NOTES AND CORRESPONDENCE

A Climatology of Collective Lake Disturbances

CHRISTOPHER C. WEISS AND PETER J. SOUSOUNIS

Department of Atmospheric, Oceanic, and Space Sciences, University of Michigan, Ann Arbor, Michigan

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ABSTRACT

This study examines the frequency and intensity with which collective lake disturbances (COLDs) develop. These disturbances develop when cold air overspreads the Great Lakes region in winter. The heat and moisture that is transferred from the Great Lakes aggregate into the lower atmosphere, and that spreads across a large region, allows eventually for the development of a meso-α-scale pressure perturbation and circulation.

Cases from the period 1980–90 were identified based on the existence of a surface trough or closed low over the Great Lakes region in the presence of cold air. Output from the Limited-Area Fine Mesh (LFM) model was used rather than performing numerous with-lake and no-lake numerical simulations to determine whether the feature was indeed the result of aggregate heating by the lakes. The LFM did not include the lakes in its simulations, so the 24-h forecast served as an optimal no-lakes simulation. Subtracting the initialization sea level pressure (SLP) field valid at the same time allowed for an assessment of the COLD events in terms of the SLP perturbation.

An average of 33 events per year with an average SLP perturbation of 3–4 hPa was found for the 10-yr period. The synoptic-scale conditions for weak events with SLP perturbations less than 3 hPa differed significantly from those for strong events with SLP perturbations greater than 9 hPa. The weak scenario was characterized by a weak trough over the Great Lakes with high static stability and weak cold advection below 500 hPa and weak vorticity advection at 500 hPa. The strong scenario was characterized by a nearly closed low over the Great Lakes with low static stability and strong cold advection below 500 hPa and strong positive vorticity advection at 500 hPa.

The current study is the first attempt to measure the frequency and intensity with which the Great Lakes collectively generate meso-α-scale disturbances in winter. The LFM-based technique provides a result that cannot likely be obtained without a herculean effort from a numerical modeling standpoint. Future numerical studies using the identified scenarios, however, will be extremely useful to better understand the sensitivities of COLD events to the large-scale conditions.

1. Introduction

It has long been known that the Great Lakes have a considerable impact on the weather in that region. The most celebrated if not the most frequently studied of these impacts has been the lake-effect snowstorms that develop as cold air flows over the relatively warm lakes in late fall and winter. During such cold air outbreaks, the heat and moisture provided by the lakes destabilizes the lowest 300 hPa of the troposphere, and the resulting convective precipitation is responsible for 50%–80% of snowfall accumulations along the lee sides of the Great Lakes (cf. Holroyd 1971). Dozens of studies have documented various aspects of lake-effect storms (e.g., Mitchell 1921; Forbes and Merritt 1984; Kelly 1986; Hsu 1987; Hjelmfelt 1990).

While lake-effect snows are likely the most noticeable, and arguably the most intense effect of the Great Lakes on the atmosphere in winter, there exists a larger, lake aggregate scale[1] or collective lake disturbance (COLD) that is also meteorologically and climatologically significant. This COLD event also develops during cold air outbreaks, but as the result of heat and moisture from all of the lakes spreading across a large region. This aggregate scale influence by the Great Lakes was first noted by Cox (1917), who demonstrated that the lakes tend to attract and strengthen lows in the winter and highs in the summer.

Since that observational study by Cox (1917), few

Corresponding author address: Dr. Peter J. Sousounis, Atmospheric, Oceanic and Space Sciences Department, University of Michigan, Ann Arbor, MI 48109-2143.
E-mail: sousou@umich.edu

[1] The term aggregate scale refers to a horizontal distance spanned approximately by the upper Great Lakes (e.g., Superior, Michigan, and Huron) from west to east.

studies have focused on aggregate effects, in direct contrast to the comprehensive studies that have focused on the individual-lake effects (e.g., lake-effect storms). Petterssen and Calabrese (1959) determined theoretically that the lake aggregate can generate a sea level pressure perturbation of 6–7 hPa. Danard and Rao (1972), Danard and McMillan (1974), and Boudra (1981) used numerical models to examine the collective thermodynamic and dynamic effects of the lake aggregate on relatively strong synoptic-scale extratropical cyclones. They found that the strengths of the cyclones appeared to overwhelm and obscure many of the effects of the lakes. In particular, Boudra (1981) simulated a storm that had a central mean sea level pressure of 995 hPa and six closed isobars at 4-hPa intervals when it was over Lake Huron. In that study, the differences between simulations with and without the lakes showed that the forcing from the lakes reduced the sea level pressure by about 3–4 hPa, but that neither the position nor the circulation of the storm changed significantly. In contrast, Sousounis and Fritsch (1994) demonstrated that the Great Lakes could impact significantly the paths and intensities of weak highs and lows. In their numerical study, a comparison of with-lakes and no-lakes simulations demonstrated that the with-lakes cyclone appeared to accelerate rapidly into the lakes region where it deepened 5 hPa and lingered for about 12 h, and that the no-lakes cyclone followed a steady path southeast of the lakes and only deepened 2 hPa.

The above-mentioned studies indicate that COLD events can be manifested in different ways. Figure 1a shows that for situations where a strong synoptic-scale low is exiting the Great Lakes with cold air over the region, a COLD event can appear as an enhanced pressure trough extending across the region (cf. Petterssen and Calabrese 1959). Figure 1b shows that for situations where no preexisting low is exiting or approaching the region, the heat and moisture from the lake aggregate can generate in situ a lake aggregate scale cyclonic circulation called a mesoscale aggregate vortex (MAV). Sousounis (1997) defined a MAV as a meso-α-scale warm-core vortex approximately 500–1000 km wide and 2–4 km deep with cyclonic vorticity in the lower half and anticyclonic vorticity in the upper half that develops by aggregate heating and moistening from the Great Lakes. An MAV is usually identifiable on standard surface weather charts as a weak low over the Great Lakes region with approximately one to three closed isobars at 2-hPa intervals. The outermost closed isobar typically encloses an area approximately as large as that spanned by the upper Great Lakes (e.g., Lakes Superior, Huron, and Michigan). Sousounis (1998) concluded that synoptic-scale diabatic heating was important early but that MAV-enhanced warm advection (WAD) and MAV-enhanced diabatic heating were significant contributors to mesoscale aggregate vortex development later.

With the exception of the observational study by Cox (1917), the few aggregate studies that have been performed to date have been on a case-by-case basis and have precluded a focus on the climatological aspects of COLD events. For example, while it has been known for some time that the lakes influence paths and intensities of low pressure systems in winter, the frequency and the magnitudes of the influences that occur over a winter or over a period of several winters are not known. Lake aggregate effects likely occur frequently because they appear to be induced under synoptic-scale conditions that are similar to those that generate lake-effect storms. Additionally, the magnitudes are likely significant. Ten-day numerical simulations with and without the lakes by Bates et al. (1993) indicated that a negative pressure perturbation with amplitude $\sim$3 hPa and hor-

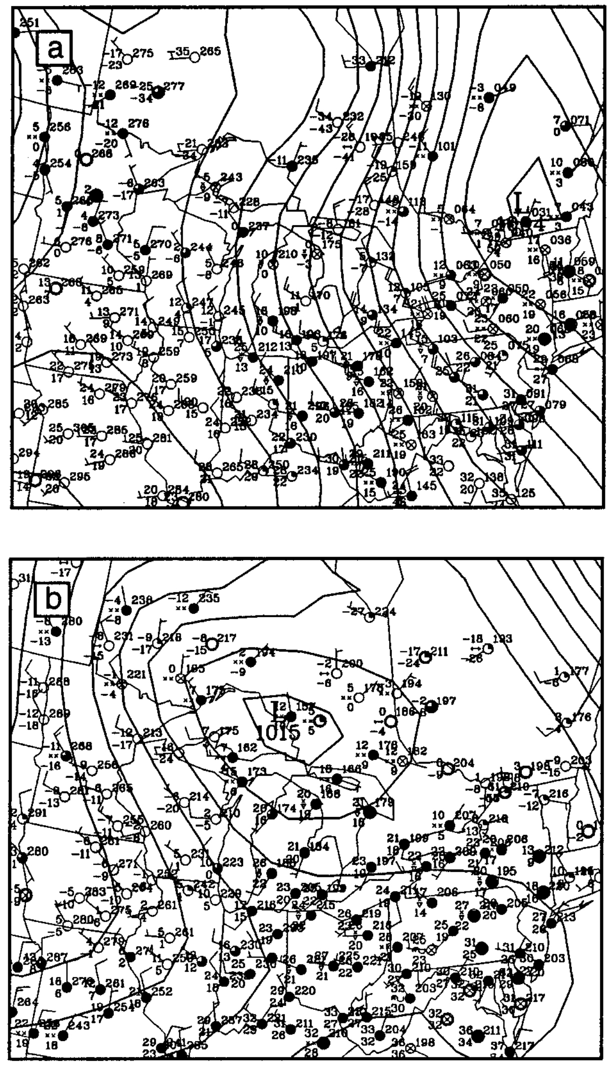

FIG. 1. Surface observations and LFM sea level pressure analyses (2-hPa contour interval) for two COLDs that formed under different types of synoptic-scale conditions. (a) Enhanced trough, which formed in strong flow. Conditions valid 1200 UTC 2 Jan 1981. (b) Closed low, which formed in weak flow. Conditions valid 1200 UTC 9 Feb 1986.

izontal scale ~800 km was located northeast of Lake Huron during the period.

The absence of climatological studies dealing with COLD events may be a consequence of the fact that these effects are not easily identified, either because they are embedded within strong synoptic-scale flow or because they appear to be synoptic-scale features that are generated dynamically. The few case studies that have been performed have revealed many important aspects of COLD events. These studies implemented the strategy of comparing numerical simulations with all of the Great Lakes included to those with none of the Great Lakes included. While the strategy of comparing with-lakes versus no-lakes numerical simulations has proven to be beneficial from a case study standpoint, it is computationally inappropriate from a multiple case (multiple year) study (e.g., climatological) standpoint.

The objective of this paper is to present a climatology of COLD events without having to perform with-lakes and no-lakes numerical simulations for all of the identified cases. In section 2, the datasets and methodology that are used to identify lake aggregate effects are described. In section 3, the climatological aspects of COLD events including the frequency of occurrence and the typical ranges of intensity are presented. The preferred synoptic-scale situation(s) associated with COLD events is also examined. A summary and conclusions are presented in section 4.

2. Methodology

An underlying premise based on previous work is that COLD events develop as a result of thermal destabilization by the lakes. Thus the months of November through February were identified for this study as those that would support the development of lake aggregate disturbances. Surface and upper-air observations, as well as model output from the Limited-Area Fine Mesh (LFM) model[2] were available for a 10-yr period from 1980 to 1990, so that a total of 40 months of data from November 1980 to February 1990 was examined.

The first phase of the identification procedure involved a visual assessment of COLD events. This assessment was accomplished in part by inspection of National Oceanographic and Atmospheric Administration (NOAA) Daily Weather Maps (DWMs) to identify COLD days (e.g., days on which a COLD event was found to exist). These maps provide the sea level pressure (SLP) field with a 4-hPa contoured analysis, selected surface observations, 500-hPa heights, 24-h liquid equivalent precipitation, snow depth, and 24-h high and low temperatures for 1200 UTC of each day. For a given day to be included as a COLD day, one important criterion was that a trough/closed low in the sea level pressure field on the scale of the lake aggregate had to have been present in the ‘‘Great Lakes region’’ as defined by Angel and Isard (1997) and as depicted in Fig. 2. For example, the existence of a ‘‘synoptic-scale wake trough’’ (cf. Fig. 1a) signaled a possible lake aggregate influence. A second important criterion was that the air upwind from the lakes had to be colder than the lakes at the given time for the selected day or within the past 24 h in order for positive heat and moisture fluxes to have existed over the lakes. For example, a November day characterized by surface air temperatures below 0°C along the north shore of Lake Superior with a lake surface temperature of ~4°C with northerly flow was acceptable. A January day characterized by southwesterly flow over the region with surface temperatures near 5°C in Wisconsin was also acceptable if surface temperatures on the preceding day were such that heat and moisture fluxes over the lakes were positive (e.g., heat/moisture transfer from the lakes to the air). The motivation for including these warm advection cases was to include events that apparently were initiated because of aggregate heating but that may have been enhanced later by synoptic-scale forcing. Other thermodynamic criteria such as 850-hPa temperatures or 1000–500-hPa thicknesses were not used. An upper thickness limit was not imposed because 1) it was found that low surface temperatures and positive heat and moisture fluxes could still exist at the surface even at high thickness values (e.g., >552 dam) and 2) other synoptic-

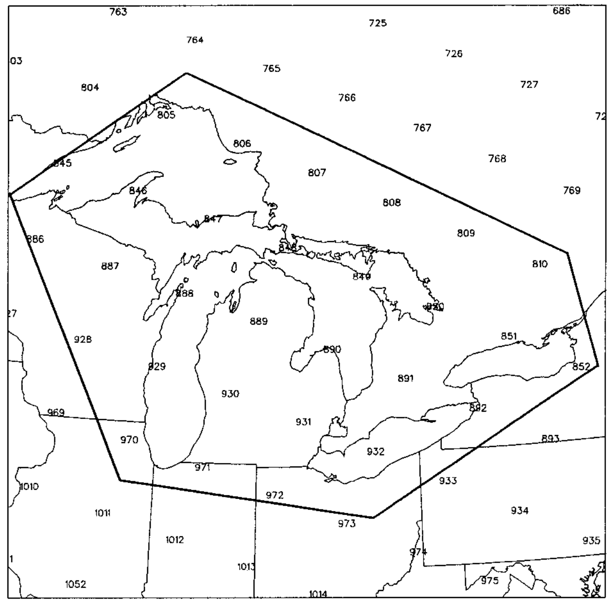

FIG. 2. Location of Great Lakes region (solid polygon) for purposes of identifying COLD days. Numbers indicate grid points used for locating maximum SLP perturbations.

[2] The LFM model output used for this study was provided via magnetic tape by the National Center for Atmospheric Research in Boulder, Colorado. The tapes contained all of the initialization and forecast output from the 0000 and 1200 UTC runs for the 40-month period. The LFM itself has a grid resolution of 190.5 km. Additional information regarding this model is available in Gerrity (1977).

scale factors (e.g., strong positive vorticity advection could have compensated for ''marginally'' cold air. Finally, the warm-core characteristics of MAVs were not considered as a criterion because MAVs represent only a subset of COLDs.

Visual inspection of the DWMs was convenient for identifying possible COLD days. But, it was not sufficient to identify all possible days because a 4-hPa analysis was not sufficiently fine to identify the weaker cases where the true perturbation in pressure due to the lake aggregate influence was less than 4 hPa. Also, the fact that the DWMs were hand-drawn brought a subjective component into the technique. Thus, use of the DWMs was complemented in the first phase by inspection of the initialized sea level pressure field from the LFM model. This field was contoured using existing software at a 2-hPa interval, which allowed for the detection of the weaker cases. Inspection of DWMs and LFM model output led to nearly 500 potential COLD days being identified in which a lake aggregate scale trough or low existed in the presence of cold surface air.

Because lake aggregate scale troughs and lows are commonly found over the Great Lakes region during the winter months and develop for a variety of reasons, it was necessary to decide in the second phase of the identification procedure whether each trough and low was really the result of lake aggregate forcing (e.g., a COLD) or synoptic-scale forcing (e.g., a shortwave) or both. As was mentioned in the introduction, to make this decision by comparing with-lake and no-lake numerical simulations for each case would have been extremely time consuming and expensive. A more efficient and cost-effective method was to use the LFM model output once again. The LFM model never included heating or moistening from the lakes in its calculations and in fact omitted the Great Lakes altogether. It was only through the initial conditions themselves that there was ever any inclusion of the influence of the lakes.

To optimize the model's utility as a valid no-lakes simulation, a forecast hour had to be selected that was sufficiently late so that the lake heating and moistening that was present at the initialization had been advected from the region, but sufficiently early so that the model forecasts were still accurate in regions away from the lakes. The 12-h LFM forecast was in general a poor choice for a no-lake simulation because the heat and moisture effects from the lake-influenced initial conditions were likely present to some degree.[3] Likewise, the 36- and 48-h LFM forecasts were in general poor choices for no-lakes simulations because the systematic errors that generally resulted from the coarse grid resolution and inadequately simulated physical processes (e.g., convection, radiation) rendered the simulations at these times relatively inaccurate. Thus, the 24-h LFM model forecast was a reasonable compromise and served as an optimal no-lakes simulation.

For each potential case that was identified in the first phase, the lake aggregate induced pressure perturbation was obtained in the second phase by subtracting the 24-h forecast sea level pressure field from the initialization field valid at the same time. Using this technique, a COLD day that was identified in the first phase was retained only if a negative or positive pressure error (e.g., perturbation) was *centered* somewhere over the Great Lakes region (cf. Fig. 2) at 1200 UTC on that day. If the error was centered outside the region, then it was assumed that it was related to a poorly forecast synoptic-scale storm track or intensity and not related to lake aggregate heating. Interestingly, very few of the cases (e.g., only three) that were identified in the first phase with errors centered in the lakes region were found to have a (weak) positive pressure perturbation, which is consistent with the fact that the LFM forecast was omitting the positive fluxes of heat and moisture that were responsible for generating the negative pressure perturbations. Owing to the presence of a trough or closed low with cold air over the region at the time or immediately preceding it, and to the presence of a negative pressure perturbation, it was thereby hypothesized that the perturbation was due mainly to aggregate heating and moistening by the lakes and hence was identified as a COLD day.

Application of the LFM pressure perturbation technique yielded a total of 328 COLD days. A COLD event was then defined, of which 238 were found, as one or more consecutive COLD days under the influence of the same area of low or high pressure. Influence was determined objectively by whether a closed isobar in association with the low or high encompassed the center of the Great Lakes region as shown in Fig. 2.

The LFM analysis technique was convenient but it was not perfect. The pressure perturbations that were calculated using the LFM forecast and analysis model output were probably not entirely a result of the lakes. A systematic error in the track of a low near the Great Lakes region could have increased or decreased the pressure perturbation and could have skewed the number, intensity, and location of the COLD events that were identified. For example, the number of weak COLD cases may have been underestimated because there may have been negative systematic errors that were large enough to cause the center of the perturbation to lie outside the lakes region, or positive systematic errors that were large enough to more than cancel the lake-induced negative perturbation, thereby masking it and creating a net positive perturbation. Either of the above scenarios could have caused a case to be excluded. These scenarios are unlikely, however, because Silberberg and Bosart (1982) and Grum and Gyakum (1986)

[3] This assessment was based on the approximate time it would take for a thermal perturbation generated on the northwest side of the aggregate to be advected across the upper Great Lakes region (e.g., Superior, Michigan, and Huron). For systems moving around 10 m s^{-1} this time is $\sim$24 h.

TABLE 1. Number of COLD days found at various intensity ranges (in hPa) and average SLP perturbations (in hPa) for the 10 seasons. Percentages enclosed in parentheses. Note that the seasons 1984/85, 1985/86, and 1986/87 each had one COLD day where perturbation pressure had opposite sign, so the sums of the number of cases for those seasons do not equal the total number of days.

Season	Days	Avg pert.	$0 < P < 3$	$3 < P < 6$	$6 < P < 9$	$9 < P < 12$	$P > 12$
1980/81	26	−5.32	6 (23)	11 (42)	6 (23)	1 (4)	2 (8)
1981/82	34	−4.09	14 (41)	14 (41)	2 (6)	2 (6)	2 (6)
1982/83	27	−4.60	8 (30)	10 (37)	9 (33)	0 (0)	0 (0)
1983/84	34	−5.34	8 (24)	13 (38)	8 (24)	5 (15)	0 (0)
1984/85	33	−4.43	12 (36)	12 (36)	5 (15)	1 (3)	2 (6)
1985/86	41	−5.08	4 (10)	23 (56)	12 (29)	1 (2)	0 (0)
1986/87	26	−4.30	6 (23)	11 (42)	7 (27)	1 (4)	0 (0)
1987/88	39	−5.22	11 (28)	14 (36)	9 (23)	3 (8)	2 (5)
1988/89	39	−5.51	7 (18)	17 (44)	10 (26)	4 (10)	1 (3)
1989/90	29	−5.07	11 (38)	7 (24)	4 (14)	6 (21)	1 (3)

both found SLP errors from the 24-h LFM forecast to be approximately −1 hPa (e.g., negative systematic error) for situations that included all lows during the 1978/79 winter and all highs during the 1981/82 winter, respectively. These studies suggested furthermore that this error was likely from omission of the lakes and not from some other systematic (non–lake related) bias. The average of all the perturbation magnitudes should thus be a good indication of the true impact of the lakes because other relatively small errors would cancel.

3. Results

General findings regarding the COLD cases, and their sensitivity to the synoptic-scale flow, are described below.

a. Frequency and intensity of COLD events

Table 1 shows the number of COLD days per year. Six of the 10 years had 33 or more days. The least number of days in any year was 26 and the most was 41. Interannual variability is apparent in the numbers; especially noticeable are the minima that occurred during the 1980/81, 1982/83, and 1986/87 winter seasons. These seasons were the three warmest of the 10 in Wisconsin, which suggests a relationship between the average temperature upwind of the lakes and the number of cases per winter.

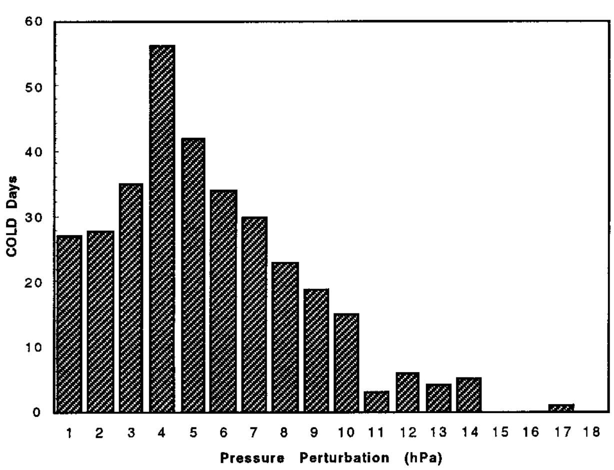

FIG. 3. The number of COLD days that were found for each season (e.g., Nov–Feb) in the study.

Table 1 shows also that there is no apparent relationship between the number of cases in a season and the average intensity of those cases. Additionally, the percentage of cases in each of the categories does not seem to be related to the total number of cases. For example, the five seasons with the fewest total number of cases and the five seasons with the most total number of cases both consisted of ~66% of the cases exhibiting an SLP perturbation less than 6 hPa, and ~10% of the cases exhibiting an SLP perturbation greater than 9 hPa. If one assumes a direct relationship between the intensity of the cold air and the strength of COLD events, then these results suggest that years that are characterized by infrequent cold air outbreaks are not necessarily characterized by weak cold air outbreaks.

Figure 3 shows that the greatest number of cases had pressure perturbations of 3–4 hPa. This result is consistent with those from other studies (e.g., Bates et al. 1993) that found cold season lake aggregate induced SLP perturbations of approximately 3–4 hPa in magnitude. Approximately 10% of the cases had SLP perturbations exceeding 9 hPa. Such large-amplitude perturbations from lake aggregate forcing during dynamically weak conditions have not been noted in the few studies that have been published. However, the large-amplitude cases identified in this study were usually accompanied by strong dynamical forcing of some sort. Feedbacks between the forcing and the aggregate heating could explain these large pressure perturbations.

b. Synoptic-scale impacts

Figure 4 shows some of the average synoptic-scale conditions that accompanied the average COLD event based on all of the cases identified. Surface conditions were characterized by the presence of a well-defined trough over the Great Lakes region that tilted westward with height. The trough had an amplitude of ~3 hPa

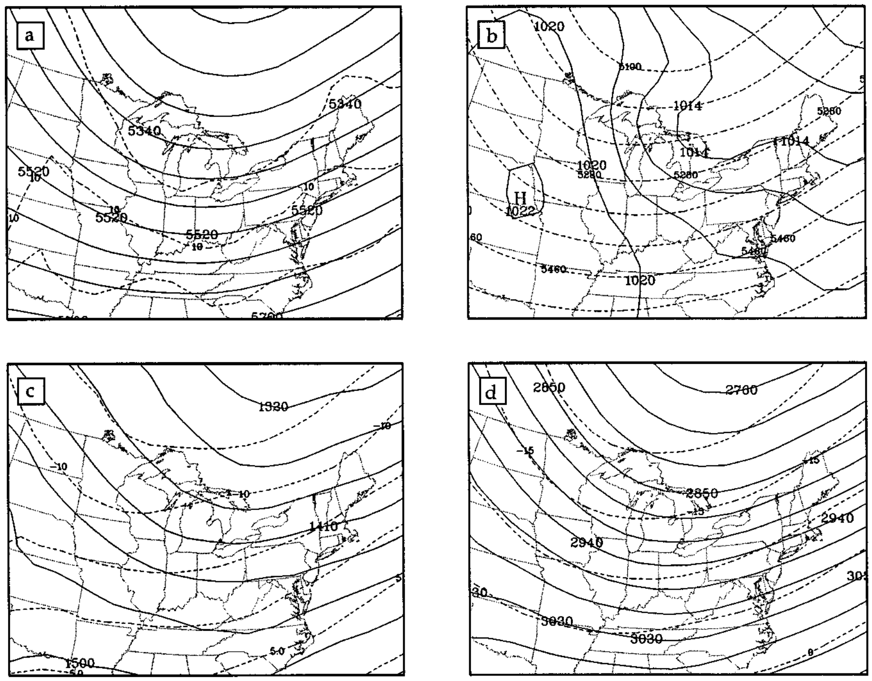

FIG. 4. Average synoptic-scale conditions for all COLD days found from LFM analyses. (a) 500-hPa heights (solid, 60-m contour interval) and absolute vorticity (dashed, 2×10^{-5} s^{-1}). (b) Sea level pressure (solid, 2-hPa contour interval) and 1000–500-hPa thickness (dashed, 60-m contour interval). (c) 850-hPa heights (solid, 30-m contour interval) and temperature (dashed, 5°C contour interval). (d) 700-hPa heights (solid, 30-m contour interval) and temperature (dashed, 5°C contour interval).

(cf. Fig. 5), which is consistent with the results in Fig. 3, and was located over Sault Ste. Marie, Michigan. Other characteristic features associated with the average COLD event include the existence of cold advection (CAD) at 700 and 850 hPa. The cold advection below 500 hPa was apparently the result of exiting low pressure over Nova Scotia and entering high pressure over the Great Plains. At 500 hPa, the flow was characterized by weak positive vorticity advection.

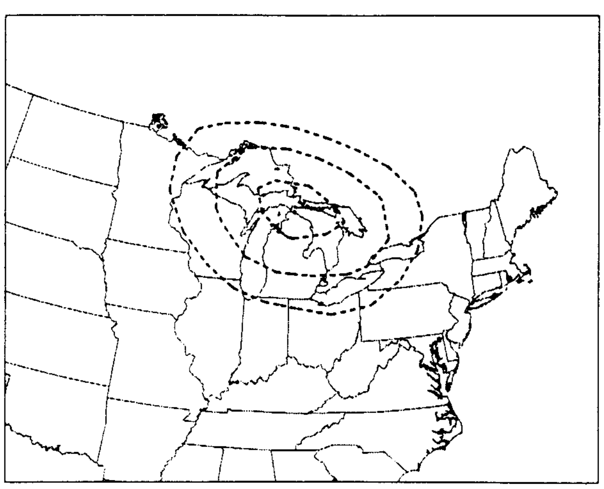

FIG. 5. Average SLP perturbation for all COLD days. Outermost dashed contour is -1 hPa. Contour interval is 1 hPa.

The conditions shown in Fig. 4 represent those for all of the COLD events that were identified. To examine whether conditions for weak events with SLP perturbations less than 3 hPa were different than those for strong events with SLP perturbations greater than 9 hPa, data were analyzed for several meteorological parameters.[4] The choices for these parameters were motivated by the similar type of environment in which COLD events and lake-effect storms develop and the documented sensitivity of lake-effect storms to certain synoptic-scale variables as noted by Burrows (1991). For example, stability, temperature advection, and vorticity advection have all been shown to correlate with lake-effect snow events as well as with larger synoptic-scale systems. Values for selected parameters for each of the cases were obtained using a nine-point average centered on the grid point where the SLP perturbation was greatest.

Table 2 indicates that the 500-hPa heights decreased steadily from weak events to strong events. Table 3

[4] Geostrophic winds were used for all of the analyses in Table 1 because actual LFM initialization winds were not recorded until December 1983.

TABLE 2. Average intensities of selected parameters for different strength categories of COLD days. Parameters from left to right include 500-mb height (H500), 850–700-mb static stability (STAB), 500-mb vorticity advection (VAD5), 700-mb temperature advection (TAD7), 850-mb temperature advection (TAD8), and 700-mb geostrophic wind speed (V7SPD) and direction (V7DIR). The values are based on nine-point averages centered on the LFM grid point where SLP perturbations are greatest. The unit VU means vorticity unit: 1 VU = 10^{-5} s^{-1}.

Size (mb)	H500 (m)	STAB °C (100 mb)$^{-1}$	VAD5 VU day^{-1}	TAD7 °C day^{-1}	TAD8 °C day^{-1}	V7SPD (m s^{-1})	V7DIR
0–3	5359	6.59	−0.09	−1.40	−1.75	13.26	266.14
3–6	5334	6.45	0.44	−1.88	−2.52	13.00	256.22
6–9	5279	6.34	3.32	−3.68	−5.28	14.13	252.77
9–12	5228	6.07	6.23	−7.72	−10.54	15.53	249.76
>12	5184	5.86	14.06	−8.23	−11.06	11.74	250.16

shows a proportionate increase in the number of cases where the 500-hPa height was lower than the average value of 532 dam (e.g., the lower-tropospheric air was cold), as the magnitude of the SLP perturbation increased. The relationship between 500-hPa heights and COLD intensity suggests that low 500-hPa heights were likely important for strong COLD development. The above results can be explained by the fact that colder air resulted in stronger fluxes of sensible and latent heat at the surface, which resulted in stronger SLP perturbations.

Table 2 indicates that the 850–700-hPa static stability decreased steadily from weak events to strong events, which suggests that low values of static stability were likely important for strong COLD development. However, Table 3 shows that the percentage of cases where the stability was greater than the average value of 6.42°C (100 hPa)$^{-1}$ was the same as the percentage of cases where the stability was less than the average value. Thus, stability values for the low stability portion of the cases in the strong event category were lower than those in the weak event category. The relationship between stability and COLD intensity can be explained by the fact that low-stability air is more likely to generate strong vertical motions and strong SLP perturbations than high-stability air.

Table 2 indicates that the 500-hPa vorticity advection increased steadily from weak events to strong events. Table 3 shows a proportionate increase in the number of cases that exhibited positive vorticity advection as opposed to negative vorticity advection as the magnitude of the SLP perturbation increased. Both of these results suggest that high values of positive vorticity advection were likely important for strong COLD development, which is consistent with the fact that upper-level dynamical forcing has been shown to enhance a low-level thermally forced circulation.

Table 2 indicates that the intensity of the cold advection associated with COLD events increased almost steadily from weak events to strong events. Table 3 shows that while more than one-third of the weak events were characterized by WAD at 700 and 850 hPa, the events become increasingly characterized by CAD as the perturbation size increased. Additionally, Table 3 indicates that the average SLP perturbation amplitude of the CAD cases was larger than that for the WAD cases. The above results suggest that strong cold advection at 850 and 700 hPa were likely important for strong COLD development. The fact that WAD cases did exist,[5] however, indicates that in cases of greater static stability, the vertical motions afforded by warm advection may have helped to make the environment more conducive for COLD events to develop. The correlation between cold advection and perturbation strength is somewhat curious, because on one hand, warm advection—not cold advection—is typically associated with strong synoptic-scale systems. On the other hand, strong cold advection may have helped to maintain strong low-level diabatic heating and low static stability, which are typically associated with strong synoptic-scale systems.

Tables 2 and 3 illustrate that the relationship between 700-hPa wind speed and COLD intensity is not as straightforward as some of the others. In this regard it is useful to recall the impact of wind speed (shear) on two other related weather phenomena. Wind shear plays a key role in the development of synoptic-scale baroclinic waves. For most cases, a larger thermal wind increases the baroclinic instability, allowing systems to develop. In contrast, wind shear can have a detrimental effect on convective systems, such as tropical cyclones. Because COLD events likely develop from a combination of thermal and dynamic forcing mechanisms, it is understandable that wind (shear) could have a complicated effect on development. An examination of Tables 2 and 3 shows approximately that wind speed increases with increasing strength, except for the strongest event category, where it drops abruptly. This behavior may be explained by the fact that higher wind speeds (and decreasing 500-hPa heights) indicate the southward approach of a baroclinic zone for weak- and medium-strength COLD events, followed by decreasing wind speeds as the baroclinic zone shifts south of the Great Lakes region for strong COLD events. Another possible explanation is that lake-induced convection during

[5] The WAD cases were not likely an artifact of the analysis technique because 1) WAD was occurring over a large (e.g., nine-point region) and because 2) air temperatures were lower than lake temperatures, at least at the beginning of the events, so that surface fluxes of heat and moisture were positive.

TABLE 3. Number of COLD days exhibiting parameter values meeting selected criteria: H500 < 5317 m, STAB > 6.42 °C/100 hPa, VAD5 > 0, TAD7 > 0, TAD8 > 0, V7SPD > 16.29 m s^{-1}, and V7DIR = south of west, for different strength categories of COLD days. Criteria values for H500, STAB, V7SPD, and V7DIR are mean values based on all COLD days. Percentages enclosed in parentheses.

Size (mb)	H500 (m)	STAB °C (100 mb)$^{-1}$	VAD5 VU day^{-1}	TAD7 °C day^{-1}	TAD8 °C day^{-1}	V7SPD (m s^{-1})	SW Wind	No. days
0–3	35 (39)	38 (42)	48 (53)	33 (37)	31 (34)	46 (51)	55 (61)	90
3–6	62 (47)	67 (51)	70 (53)	45 (34)	42 (32)	73 (55)	97 (73)	132
6–9	46 (64)	36 (50)	47 (65)	18 (25)	12 (17)	31 (43)	50 (69)	72
9–12	20 (83)	12 (50)	15 (63)	4 (17)	3 (13)	9 (38)	21 (88)	24
>12	8 (80)	6 (60)	9 (90)	1 (10)	1 (10)	4 (40)	6 (60)	10

strong COLD events may extend to greater heights as described by Niziol (1987) and Byrd et al. (1991) and thereby reduce through turbulent mixing the wind speeds at 700 hPa.

Finally, Tables 2 and 3 show that the majority of COLD events existed when 700-hPa geostrophic winds were southwesterly. The 700-hPa geostrophic wind direction became increasingly southwesterly from weak events to strong events. The presence of the southwesterly flow is surprising because (a) northwesterly winds are the ‘‘prevailing winds’’ over the Great Lakes during the winter months, and (b) the southwesterly winds suggest that warm advection should have been occurring, for which few cases were found. The curious nature can be understood by considering the fact that strong events may be associated with higher-amplitude baroclinic troughs located just to the west of the Great Lakes and the presence of stronger east–west, rather than north–south, temperature gradients (cf. Fig. 7).

c. *Weak versus strong scenarios*

The results in Tables 2 and 3 suggest that weak events, with SLP perturbations less than 3 hPa, may have developed under different synoptic-scale conditions than strong events, with SLP perturbations greater than 9 hPa. Figures 6 and 7 are similar to Fig. 4, except they exhibit the synoptic-scale conditions for weak and strong events, respectively. The weak scenario in Fig. 6 shows a weak trough in the SLP field over the Great Lakes region and a ridge oriented from north to south from Minnesota to Arkansas. A trough in the thickness field is located over the eastern Great Lakes. The strong scenario in Fig. 7 shows a nearly closed low in the SLP field with a low center just northeast of Lake Huron. A strong ridge is located farther to the west than in the weak scenario and extends north to south, from the Dakotas to Texas. Strong cold advection exists over the western Great Lakes with 1000–500-hPa thicknesses

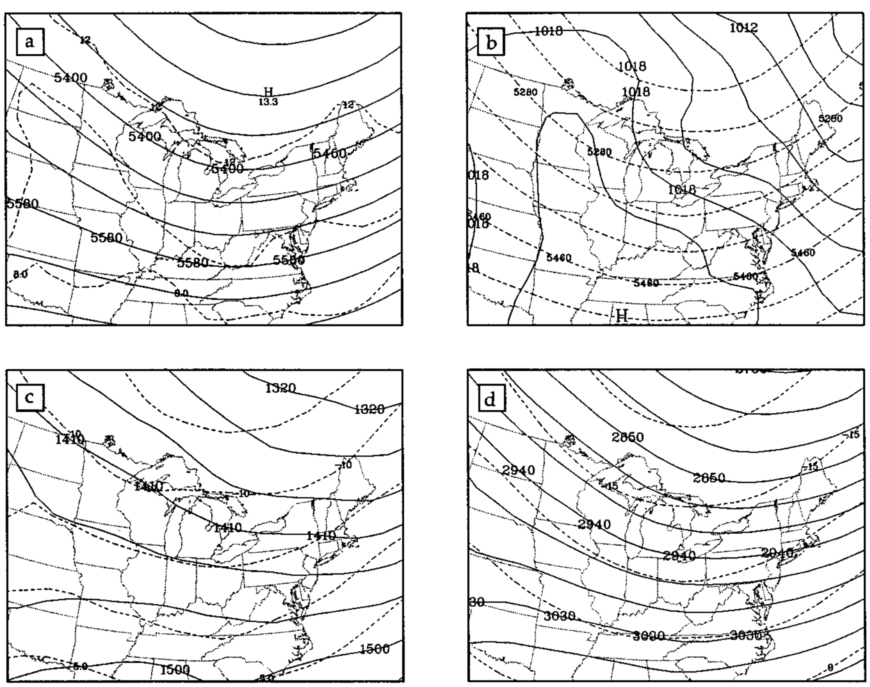

FIG. 6. Similar to Fig. 4 but showing average conditions for COLD days with SLP perturbations less than 3 hPa.

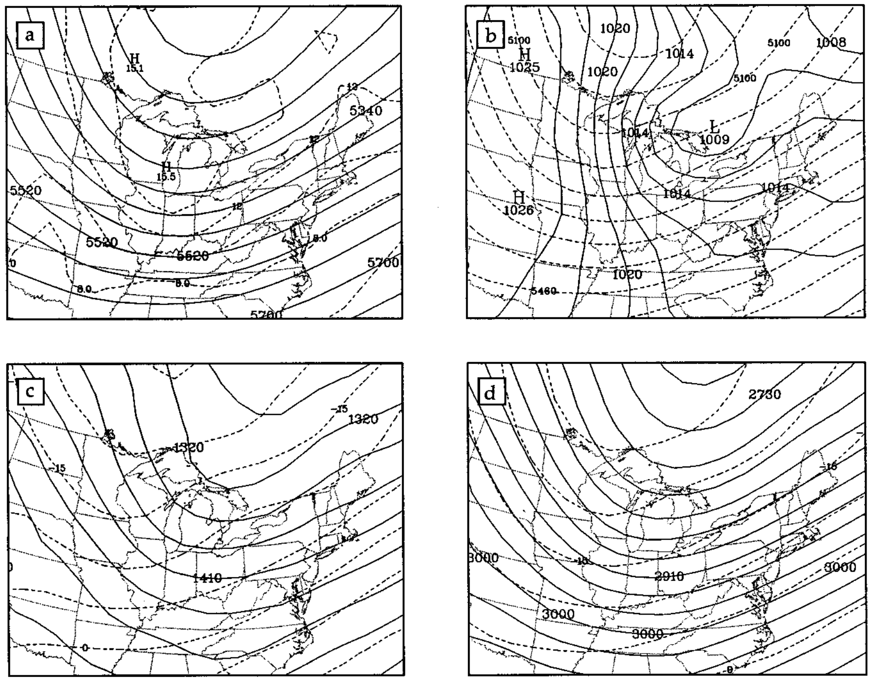

FIG. 7. Similar to Fig. 4 but showing average conditions for COLD days with SLP perturbations greater than 9 hPa.

more than 10 dam lower than in the weak scenario. The stronger cold advection (and lower thicknesses) in the strong scenario may have contributed to stronger fluxes of heat and moisture from the Great Lakes and hence may have contributed directly to the stronger SLP perturbations. Other differences between the weak and strong scenarios include the presence of weak cold advection at 850 and 700 hPa for the weak scenario and stronger cold advection over the western Great Lakes with weaker warm advection over the eastern Great Lakes for the strong scenario. Finally, the positive vorticity advection at 500 hPa is considerably weaker for the weak scenario than it is for the strong scenario.

The stronger dynamical forcing for the strong scenario suggests that strong COLD events may develop from a combination of dynamical and thermodynamical forcing. However, the simultaneous presence of very cold air, and strong temperature and vorticity advections, make it difficult to make a more specific assessment of the relative importance of each forcing mechanism.

4. Summary and conclusions

The Great Lakes affect weather on a variety of space scales and timescales in winter. On small space scales and short timescales, an individual lake can generate narrow lake-effect snow bands when cold air flows across the relatively warm lake surface. On a larger space scale and longer timescale, the Great Lakes also can generate collectively a negative pressure perturbation near the surface on the scale of the entire Great Lakes region. This collective lake disturbance (COLD) is a phenomenon that occurs in the ‘‘unstable season’’ from November to February over the Great Lakes region when there is a large difference in temperature between the lakes and the air above them. This lake–air temperature difference allows small lake-scale circulations and negative pressure perturbations to develop over each of the individual Great Lakes that can evolve eventually into a meso-α-scale pressure perturbation and circulation as the aggregate heat and moisture from all of the lakes spreads across a large region. The frequency and intensity distributions for the aggregate-scale effects have not been addressed previously in the literature.

The current study was conducted to better understand some of the climatological aspects of COLD events and some of the large-scale flow characteristics that are associated with them. The study encompassed a 40-month period including the months of November, December, January, and February for the period 1980–90. The COLD days were identified by inspection of NOAA Daily Weather Maps and comparison of LFM analyses and 24-h forecasts of SLP valid at the same time. A total of 328 COLD days (238 COLD events) were found for the 10-yr period. The number of events per season ranged from 26 to 41, with 33 events being the mean. The average SLP perturbation for all events was 3–4 hPa.

For each of the COLD days that was found, the synoptic-scale environment was assessed to determine possible correlations between COLD intensity and large-scale forcing. A correlation existed between temperature-related variables and the SLP perturbation of the COLD event. In particular, COLD intensity had a strong negative correlation with 500-hPa heights and midlevel static stability. These correlations supported the rather intuitive assertion that the COLD magnitude should be proportional to the thermal instability that creates it. Because lake-effect snows have been found to depend on this measure of thermal instability, it is understandable that COLD events would as well.

The synoptic-scale conditions for weak events with SLP perturbations less than 3 hPa differed significantly from those for strong events with SLP perturbations greater than 9 hPa. The weak scenario was characterized by a weak trough over the Great Lakes with high static stability and weak cold advection below 500 hPa and weak vorticity advection at 500 hPa. The strong scenario was characterized by a nearly closed low over the Great Lakes with low static stability and strong cold advection below 500 hPa and strong positive vorticity advection at 500 hPa.

It is concluded that the aggregate heating and moistening from the Great Lakes aggregate can generate approximately 25–40 COLD days per year. The SLP pressure amplitudes of these events are typically on the order of 3–4 hPa, but may be enhanced to values in excess of 9 hPa when dynamical forcing from favorable synoptic-scale flow exists.

The current study is a first attempt to measure the frequency and intensity with which the Great Lakes collectively generate meso-α-scale disturbances in winter. The simplistic technique of examining LFM model output provides a result that cannot otherwise be obtained without a herculean effort from a numerical modeling standpoint. Future numerical studies using the identified scenarios, however, will be extremely useful to understand better the sensitivities of COLD events to the large-scale conditions.

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