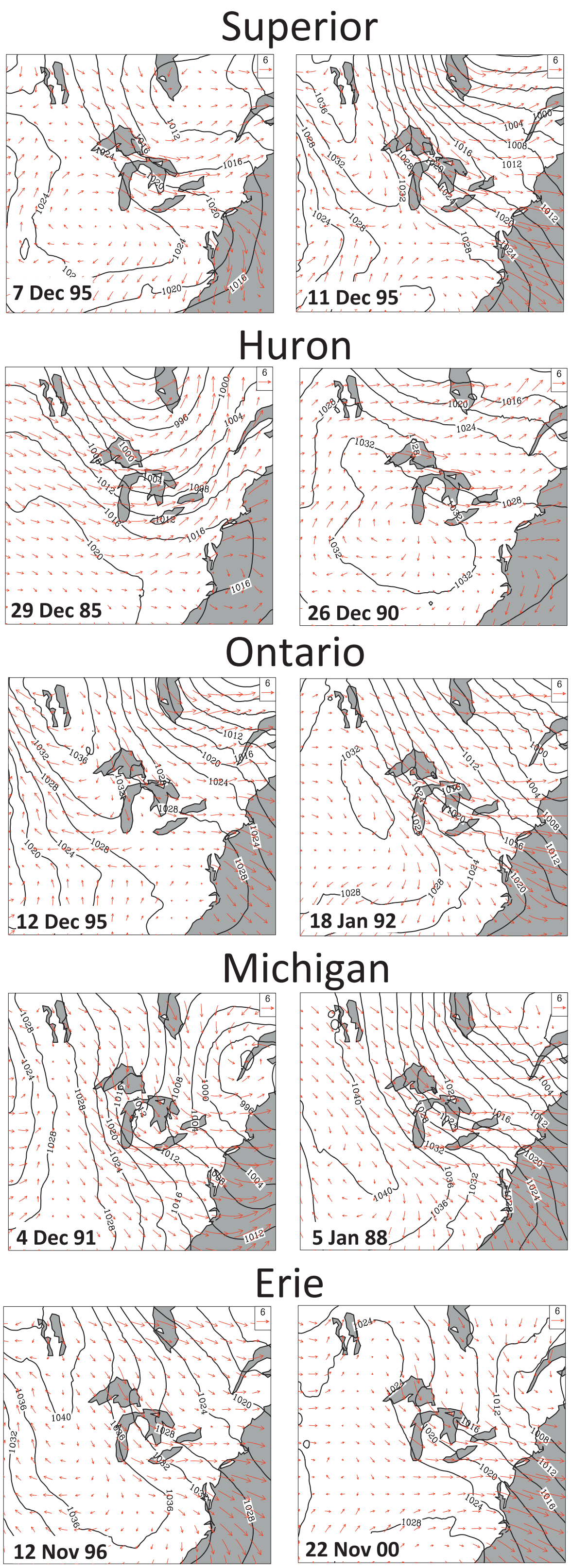

FIG. 2. Simulated sea level pressure (hPa) and 10-m wind velocities (m s^{-1}) on each of the 10 days with heavy lake-effect snowfall in the CONTROL simulation.

over-lake snowfall, especially those for Superior and Michigan. The areally averaged snowfall accumulations within the diagnostic boxes are slightly too low for the average of the 10 cases (7.5 cm simulated vs 8.2 cm observed), but within 10% of measured values. This average simulated value among all the lakes includes lakes with excessive simulated snowfall [e.g., Ontario (+29%) and Superior (+15%)] and those with totals less than observed [e.g., Erie (−48%), Michigan (−36%), and Huron (−25%)] on these HLES days. These errors during heavy events are generally consistent with the climatological snowfall biases in RegCM4, particularly the overproduction of snow downwind of Lake Ontario and the underproduction downstream of Lakes Erie, Michigan, and Huron (NZV). Among the 10 HLES days examined here, the maximum accumulation of up to 30 cm occurs just downstream of Lake Ontario, whose lake-effect snowfall pattern is usually more spatially concentrated than the other lakes, due to enhancement by local topography (Niziol et al. 1995).

b. *Individual-lake simulations*

When complete ice coverage is imposed one lake at a time (experiment IL), downstream snowfall is dramatically curtailed (Fig. 3, right panel). The reduction is especially pronounced for both Superior cases and the second Michigan case, in which lake-effect snowfall essentially disappears. Conversely, the HLES downstream of Lakes Erie and Ontario is mitigated but not eliminated by upstream ice cover (73% snowfall reduction for Erie and 80% reduction for Ontario, averaged between their two cases). Overall, the suppression of open water on the individual lakes causes over an 80% decline in downstream HLES, averaged among the 10 cases (see Fig. 4 and Table 2).

The change in ice cover in IL from the amount simulated by RegCM4 in CONTROL varies greatly from one lake to another. As shown in Table 1, the early season HLES events around Lake Erie occur before any ice has formed, whereas the model produces ice over a majority of Lake Huron in its 29 December 1985 storm. As can be inferred from a comparison with observations during these 10 events (Table 1), lake ice in RegCM4 is generally too extensive during the early to midwinter, when HLES is most common. Averaged over the 10 cases, the prognostic ice concentration on these days is 23% compared with 6% reported in the NOAA Great Lakes Ice Atlas (Assel 2003). The possible impact of this bias is addressed in section 4.

The most direct explanation for the severe decrease in lake-effect snowfall in experiment IL is that complete ice coverage chokes off the primary source of heat and moisture from the lake surfaces. This expectation is

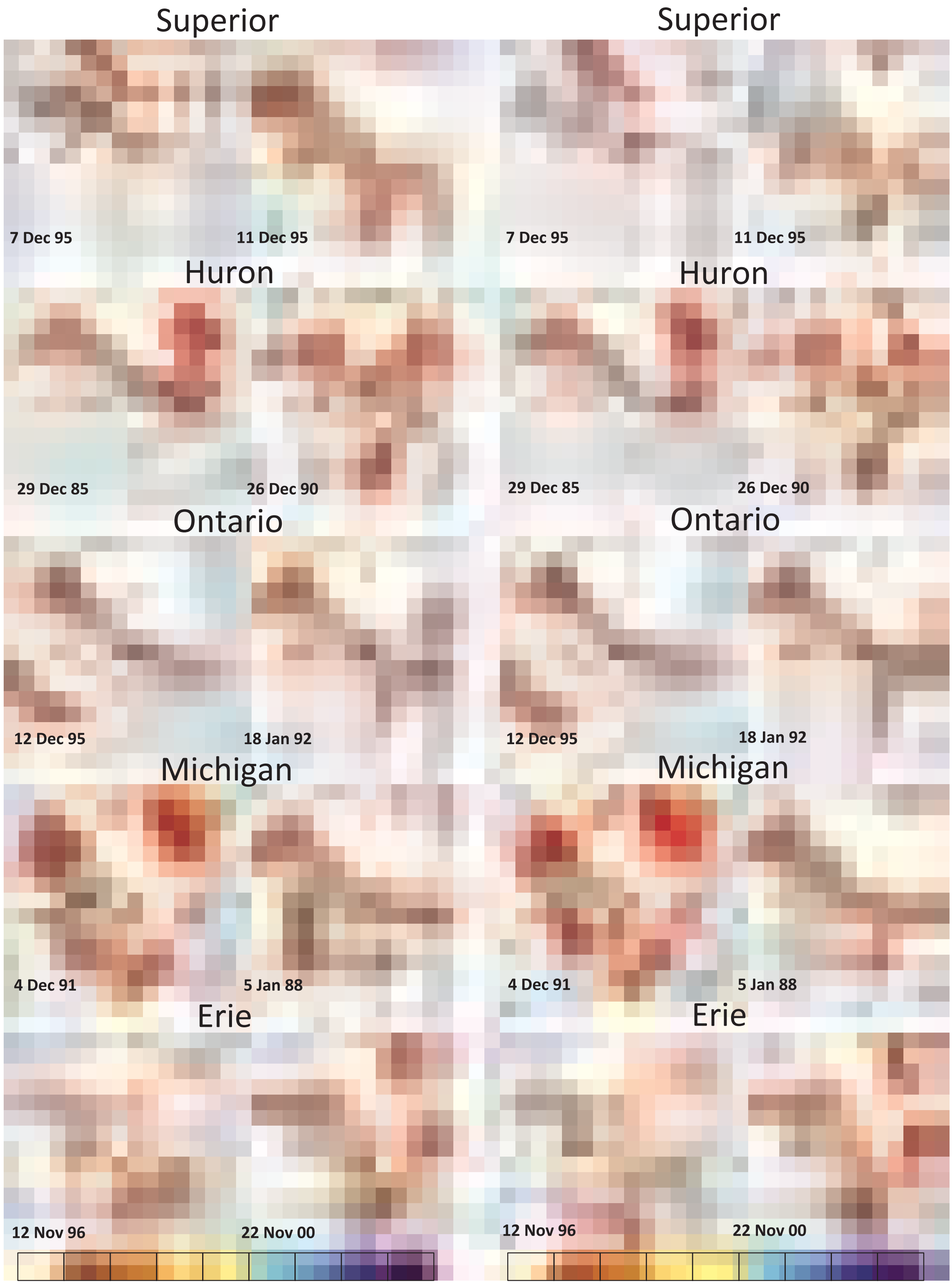

FIG. 3. Simulated daily snowfall (cm) on each of the 10 days with heavy lake-effect snowfall for the (left) CONTROL simulation and (right) IL experiment. The lake-effect region for each lake is denoted by a black box.

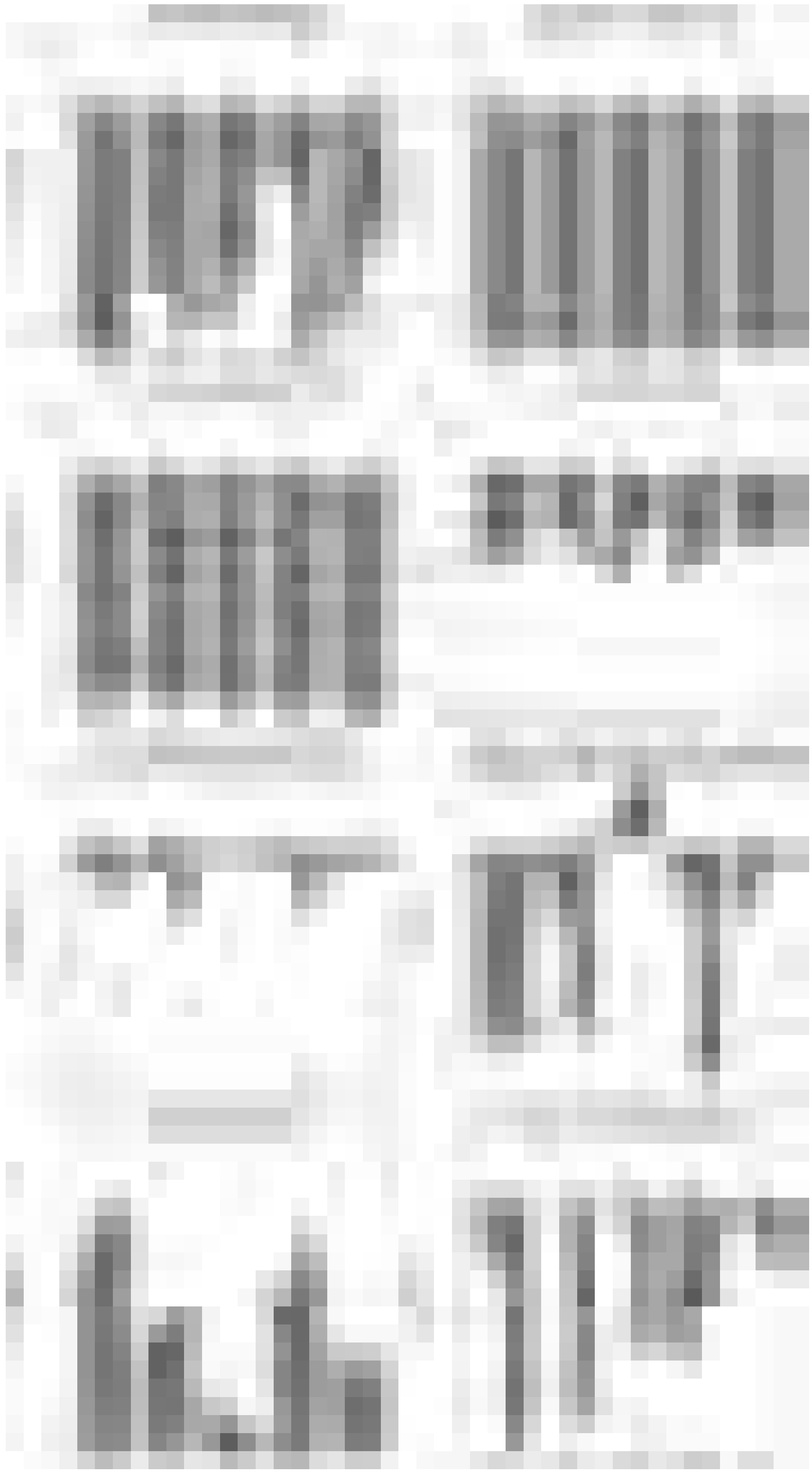

FIG. 4. Changes in downstream snowfall, precipitable water, low cloud fraction, SLP, surface air temperature and over-lake latent heat flux, sensible heat flux, and surface wind speed between experiment IL and CONTROL in all 10 cases. The changes (IL minus CONTROL) are area averages over either the entire lake or the diagnostic box adjacent to each lake and are expressed as percentages, except for SLP (hPa) and temperature (K).

TABLE 2. Average magnitude of variables among the 10 HLES events in the CONTROL (CTL), individual-lake (IL), and all-lake (AL) experiments (EXP). Statistically significant changes (95% confidence level) between IL and CTL and between AL and CTL are italicized. The significance of changes between IL and AL are indicated in the final column. Latent and sensible heat fluxes and wind speed are area averages over lake points, whereas the remaining variables are shown as area averages over the diagnostic boxes downstream of the lakes, as shown in Fig. 3.

	IL			AL		
Variable	CTL	IL EXP	IL EXP − CTL	AL EXP	AL EXP − CTL	AL − IL Significant?
Snowfall (cm)	7.46	1.17	*−6.29 (−84%)*	0.87	*−6.59 (−88%)*	No
Latent heat (W m^{-2})	160.2	1.9	*−158.3 (−99%)*	1.7	*−158.5 (−99%)*	No
Sensible heat (W m^{-2})	252.3	14.2	*−238.1 (−94%)*	16.9	*−235.4 (−93%)*	No
SLP (hPa)	1021.9	1023.2	*1.3*	1024.7	*2.8*	Yes
Wind speed (m s^{-1})	10.5	7.1	*−3.4 (−33%)*	6.7	*−3.8 (−36%)*	No
2-m air temperature (°C)	−9.3	−12.4	*−3.1*	−14.3	*−5.0*	No
Precipitable water (kg m^{-2})	3.54	3.15	*−0.39 (−11%)*	2.78	*−0.76 (−21%)*	Yes
Low cloud fraction	0.30	0.19	*−0.11 (−34%)*	0.14	*−0.15 (−51%)*	No

borne out by the huge reduction in simulated latent heat flux averaged over each lake, which approaches 100% in all 10 cases (Fig. 4). A similarly robust, but not quite as extreme, decline in sensible heat flux is also evident, as the relatively warm open-water portion of each lake is replaced with a much colder ice surface. These percentage decreases translate into a tremendous change in the absolute amount of energy transfer, considering that the over-lake latent and sensible heat fluxes average 160 and 250 W m^{-2}, respectively, among the 10 cases in CONTROL and exceed 210 (latent) and 440 W m^{-2} (sensible) over Lake Superior during the 11 December 1995 event.

Besides suppressing surface energy transfer, the prescribed switch to complete ice cover also induces significant changes in the local atmospheric circulation around each lake. For example, a remarkably consistent and important impact is a reduction in near-surface wind speed over every lake, both as a percentage change (Fig. 4) and an absolute change (Fig. 5). The surface wind weakens significantly in every case in experiment IL and by a rather similar amount. The lake-averaged decreases among the 10 events range only between 2.6–4.3 m s^{-1} (24%–44%), and all lakes show regions with reductions of at least 5 m s^{-1}.

In addition to this circulation response, other changes downstream of the lakes induced by the removal of open water include decreasing atmospheric moisture (precipitable water and near-surface specific humidity), low cloud amount, and near-surface air temperature, as well as increasing SLP (Fig. 4). Averaged among the 10 events, all of these quantities show statistically significant changes (95% confidence) between IL and CONTROL within the diagnostic downstream boxes, as do the variables discussed above (downstream snowfall and over-lake turbulent heat fluxes and wind speed). Precipitable water (PW) declines in all 10 cases (mean reduction of 11%), presumably due to the severely reduced upstream moisture supply over the fully ice-covered lakes. A similar explanation probably applies to low clouds, which show a much more varied response than PW, but a larger percentage decline in the mean (34%). The changes in low cloud amount and PW across the 10 cases are very highly correlated (0.86). In addition, the icier lake surfaces in IL cause a significant cooling downstream (mean reduction of 3.2 K), although a couple of cases produce a slight warming. A robust increase in air pressure is also evident, ranging from a modest rise downstream of Lake Ontario (<0.5 hPa) to the strongest response near Lake Superior (>2 hPa). The spatial extent of the pressure rise varies dramatically by lake (Fig. 6). Not only does the addition of ice cover in IL induce the largest (smallest) SLP increase just downstream of Lake Superior (Ontario), but the spatial extent of these pressure changes differs accordingly. Increases of over 2 hPa are seen directly over Lake Superior but also far downwind, spreading over Lake Huron and part of Lake Michigan. Similar but muted behavior is apparent for Lake Michigan's two IL events. By contrast, imposing complete ice cover on Lake Ontario produces only a relatively localized pressure change that is confined almost exclusively to the lake itself and New York State, while not affecting any other Great Lake.

As displayed in the vertical–longitude cross sections in Fig. 7 (left), the weakening of near-surface winds over the lakes in IL generally coincides with stronger winds aloft and an opposite vertical dipole pattern of wind changes downstream over land, where winds are stronger at low levels but weaker aloft. The corresponding cross sections for the change in vertical motion in Fig. 7 (right) illustrate that these wind anomalies are part of a circulation cell in the lower troposphere caused by the imposition of complete ice cover. In every case, anomalous

FIG. 5. The change in surface wind speed (m s^{-1}) between IL minus CONTROL in all 10 cases.

FIG. 6. Simulated change in daily mean sea level pressure (hPa) in experiment IL minus CONTROL for the 10 HLES events.

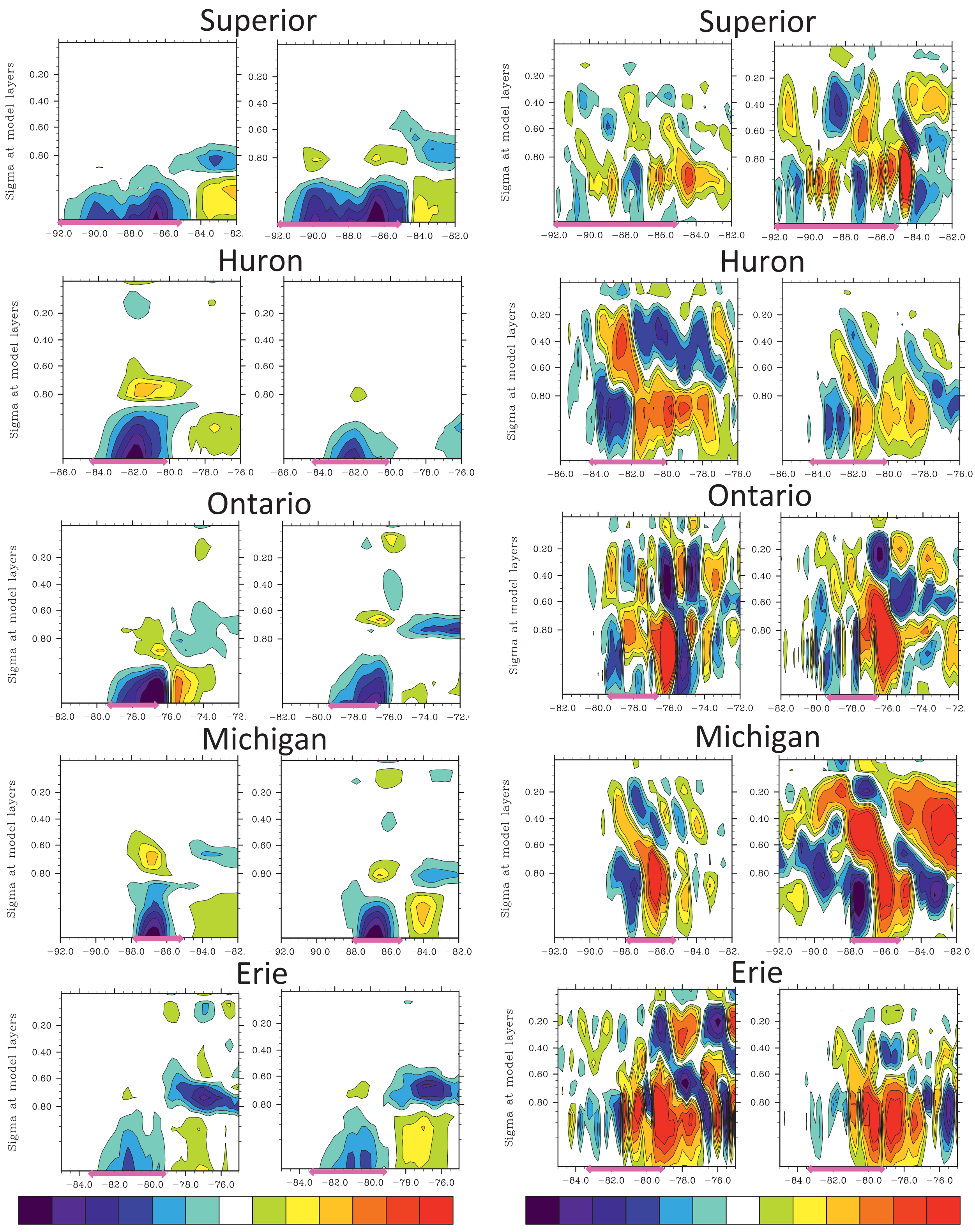

FIG. 7. Vertical–longitude cross sections of the change between IL minus CONTROL in (left) wind speed (m s^{-1}) and (right) vertical motion (hPa s^{-1}) in the 10 cases, using the same order as in Fig. 3. The approximate width of each lake is indicated by the pink arrows above the longitudes on the *x* axis. The data are averaged over the following latitudes: Superior 45.9°–49.2°N, Michigan 41.5°–46.2°N, Huron 43.0°–46.5°N, Erie 41.2°–43.3°N, and Ontario 43.1°–44.4°N.

descent occurs over the eastern portion of the lake and downstream over land in IL, whereas anomalous ascent is produced over the western portion of the lakes. Because the over-lake winds during all events in CONTROL are westerly to northwesterly (Fig. 2), the sign of the wind speed anomalies with height in Fig. 7 implies that this vertical motion couplet comprises a circulation cell with an anomalous easterly wind component just above the lake and a strengthened westerly component aloft. In each case, the largest surface-based wind reductions are bracketed to the east by particularly pronounced anomalous sinking in the lowest layers and to the west by anomalous ascent (Lake Superior shows two such cells that correspond to the double peak in weakened boundary layer winds around 86.5° and 90.0°W).

The location of anomalous descent in the IL experiments also agrees with the changes in sea level pressure (SLP), as the strongest pressure increases above the lakes occur in eastern sectors and even larger rises are seen downstream over land (Fig. 6). A gain in surface pressure in the IL simulations is expected as a linear response to boundary layer cooling caused by imposed ice cover. However, the core of the anticyclonic anomalies in IL is situated eastward of the strongest turbulent heating reductions over the lakes and is collocated with the maximum surface cooling that occurs just downstream (not shown). This horizontal displacement in anomalous temperature and SLP is driven by cold-air advection from the background wind flow (Fig. 2). A simpler alternative hypothesis to explain the weaker over-lake winds is that the ice cover has a higher surface roughness than open water. However, this possibility is rejected because RegCM4 uses the same roughness length for both surface types, consistent with observational estimates of smooth, first-year sea ice and open ocean conditions (Smith 1988; Wadhams 2000).

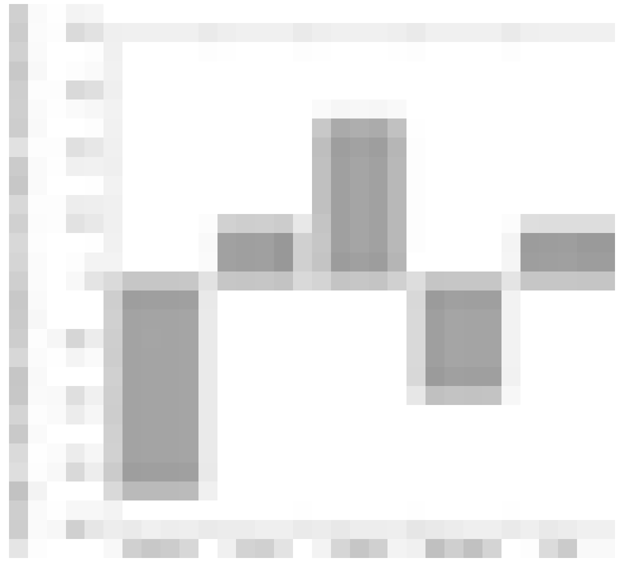

FIG. 8. Measure of the relative sensitivity of each lake to imposed total ice cover in experiment IL minus CONTROL. See text for an explanation of how this graph is constructed.

The variation among the Great Lakes in their overall response to imposed ice cover in this study is encapsulated in Fig. 8. This summary graph provides the relative sensitivity S of the meteorological response around each lake in experiment IL versus CONTROL. The S values represent the overall response in IL for each lake compared with the average response of all five lakes, on the basis of nine variables (those presented in Fig. 4 plus downstream near-surface specific humidity). The departure from the mean all-lake response in IL of each variable for each lake was calculated as the difference between the change in the variable for that lake minus the 10-case average change over all the lakes, divided by the standard deviation of the variable's response across all 10 cases:

$$S_{1,2} = \sum_{V=1}^{9} \frac{\dfrac{(\mathrm{IL}-\mathrm{CONTROL})_{\mathrm{LAKE}} - \overline{(\mathrm{IL}-\mathrm{CONTROL})_{\mathrm{LAKES}}}}{\sigma_{(\mathrm{IL}-\mathrm{CONTROL})_{\mathrm{LAKES}}}}}{9}, \tag{1}$$

where $S_{1,2}$ is the sensitivity index for a lake in each of its two HLES cases (the average of the two S values is plotted in Fig. 8), V represents the variable, the overbar denotes the all-lake average, and σ is the standard deviation of a variable's response across all 10 cases.

Because the mean change in all variables except SLP was negative in IL minus CONTROL, we converted the normalized SLP change from positive to negative for consistency before calculating the normalized average of all nine variables for each of the 10 cases. The sign of the numbers shown in Fig. 8 indicates whether a lake's sensitivity to imposed ice cover is higher or lower than the average sensitivity of all the lakes (negative meaning more sensitive, positive meaning less), while the magnitude quantifies how far a lake's sensitivity departs from the all-lake mean sensitivity (large values mean a larger departure). For example, the large negative index for Lake Superior (−0.68) denotes that freezing over this lake produces the most pronounced atmospheric response of all in experiment IL, while Lake Michigan (−0.37) shows the second strongest response and one that is greater than the Great Lakes average. By contrast, imposing ice cover on the other three lakes induces weaker-than-average atmospheric changes, with Lake Ontario (+0.52) the weakest of all.

c. All-lake simulations

The results from the IL experiments suggest that surface conditions on one lake can significantly affect another lake downstream and that this influence depends strongly on the particular Great Lake. To investigate this point further, we conducted an analogous set of ice-cover experiments for the same 10 events, but this time we imposed 100% ice concentration (1 m thick) simultaneously over all five of the lakes. Comparing the meteorological response from these AL to the IL simulations provides a measure of the regional influence of lake ice cover. In general, the added impact of freezing all the lakes (AL) compared with only an individual lake (IL) is smaller than the atmospheric response between IL and CONTROL (Figs. 4 and 9). This is especially true for the turbulent heat fluxes, which show an insignificant change from IL to AL because they both were already nearly shut off upon complete ice coverage on a single lake (Fig. 4). However, of the variables that do change between IL and AL, the effect of icing over all the lakes is to amplify the response to freezing up a single lake. For example, most cases produce an even greater reduction in snowfall in AL, although the difference is only pronounced for Lake Ontario. In addition, the surface winds weaken even more in all but the two Lake Ontario cases, and downstream cooling is enhanced except for Lake Superior (unchanged from IL). A strong and consistent amplification of the response from freezing all lakes is seen in SLP, which increases in every case and by amounts comparable to the pressure rise caused by icing over individual lakes. The 10-case average increase in SLP and the corresponding average decrease in PW were the only two changes shown here that were statistically significant (95% level) between the response in AL compared with IL (Table 2).

The most noticeable result in comparing AL to IL versus IL to CONTROL is the contrast between Lake Superior, which shows very little additional sensitivity to imposed regionwide ice cover, and Lake Ontario, which is strongly affected by surface conditions on the other lakes. Among the five Great Lakes, Ontario changes the most between IL and AL in terms of downstream snowfall, precipitable water, low cloud amount, SLP, and air temperature (Fig. 9). The contrasting dependence on regional versus local ice conditions between these two lakes is depicted by the spatial pattern of simulated snowfall in the three sets of experiments (Fig. 10). The heavy snowfall downstream of Lake Ontario in both CONTROL cases is reduced but not eliminated by freezing over the lake in IL. However, all lake-effect snowfall adjacent to Ontario disappears when the entire Great Lakes become ice covered in AL, presumably due to the influence from the upstream lakes (especially Huron). Conversely, Superior's lake-effect snowfall is completely suppressed when it ices over (IL) and remains so when all the lakes are ice covered.

The influence of the upstream lakes on the Lake Ontario response in AL is exemplified by the behavior of precipitable water (Fig. 11), whose spatial pattern resembles that of SLP and temperature, which also amplified considerably between IL and AL (not shown). In Ontario's IL cases, ice cover lowers PW partially over the lake but also in a narrow band downstream of the lake, matching the location of heavy lake-effect snowfall in CONTROL (Fig. 10). However, the magnitude of this reduction is rather small (<0.75 kg m^{-2} right next to the lake and even lower values farther inland). By contrast, the impact on PW above and downwind of Lake Ontario in the AL experiments is much more dramatic. The PW concentrations fall by more than 1 kg m^{-2} both above the lake and inland, and large decreases of 0.75 kg m^{-2} extend throughout central New York. Lake Huron is also strongly affected by upstream conditions (on Lake Superior), such that its over-lake decrease in PW is nearly the same in Huron's two IL cases and Superior's two IL cases (not shown). On the other hand, the PW change around Lake Superior in both cases is nearly identical in IL and AL. Instead, the major difference is the much more expansive *regional* drying of the atmosphere in AL, such that PW declines by over 0.25 kg m^{-2} across nearly the entire Great Lakes Basin and beyond.

The simulated influence of one Great Lake affecting snowfall on another is consistent with other studies. Byrd et al. (1995) identified a strong dependence of Lake Ontario's lake-effect snowfall on upstream air-mass modification by Lake Huron. Sousounis and Mann (2000) described how the existence of multiple lakes causes aggregate-scale circulations to develop over the Great Lakes from clusters of heat sources and sinks, which in turn affect lake-effect snowfall. A similarly complex set of multiple-lake interactions during cold-air outbreaks was documented by Mann et al. (2002), who showed that the resulting regional-scale circulations can cause a time-dependent response in lake-effect snowfall that evolves over the course of a lake-effect storm.

FIG. 9. As in Fig. 4, but for the difference between AL minus CONTROL compared with IL minus CONTROL (i.e., the supplemental impact of freezing all lakes compared with freezing an individual lake). The scales are as in Fig. 4 to facilitate comparison.

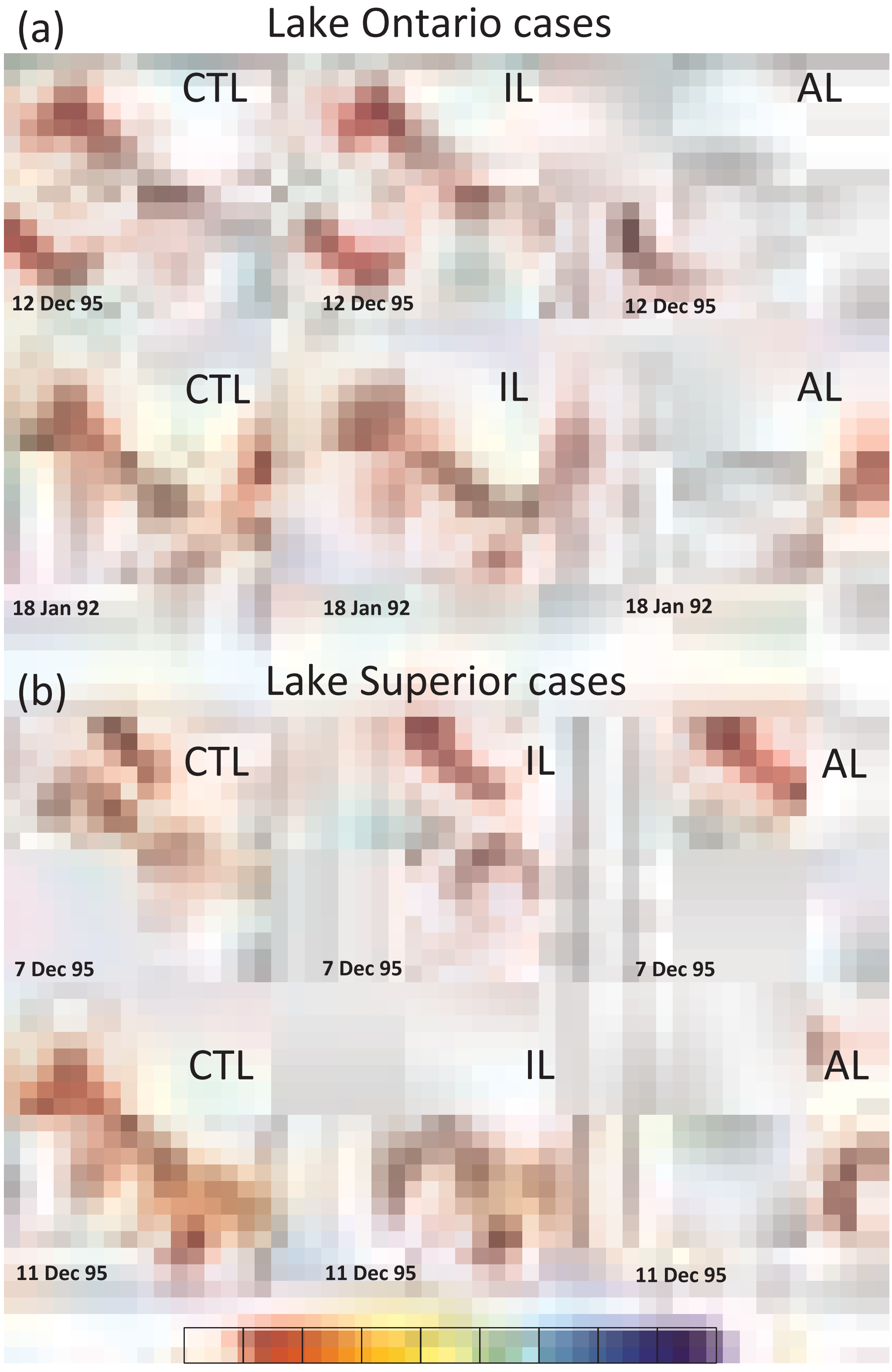

FIG. 10. Daily snowfall (cm) for the CONTROL (CTL), IL, and AL experiments for (a) Lake Ontario [(top) 12 Dec 1995, (bottom) 18 Jan 1992] and (b) Lake Superior [(top) 7 Dec 1995, (bottom) 11 Dec 1995].

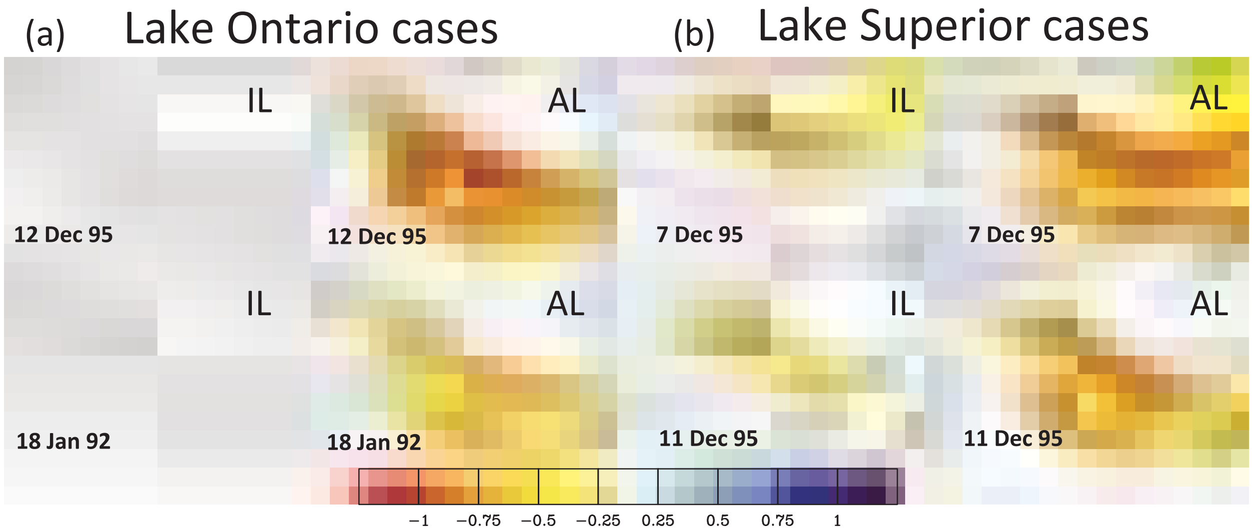

FIG. 11. Change in precipitable water (kg m^{-2}) from the CONTROL simulation for the IL and AL experiments for (a) Lake Ontario [(top) 12 Dec 1995, (bottom) 18 Jan 1992] and (b) Lake Superior [(top) 7 Dec 1995, (bottom) 11 Dec 1995].

We summarize these inter-lake differences between the atmospheric response to local versus regional ice cover through an index shown in Fig. 12. The values reflect the difference for each lake between its response in experiment AL (minus CONTROL) compared with experiment IL (minus CONTROL), on the basis of the same nine variables used in Fig. 8. The normalized difference of each variable between AL and IL is calculated as the variable's change in experiment AL (i.e., AL − CONTROL) minus the variable's change in experiment IL (IL − CONTROL), divided by the standard deviation of this difference among all 10 cases. Because we are primarily interested in the *magnitude* of the response in experiment AL relative to IL, all the normalized differences are taken as absolute values when calculating the nine-variable mean shown here:

$$D_{1,2} = \sum_{V=1}^{9} \frac{\left| \dfrac{(\mathrm{AL}-\mathrm{CONTROL})_{\mathrm{LAKE}} - (\mathrm{IL}-\mathrm{CONTROL})_{\mathrm{LAKE}}}{\sigma_{[(\mathrm{AL}-\mathrm{CONTROL})-(\mathrm{IL}-\mathrm{CONTROL})]_{\mathrm{LAKE}}}} \right|}{9}, \tag{2}$$

where $D_{1,2}$ is the difference index for a lake between AL and IL, V represents the variable, and σ is the standard deviation of a variable's response between AL and IL across all 10 cases.

Small values, such as for Lake Superior (0.05), indicate very little difference between the response in experiments IL and AL and thus a high degree of independence in how the atmosphere around the lake reacts to ice cover locally versus regionally. By contrast, the large value for Lake Ontario (2.12) indicates a large difference in this lake's response between the two experiments, such that Ontario is highly affected by ice cover elsewhere in the Great Lakes Basin.

4. Discussion

Although most of the simulated changes in the atmospheric response to imposed lake ice cover described here are robust and statistically significant, in this section we consider some possible caveats. One natural question is whether the relatively small set of 10 HLES events in this study is representative of the larger variety of observed lake-effect snowstorms. Ideally, RegCM4 would be used to simulate many more events than the two per lake presented here, but practically such an approach is unfeasible. We assume instead that the major features in the selected cases are fairly typical of other heavy storms and base this assumption on several factors. First, the spatial distribution of lake-effect snowfall in these events is similar to observed climatologies of favored lake-effect snow zones (Scott and Huff 1996), suggesting that the processes producing heavy snowfall in the model occur commonly. Second, the high degree of consistency, statistical significance, and physical plausibility among the simulated HLES events suggests that the major findings should apply generally,

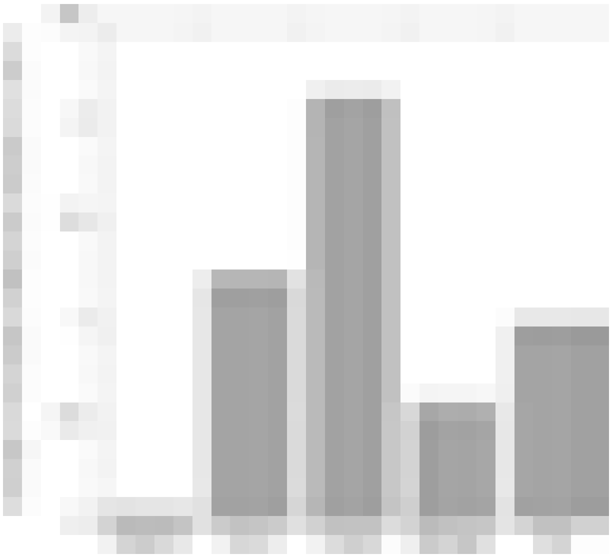

FIG. 12. Measure of the overall difference in the atmospheric response around each lake between experiment AL minus CONTROL vs experiment IL minus CONTROL caused by imposing complete ice cover. See text for an explanation of how plot was constructed.

even though certain heavy lake-effect snowstorms may not conform. Third, the synoptic patterns among the cases in this study are similar to each other and representative of those during typical heavy lake-effect snowstorms, consisting of northerly to westerly outflow from a polar anticyclone advecting cold air over the Great Lakes (NZV). These factors support our contention that the main conclusions drawn from the 10 events presented here can be applied to other heavy lake-effect snowstorms in the region.

A second possible concern surrounds the simulated ice cover in the CONTROL simulations; namely, the large variation among events and RegCM4's excessive ice concentrations (Table 1). This model bias could skew our results by reducing the influence of ice cover changes between completely frozen lakes in IL and AL and partially frozen lakes in CONTROL. This discrepancy means that our simulations generally provide a conservative estimate of the impact of ice cover, although we note that in 3 of the 10 events the simulated ice concentration is equal to or greater than observed. Furthermore, latent and sensible heating are almost completely suppressed in the IL and AL experiments, regardless of the open water amount in CONTROL, although a caveat is that the RegCM4 ice model does not contain leads (cracks of open water within a frozen grid box) and thus could exaggerate the impact of ice cover. We also find no significant correlation between the downstream snowfall in IL and the corresponding lake's evaporation rate or ice concentration in the CONTROL runs. In addition, Gerbush et al. (2008) observed that sensible heat fluxes over Lake Erie decrease nonlinearly with ice coverage, with very weak sensitivity to ice concentrations below 70%, although latent heat fluxes were found to vary linearly. In only one event did RegCM4 simulate a lake to be mostly ice covered in CONTROL; in every other case, the model agreed with observations that the majority of the lake was open water.

A related issue is that our experiments include not only the impact of a change from open water to ice cover, but also a change from regions of relatively thin ice cover in CONTROL (<50 cm) to much thicker ice (1 m) in IL and AL. In our analysis, we have implicitly assumed that the former transition dominates the atmospheric response, but the shift from thin to thick ice could conceivably affect our results by also cooling the lake surface and curtailing heat fluxes. However, several factors suggest that the change from open water to ice cover is much more important. First, the fact that the lakes were predominantly open in the CONTROL cases (and even ice-free on Lake Erie) means that the largest areal change in surface conditions between CONTROL and IL/AL is from open water to ice cover, rather than from thin ice to thicker ice. Second, the spatial pattern of turbulent heat fluxes over the lakes in CONTROL indicates much larger energy loss from regions of open water than over ice-covered areas (not shown) and therefore a predominant influence from freezing over these ice-free sectors. Third, one-dimensional thermodynamic ice models show a dramatic, nonlinear sensitivity in sensible and latent heat fluxes as a function of thickness; surface energy losses decline sharply between open water and thin ice, whereas the difference in heat fluxes between thin and thick ice is relatively smaller (Maykut 1978).

5. Conclusions

A 20-km version of the RegCM4 regional climate model has been employed to investigate the impact of ice cover on heavy lake-effect snowstorms in the Great Lakes region. The model simulates a set of 10 HLES cases reasonably well, generating lake-effect snowfall amounts within 10% of observed for the all-event average. RegCM4 simulates a strong influence of thick ice cover on a wide range of atmospheric variables above and downstream of the lakes, including snowfall, surface energy fluxes, wind speed, temperature, moisture, clouds, and air pressure. Compared with the CONTROL simulations, the effect of completely freezing over individual Great Lakes in the IL experiments is to significantly decrease downstream snowfall (by 84%), low cloudiness (33%), and temperature (3.1 K) when averaged over all 10 events (Table 2). In addition, the ice cover nearly eliminates sensible and latent heat fluxes over the lakes (>90% reduction) and causes a large decline in over-lake, near-surface wind speeds (33%) that is linked to an increase in SLP centered over and east of each frozen lake. The primary mechanism for the reduced lake-effect snowfall in the IL simulations is interpreted to be thermodynamic: the dramatic suppression of turbulent heating over the lake (especially evaporation) due to ice cover. However, this study also finds that the weakened surface winds represent a secondary, dynamical effect that amplifies the thermodynamic reduction of lake-effect snowfall by 1) reducing the turbulent heat fluxes over the lake and 2) reducing the downstream surface wind convergence that promotes HLES. If we reverse the sign of these changes, this finding suggests that heavy lake-effect snowfall is supplemented dynamically by a localized circulation response to open water.

By quantifying the impact of ice cover from a particular lake on downstream snowfall and associated atmospheric variables, this study suggests that Lake Superior (and secondarily Lake Michigan) has the strongest and most widespread influence on heavy snowfall in the basin and Lake Ontario the least. A similar conclusion of the prominence of Lake Superior was reached by Scott and Huff (1996), who reasoned that the lake's large surface area, great depth, and west–east orientation parallel to the prevailing wintertime wind direction are factors favorable for producing lake-effect snowfall. In the current study we identify another important element for the varying influence of each lake on the atmospheric response to ice cover: their location relative to each other and to the large-scale wind flow during HLES. Not only is Lake Superior the largest of the Great Lakes, but its upstream location to the northwest of the other lakes gives it the greatest potential to affect other lakes downwind (but not be affected by them), as evidenced by the anomaly maps of SLP and PW (Figs. 6 and 11). Similarly, conditions on Lake Michigan can propagate downstream when the large-scale flow is westerly, although this lake can also be influenced by Lake Superior under a north-northwesterly flow pattern. On the other hand, Lake Ontario is both the smallest of the Great Lakes and located such that it can be influenced by Lakes Superior, Huron, and even Michigan. There is a strong relationship between the relative sensitivity of each lake to imposed ice cover locally (Fig. 8) and the relative independence of each lake's response to regional (vs local) ice conditions (Fig. 12). The correlation coefficient of these two indexes across the five lakes is extremely high ($r = 0.96$), and the large spread between Lakes Superior and Ontario is very apparent. One caveat is that we did not investigate the aggregate effect of the Great Lakes or the relative roles of the individual lakes in initiating mesoscale circulation anomalies during these events. Such collective lake disturbances (CoLD) during cold-air outbreaks have been shown to affect lake-effect snowfall in a complex and time-transgressive manner (Weiss and Sousounis 1999; Sousounis and Mann 2000; Mann et al. 2002).

Although this study is focused on contemporary climatic conditions, the findings are relevant for assessing how future climate change may affect lake-effect snowstorms. The expected warming this century should accelerate the trend toward reduced ice cover on the Great Lakes, which has been occurring over the past few decades and is well replicated by RegCM4 (NZV). The results of the current study suggest that this change toward more open water should favor significantly greater lake-effect snowfall. At the same time, however, climate models project fewer extreme cold-air outbreaks over the Great Lakes (Vavrus et al. 2006), a shift that makes heavy lake-effect snowfall less likely. We will be researching the interplay between these two factors in our future work to foster improved societal responses and adaptation measures to impending Great Lakes climate change.

Acknowledgments. This study was funded by grants from NOAA Climate Change Data and Detection (Grant NA09OAR4310108), along with EPA-funded contracts from the Michigan Department of Natural Resources (Contracts 751B0200072 and 751P1301081). The authors are thankful for assistance from ICTP, particularly Drs. Graziano Guiliani, Nellie Elguindi, and Filippo Giorgi, in debugging and improving the RegCM4 model. Computational resources were provided through the Teragrid from the University of Texas at Austin and the University of Illinois at Urbana–Champaign (supported by the National Science Foundation Office of Cyberinfrastructure, Grant 0503697 ''ETF Grid Infrastructure Group: Providing System Management and Integration for the TeraGrid'').

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