P2.1

SIMULATIONS OF ENVIRONMENTAL CONDITIONS CONDUCIVE TO THE FORMATION OF LAKE-TO-LAKE BANDS

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ABSTRACT

Lake-effect snow has a large influence on the Great Lakes region, often enhancing snowfall downwind of the lakes. Several different types of lake-effect bands exist, but lake-to-lake (L2L) bands remain the least studied. L2L bands are cloud bands that extend from one lake across intervening land and a downwind lake, and occur due to preconditioning in the heat and moisture fields of the atmosphere by the upwind lakes. The purpose of this study is to examine conditions conducive to the development of L2L bands. The Weather and Research Forecasting-Advanced Research WRF (WRF-ARW) Version 2.2 model is used to simulate and compare two cases of cold air outbreaks over the western Great Lakes. The case of 02 December 2003 produced L2L bands while the case of 15 February 2007 started with broken lake-effect band production and later converted into a L2L band case. Sensitivity studies involving both an increase and a decrease of 5°K in the lake surface temperatures, as well as a change in the surface roughness length of the land surrounding the western Great Lakes, were conducted to investigate impacts on the L2L bands.

Examination of cross-sections and plan-view images of model output reveal that the two cases were accurately simulated by the model. Of significant importance for the formation of lake-effect bands of any type include instability, wind speed and direction, and lake surface-850 hPa temperature difference. For the December case, the northwesterly flow of air over Lake Superior created superadiabatic lapse rates over the water surfaces with mixing continuing over land surfaces. Only a slight decrease in boundary layer depth occurred as air passed over the Upper Peninsula of Michigan. With the lake surface temperatures increased by 5°K, the well-mixed boundary layer and cloud bands expanded in depth and more lake-effect band formation occurred to the south as more heat and moisture became available from the lake surface. The opposite effect happened when the lake surface temperatures were decreased by 5°K; a reduction in bands over land masses was apparent. The February case showed similar boundary layer conditions to the December case initially. However, when the lake surface temperatures were increased by 5°K, the boundary layer remained similar to the base simulation, but L2L bands were delayed, suggesting that very large lake-land temperature differences may hinder L2L bands. The decrease in 5°K again produced thinner cloud and boundary layers. For both events, the change in land-surface roughness length produced no variations in the simulations.

Overall, inversion heights were comparable with time between the two cases. Lake surface-850 hPa temperature differences were typically higher for the February simulations, but L2L bands were generally observed to form between 15°C and 24°C in the simulations. Wind speed and direction appears to affect L2L band formation the most in this study. Surface wind speeds at the start of L2L convection were approximately 15 m s^{-1} in both events. With diminished wind speeds, a change of wind direction from northwest to westerly resulted in the end of L2L bands.

1. INTRODUCTION AND BACKGROUND

The Great Lakes are a major influence on the weather of the surrounding region. During the winter season the Lakes especially modify snowfall, increasing snowfall totals over and downwind of the lakes. In a study by Braham and Dungey (1984), it was estimated that one-fourth to one-half of the annual snowfall that falls over Lake Michigan can be credited to lake-effect snow.

When cold air crosses over a lake of relatively warm water, lake-effect snow can occur. In the winter months, lakes are typically warmer than the air from arctic air outbreaks, due to the heat capacity of water. When the cold air traverses the warmer lake, instability is created as a result of the ensuing large temperature gradient.

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Past studies have grouped the convective clouds into four different classes: wind-parallel bands or widespread disorganized convection, shore-parallel bands, midlake bands, and mesoscale vortices (e.g., Hjelmfelt, 1990; Niziol *et al.,* 1995). More recently, the importance of bands extending from lake-to-lake (L2L bands) has been recognized (Rodriguez *et al*., 2007). Lake-to-lake bands are identified as lake-effect snow bands that extend over two or more lakes at one time. Forecasting of this specific type of lake-effect band remains difficult, partially due to insufficient research (Rodriguez *et al*. 2007).

Four processes have been suggested as possibly important to L2L formation: (1) propagation of heat and moisture plumes downwind, (2) lake-induced circulation growth to a downwind lake, with a major contribution in growth as a result of large temperature differences between the air and the surface of the lake (Chang and Braham, 1991), (3) gravity wave initiation by an upwind lake, and (4) band internal microphysical and radiational processes able to sustain propagation of the band. Hjelmfelt *et al.* (2004) examined the impact of Lake Superior on lake-effect snow over Lake Michigan. Simulation results showed rapid warming and moistening of near-surface air. A sensitivity study removing Lake Superior from simulations revealed a boundary layer over Lake Michigan half the depth of a boundary layer with Lake Superior present. These effects were found to continue across to northern Lower Michigan; it was hypothesized that the distance of the land between Lakes Superior and Michigan is not sufficiently long enough to disrupt the boundary layer above it. Similar results were obtained observationally by Chang and Braham (1991) and Kristovich *et al*. (2003). In the observational study by Kristovich *et al*. (2003), sounding, radar, and aircraft data were investigated to determine Lake Michigan's effect on the boundary layer over the lake and the state of Michigan. In addition to the warming and higher humidity encountered, wind shear was explored to examine its role in boundary layer evolution. Wind speed shear was present in the case study, but little directional wind shear was present. Mann *et al*. (2002) looked at the Great Lakes collectively and found that the combined effects of the Great Lakes increased the convective depth by lifting the inversion base and lowering cloud bases (due to increased moisture near the surface). Influences from Lake Superior were both direct and indirect; Lake Superior directly altered the boundary layer by dynamic changes due to the presence of a heat plume, and indirectly by interaction with the air over Lake Michigan downwind (Mann *et al*., 2002).

More recently, L2L cases were studied to determine weather patterns under which L2L bands occurred. Soundings were compared between L2L and non-L2L lake-effect cases, and the results showed that L2L bands are not more likely to develop under the most intense lake-effect conditions (Kristovich, 2008, personal communication). Also, an analysis of near-surface air downwind of Lake Michigan found that deep surface-based stable layers rarely formed when lake-modified air advected inland (Kristovich, 2008, personal communication). Instead, the stable layer over the cold landmass stayed very shallow and did not mix out, giving more of a neutral layer farther above the surface.

An observational study by Rodriguez *et al*. (2007) determined the frequencies of L2L bands over the Great Lakes. Most L2L cloud bands were found to originate over Lake Superior during the study period of October to March 2000 to 2004. This is consistent with the fact that most cold air outbreaks originate northwest of the Lakes. Lake Superior is also the largest of the Great Lakes by area and depth, and its east-west orientation yields a large fetch. Both of these conditions make it favorable for air traveling over Superior to attain a large amount of heat and moisture from the water.

Another L2L feature noted by Rodriguez *et al*. (2007) was the wavelength characteristics of the bands as they evolved downwind of a lake. As cloud bands originate over Lake Superior, the wavelengths of the bands are generally small; one case study noted cloud band wavelengths of less than 3 km. Downwind of Lake Superior, the wavelengths increased to an average of 5-7 km. This structural change of the clouds is indicative of growth of the boundary layer (Young *et al*., 2002), and suggests that the depth of the boundary layer has increased as the air crossed over the lake and land surfaces.

The purpose of this study is to examine the environmental conditions that allow lake-effect clouds to continue across the intervening land to the lake(s) downwind. The Weather and Research Forecasting-Advanced Research WRF (WRF-ARW) Version 2.2 (Skamarock *et al*., 2005) model was used to simulate two separate cases. The first case is an L2L band case from 02 December 2003 that was chosen due to its production of well-defined L2L bands. The second case is a similar cold air outbreak case from 15 February 2007 that produced broken lake-effect convection, hereafter referred to as typical lake-effect bands, and only later evolved to an L2L case. The February case will be referred to as the non-L2L case in some descriptions in following sections. The boundary layer interactions in this second case were used as comparison to the interactions in the actual L2L event. Besides performing base runs of the cases, sensitivity studies were designed to investigate some of the hypotheses regarding conditions that allow L2L bands.

2. OBSERVATIONS OF THE CASE DAYS

2.1 *The 02 December 2003 Case Day*

L2L bands began to form over Lake Superior and continued over Lake Michigan in the early morning hours of 02 December 2003, in response to cold, northwesterly flow. At 12 UTC on 02 December 2003, after the L2L event had already begun, a surface high pressure system was centered over western Wisconsin and Illinois and a surface low pressure system off the northeast coast of the United States. An 850 hPa ridge was centered over the Minnesota and Iowa region. In the northwesterly flow, 850-hPa temperatures at 12 UTC on 02 December ranged from -6°C to -14°C over western to eastern Lake Superior, respectively, and from -8°C to -14°C over southwestern to northeastern Lake Michigan, respectively. Surface air temperatures (Fig. 2.1) ranged from -14°C over the shores of western Lake Superior to -4°C over the state of Michigan at 12 UTC. Corresponding water temperatures (Fig. 2.2) were about 1°C over central Lake Superior and 2°C surrounding the center, and had an average of 5°C over northern Lake Michigan, with cooler temperatures in the shallower perimeter of the Lake. The temperature differences between 850 hPa and the lake surface were 8°C over western Lake Superior, 14°C over eastern Lake Superior, and 17-19°C over northern Lake Michigan. However, it is important to note that 850-hPa temperatures were actually colder over Lakes Superior and Michigan at 00 UTC 02 December 2003, prior to the event (not shown). Figure 2.2 also shows the ice cover over the Lakes for the event date. Based on the limited coverage and the results of Gerbush *et al*. (2008), it was concluded that there was too little ice cover on Lakes Superior and Michigan to be a major factor to L2L band formation and evolution.

A visible satellite image of L2L bands on 02 December 2003 from the Terra Moderate Resolution Imaging Spectroradiometer (MODIS) from the Space Science and Engineering Center (SSEC) at the University of Wisconsin-Madison is shown in Fig. 2.3. L2L bands at this time — 1715 UTC — were seen forming on Lake Superior and continuing downwind over Lake Michigan. Very little snow cover was present over the Great Lakes region through 17 UTC for this event, as the fraction of snow was 0% over most of the land and the open water at this time. Only upwind of Lake Superior and in bands over the Upper Peninsula of Michigan and northern Lower Michigan did the fraction reach 30-50 percent.

By 00 UTC 03 December 2003, the high pressure system was centered over western Lake Michigan and weaker, more variable winds occurred over the Great Lakes region. The L2L bands had broken up and other lake effect snow tapered off.

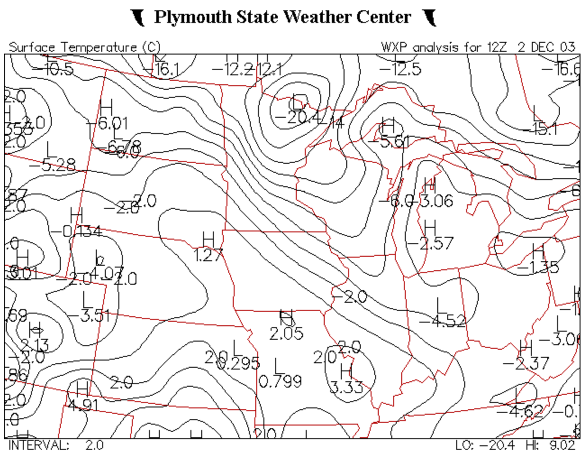


Figure 2.1. Surface temperature (°C) map for 02 December 2003 at 12 UTC from Plymouth State archives. Temperature contours are every 2°C.

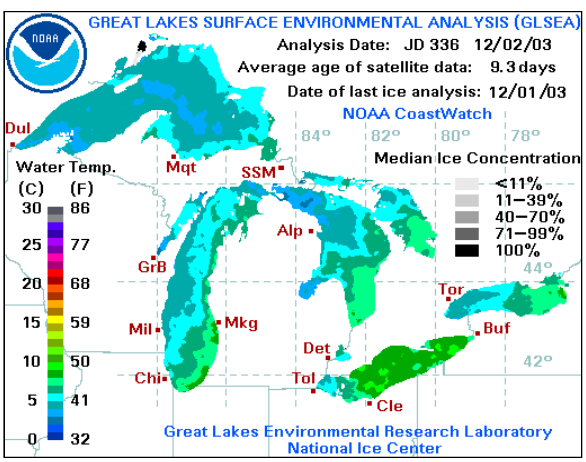


Figure 2.2. Lake surface temperatures and median ice concentration for the Great Lakes on 02 December 2003 from the Great Lakes Environmental Research Laboratory.

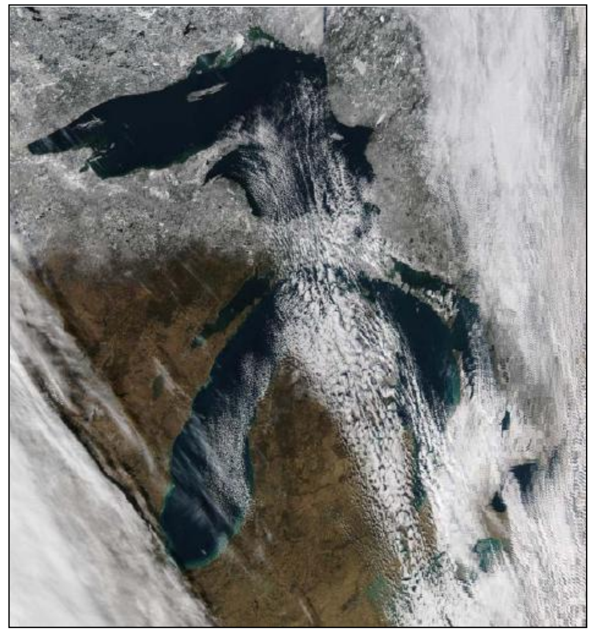


Figure 2.3. Terra MODIS visible satellite observations at 1715 UTC on 02 December 2003. L2L bands are observed over Lakes Superior and Michigan, as well as the Upper Peninsula and Lower Michigan.

2.2 *The 15 February 2007 Case Day*

The 15 February case is similar to the 02 December case in that it is a cold air outbreak day that produces lake-effect snow. However, L2L bands do not form until late in the period. Lake-effect snow initiated over Lake Superior in the morning hours of 15 February. The synoptic situation for this date is similar to the 02 December case; a high pressure system resided to the west of the Great Lakes in the wake of a low pressure system that was located over Maine at 12 UTC. The 850-hPa ridge was a bit farther to the west than the previous case, over the western Dakotas. Wind flow was from the northwest throughout the event, with surface wind speeds varying from 5-13 m s^{-1} (10-25 knots) over and upwind of the Great Lakes. At 850-hPa, temperatures varied from -18°C to -22°C over the western Lakes at 12 UTC. Surface air temperatures at 12 UTC, shown in Figure 2.4, over northern Lower Michigan and eastern Wisconsin, averaged -18°C, while over southwestern Michigan temperatures reached a maximum of -13°C. Temperatures upwind of Lake Superior and over the Upper Peninsula of Michigan ranged from -22°C to -16°C, with warmer temperatures of -10°C to -14°C along the shore of southern Lake Superior. Lake surface temperatures for 15 February, Fig. 2.5, show an average of 3°C for Lake Superior and 1°C for northern Lake Michigan. For this case, the lake surface to 850 hPa temperature difference over the region ranged from 21-23°C. Much of western Lake Superior and some of its perimeter is covered with some ice, and most of northern Lake Michigan is ice-covered as well (Fig. 2.5). Therefore, a possibility exists for the heat and moisture exchange between the Lakes and the air to be reduced in this case. This effect, however, may not be as great as expected, as recent research indicates that appreciable reduction in surface fluxes does not occur until ice coverage is nearly complete (Gerbush *et al*., 2008).

The Terra MODIS visible satellite image at 1615 UTC on 15 February 2007 is shown in Fig. 2.6. Though snow and ice cover on the land and water surfaces confuses the view of the clouds, bands were seen over Lake Superior and then over the eastern parts of Lake Michigan. Around 18 UTC, L2L bands formed in place of the typical lake-effect bands. This will be shown and discussed in Section 4. Lake-effect snow diminished shortly after 00 UTC 16 February 2007.

Much more snow was present in this event than in the December event. Fractional snow cover at 16 UTC on 15 February 2007 over land varied from 30-55% over the western Upper Peninsula and from 65-100% over the eastern Upper Peninsula.

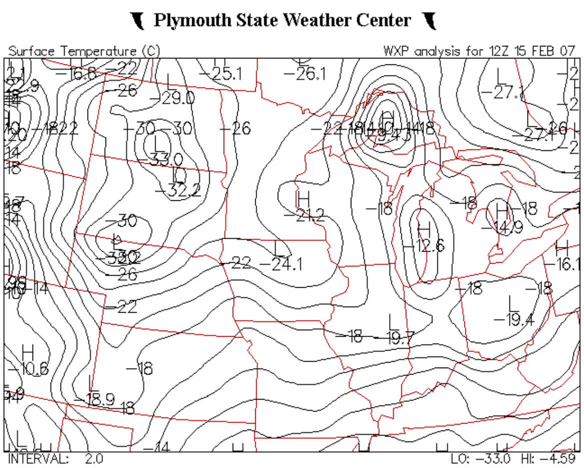


Figure 2.4. As in Fig. 2.1, but for 15 February 2007 at 12 UTC.

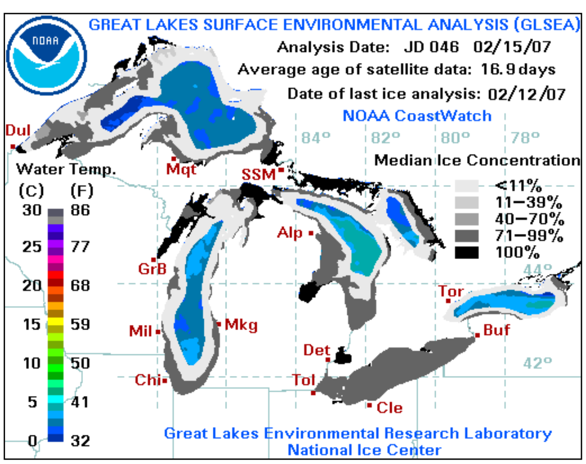


Figure 2.5. As in Fig. 2.2, but on 15 February 2007.

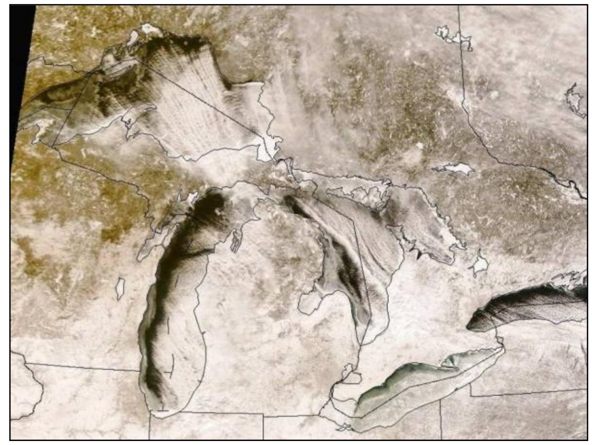


Figure 2.6. As in Fig. 2.3, but at 1615 UTC on 15 February 2007. Broken lake-effect convection is observed over Lakes Superior and Michigan.

3. MODEL DESCRIPTION

The Weather Research and Forecasting-Advanced Research WRF (WRF-ARW) model is an Eulerian, non-hydrostatic, fully compressible, three-dimensional model that utilizes a terrain-following hydrostatic-pressure vertical coordinate (Skamarock *et al*., 2005).

Three nested model domains were used in this study. The outer domain (domain 1) covered a broad region and consisted of 130 X 135 grid points with 15-km grid spacing. The second domain (domain 2) covers the Great Lakes region and used 250 X 196 grid points with 5-km grid spacing. The innermost domain (domain 3) encompassed the region of the western Great Lakes where L2L band formation was observed with 1.66-km grid spacing over 244 X 391 grid points. The domains, shown in Fig. 3.1, were configured with a Lambert Conformal map projection. For the outer two domains, 5-minute geographic data resolution was set, and for the inner domain, 30-second geographic data resolution was employed. In the vertical, 35 levels were utilized, with approximately 13 levels in the boundary layer.

Physics options include the Thompson *et al*. (2004) graupel scheme for cloud microphysics, the Noah land surface model (Noah LSM, Chen and Dudhia, 2001a), the Yonsei University (YSU) boundary layer scheme (Hong *et al*., 2006), and the Kain-Fritsch cumulus parameterization (Kain and Fritsch, 1993). The Kain-Fritsch cumulus parameterization was not used in the innermost domain.

Initial and boundary conditions for the simulations were provided by North American Regional Reanalysis (NARR) from the National Centers for Environmental Prediction (NCEP) (Mesinger *et al*., 2006). Lake surface temperature and ice data were provided by the Great Lakes Surface Environmental Analysis (GLSEA) from the Great Lakes Environmental Research Laboratory (GLERL, see Figs. 2.2 and 2.5). The GLERL data was retrieved as a pixel map in GIF format and was only incorporated into the innermost domain.

Each case was initialized at 18 UTC on the day prior to the event (01 December 2003 and 15 February 2007, respectively). This allowed for model spin-up time before the lake-effect bands began. Each run was completed at 06 UTC on the day after the event (03 December 2003 and 16 February 2007), when all evidence of lake-effect bands had ceased.

Figure 3.1. Domains selected for the numerical simulations. The outer box (domain 1) is 130 X 135 grid points with grid spacing of 15 km. The middle box (domain 2) centered over the Great Lakes is 250 X 196 grid points with grid spacing of 5 km. The inner box (domain 3) is centered to show L2L bands over the western Great Lakes, and is 244 X 391 grid points with grid spacing of 1.66 km.

A base run was completed for each of the cases, with additional sensitivity simulations completed to investigate L2L band and boundary layer behavior. Usually the lake-effect clouds are unable to continue across the intervening land, and the sensitivity studies investigate some of the hypotheses concerning conditions that allow L2L bands to form. Two sensitivity studies implemented a change in lake surface temperatures, one with a 5°K increase from the base case temperatures and the other with a 5°K decrease in lake surface temperatures. Only the lake water surface temperatures were changed; the temperatures of the ice and land surfaces did not change. Ice was also not created when lake surface temperatures dropped below freezing (WRF-ARW Version 2.2, Wang *et al.,* 2004, uses a temperature of 271°K to delineate ice) as a result of the 5°K temperature decrease. In this manner, ice neither froze nor melted. This is important because in the model, ice is treated as a land-surface, so keeping the ice cover unchanged keeps the model land-surface unchanged as well.

Another sensitivity study modified the roughness length over snow. The snow-dependent roughness length was replaced with background land-surface roughness length to examine what happens when no snow is present to reduce the roughness of the land surface. In the Noah LSM (Chen and Dudhia, 2001a), the presence of snow cover reduces the surface roughness to a low value, even for forests and urban land uses, Fig. 3.2. Thus the experiment was designed to investigate the influence of surface roughness induced mixing on the bands passing over the Upper Peninsula of Michigan. Details are provided in George (2008).

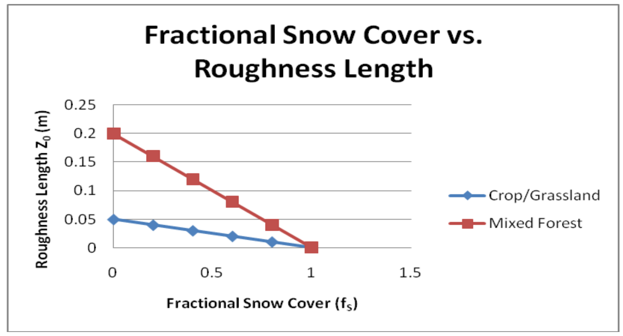


Figure 3.2. Fractional snow cover and roughness length plot for land use types crop/grassland and mixed forest. The reduction in roughness length due to an increase in snow cover is depicted.

Figure 3.3 shows the along-band cross-section taken from domain 3 shown in subsequent figures. George (2008) also shows additional cross-sections taken perpendicular to the bands along the southeast shoreline of Lake Superior and along the northwest shoreline of Lake Michigan to compare the boundary layer before and after interaction with the intervening land of the Upper Peninsula of Michigan.

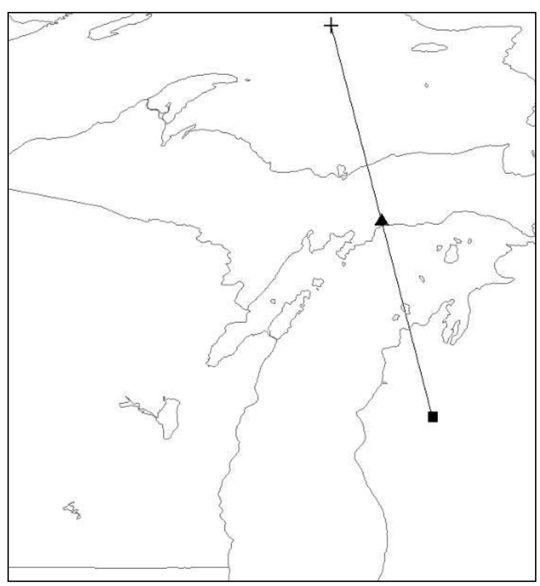


Figure 3.3. Location of Cross-section in model domain 3 for potential temperature and relative humidity plots.

4. RESULTS

4.1 *Model Verification*

To determine if the WRF-ARW model correctly portrayed each lake-effect event throughout the entire run times, a comparison of observations and model results was conducted. Sea level pressure, surface winds, surface air temperatures, and cloud cover were among the properties compared. Archived observations of sea level pressure, surface winds, and surface air temperatures used for model verification were collected from the National Climatic Data Center (NCDC) and the University Corporation for Atmospheric Research (UCAR) while archived observations of Terra Moderate Resolution Imaging Spectroradiometer (MODIS) satellite images were collected from the Space Science and Engineering Center (SSEC) at the University of Wisconsin-Madison and from UCAR.

One aspect of the comparisons was to verify if the events were accurately portrayed through placement of the low and high pressure systems. Sea level pressure and surface wind analyses at 21 UTC on 01 December, Fig. 4.1, indicate that the model results in image (a) are consistent with the observations in image (b). Placement of the isobars in the model corresponded well to placement of the isobars in observations in (b). For example, the 1024-hPa isobar in (a) and (b) each ran through central Lake Michigan and up through western Lake Superior. As for the winds, it is important to note that the model results are from domain 2, which had 5-km grid spacing, so the resolution of the modeled wind barbs was much finer than the resolution of the wind barbs on the UCAR archived observation map shown in Fig. 4.1(c). The model appears to have overestimated the wind speed by 3-5 m s^{-1} (5-10 knots) in general over the second domain region. However, gusts were noted frequently in the observations, so given that the model surface winds are instantaneous values at that time, at least part of the difference accounts for these gusts. Later in the period, at 18 UTC on 02 December, the consistency between the model and the observations continued, as shown in Fig. 4.2. In both image (a) and (b), the high pressure system was centered over west-central Wisconsin. The NCDC pressure observations indicate a slightly higher pressure at the center of the surface high than the model results, but the pressure locations matched relatively well. The wind barbs in

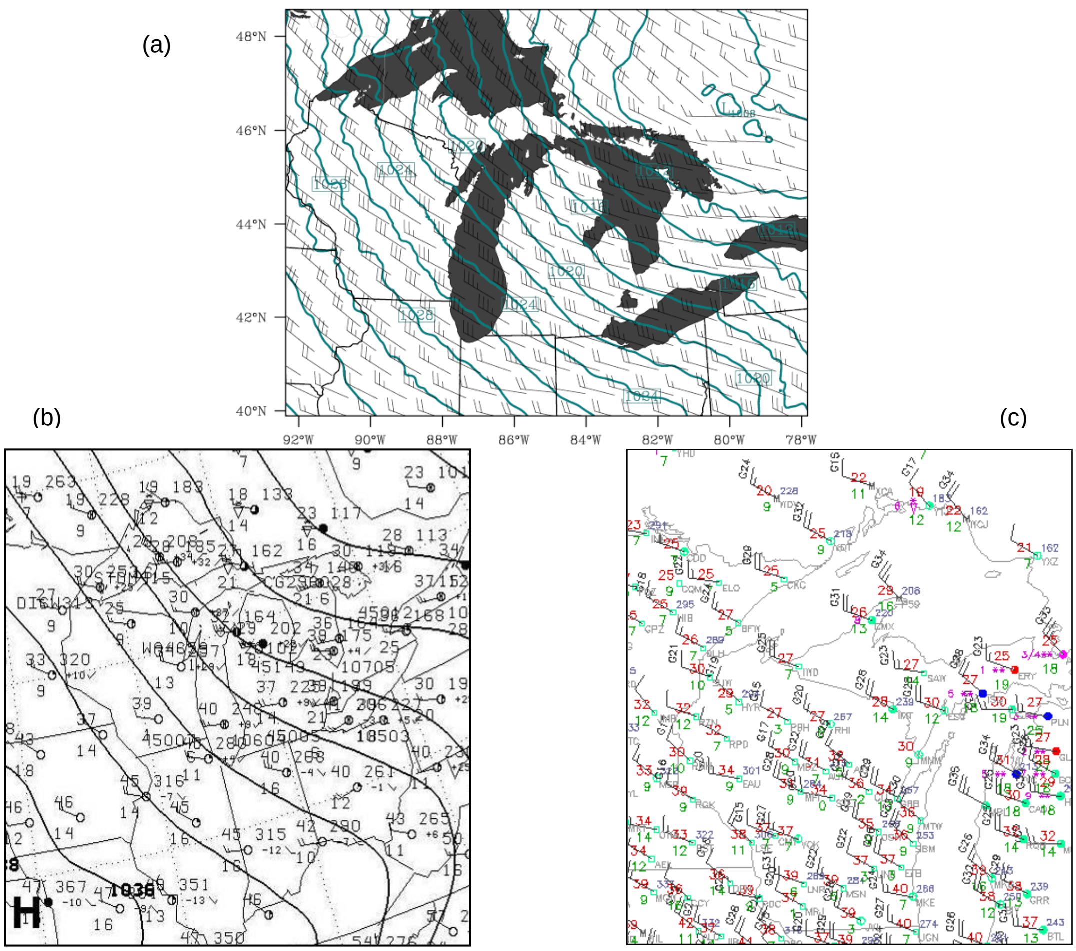


Figure 4.1. Comparison of (a) model analysis of sea level pressure and wind fields for domain 2 with (b) NCDC-observed sea level pressure values at 21 UTC and (c) UCAR-observed surface wind values at 2143 UTC on 01 December 2003. Image (a) has pressure contours every 2 hPa while (b) has pressure contours every 4 hPa. 10 kts is approximately equivalent to 5 m s^{-1}. Temperatures are shown in °F.

image 4.2(c) mostly matched to the model in image (a), with the exception of slight overestimation of wind speeds in the model over the state of Michigan. The consistency of the model with the observations continued for this case throughout the simulation, not just for the presented times, which are near initialization and near completion of the simulation.

The February base case also showed agreement between the model results and the observations. At 21 UTC on 14 February, the model in Fig. 4.3(a) and the observations in Fig. 4.3(b) both had the 1028-hPa isobar oriented north-to-south through central Wisconsin. To the east of this the isobars matched in location as well. The wind speeds as indicated by the barbs in image (a) appear to have been overestimated in domain 2 by about 3 m s^{-1} (5 knots) when compared to the observed wind values in Fig. 4.3(c), though these observations were taken over an hour later than the time of NCDC observations and the model. At 18 UTC on 15 February, in Fig. 4.4, the isobars were again found in similar locations, as the model results in (a) and the observations in (b) both show that the 1020-hPa isobar crossed through central Lakes Michigan and Superior.

The modeled wind in (a) more accurately simulated the gusts that were observed for this time, and as a result only a slight overestimation in the wind speed was seen in a comparison of Fig. 4.4(a) with the observations in Fig. 4.4(c). The model also compared well with the observations at intermediate times. The figures for both December and February support the fact that the model had successful simulations in each case. Other surface level comparisons also indicated accurate simulations by the model.

One of the most crucial elements of the simulations is the replication of the clouds over the innermost domain. Since the Terra satellite is polar orbiting, Terra MODIS visible satellite images are only available about once a day over the Great Lakes region. Therefore, high-resolution, close-up images of cloud observations could only be compared to modeled clouds at one time during each of the events. The remainder of the events had to be compared to geostationary satellite images. Figure 4.5 shows the model results for domain 3 at 17 UTC in (a) compared to the Terra MODIS satellite image at 1715 UTC in (b) for 02 December. Likewise, Fig. 4.6 shows the model results for domain 3 at 16 UTC in

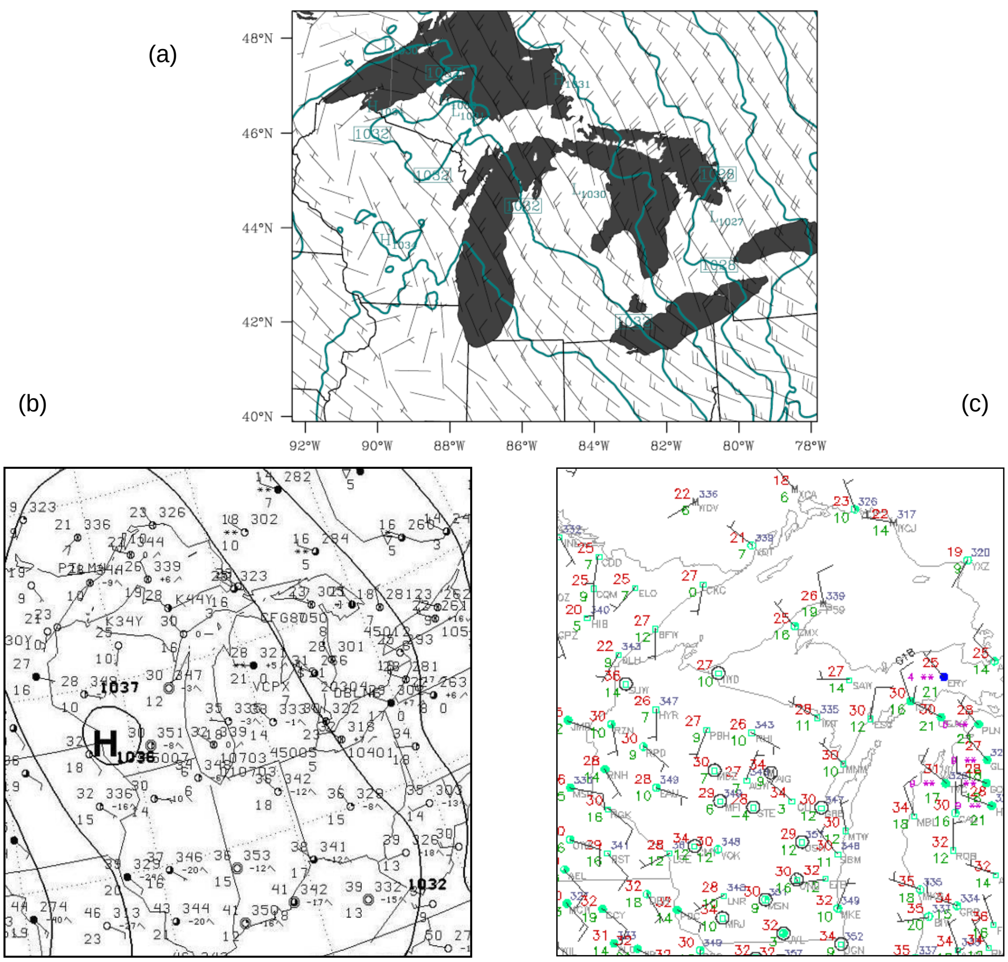


Figure 4.2. As in Fig. 4.1, but for (a) and (b) at 18 UTC and (c) UCAR-observed surface wind values at 1843 UTC on 02 December 2003.

image (a) compared to the Terra MODIS satellite observations at 1615 UTC in (b) for 15 February. Modeled clouds are displayed as cloud water mixing ratio isosurfaces of 0.05 g kg^{-1}. This value was chosen due to the adequate amount of cloud structure displayed with it. The case of 15 February 2007 had more snowfall than the December case; therefore, the clouds in the satellite observation for February blended into the snowy surface. However, a close look revealed that both the L2L bands in December and the lake-effect clouds in February were reproduced relatively accurately. Consideration must be taken in the resolution differences between the satellite and the model; the Terra MODIS visible satellite images have 250-m resolution while domain 3 in the model had 1.66-km resolution and wavelengths in the model are aliased to more than 3.3 km, therefore more detail was seen in the satellite images. Geostationary visible satellite images from UCAR were used for comparison against the modeled clouds for other times during each event. The daytime hours of 02 December 2003 appeared to show correct location and structure of L2L bands based on model analysis of cloud field. It was initially thought that the February case was non-L2L. However, L2L bands developed around 18 UTC on 15 February, after the high-resolution image from Terra MODIS. Figure 4.7 shows model analysis of cloud water mixing ratio at 21 UTC 15 February 2007 in (a) and a corresponding visible satellite image at 2115 UTC in (b). The cloud isosurface of mixing ratio 0.05 g kg^{-1} shows the same bands of clouds over southern Lake Superior as well as the Upper Peninsula and northern Lake Michigan as seen on the visible satellite observation. Results were similar for 18 UTC to 20 UTC as well. To further verify the accuracy of the clouds, another simulation of the February case was completed. To test the possibility that the YSU boundary layer scheme had a tendency to over-produce L2L bands, the boundary layer scheme was changed from the YSU to the Mellor-Yamada-Janjic (Eta) TKE scheme. Correspondingly, the surface-layer option was changed from the default Monin-Obukhov scheme to the Janjic Eta scheme, since the surface-layer options are tied to particular boundary layer options in the WRF model. The results, which are presented in George (2008), indicated that the boundary

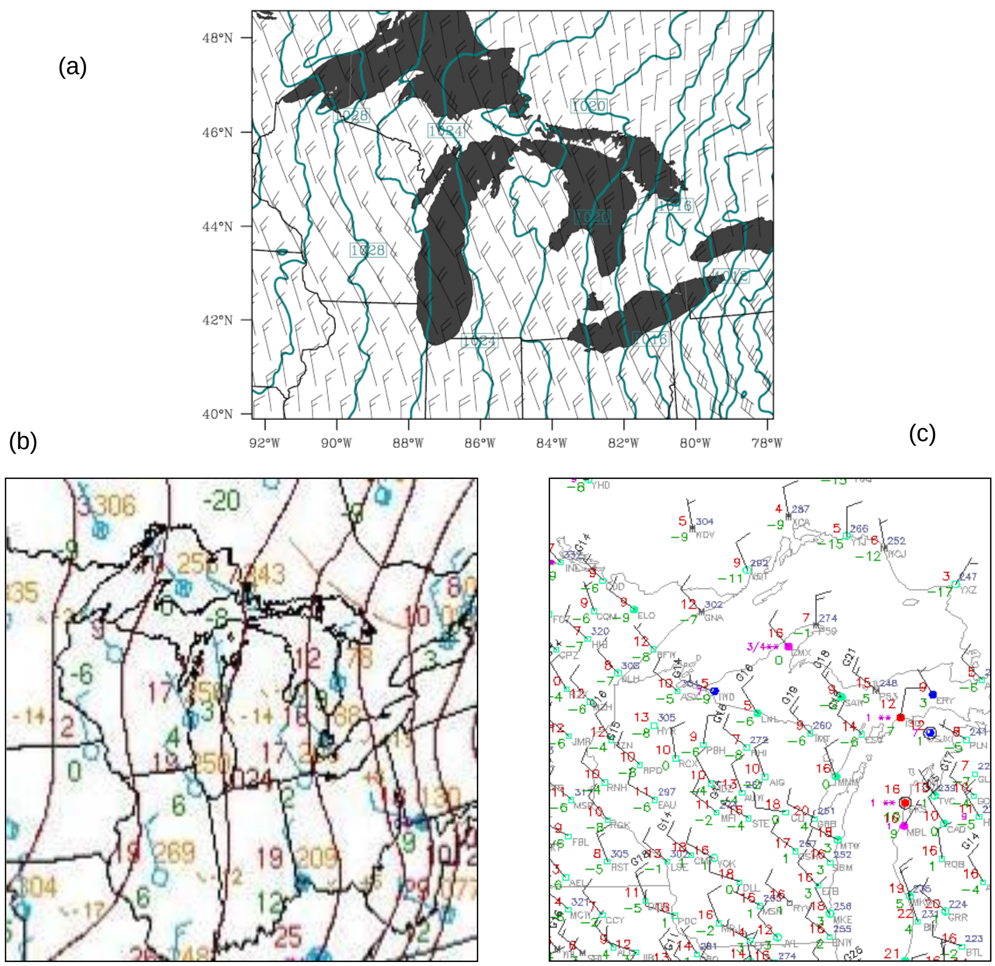


Figure 4.3. As in Fig. 4.1, but for (a) and (b) at 21 UTC and (c) UCAR-observed surface wind values at 2216 UTC on 14 February 2007.

layer scheme did not have a significant impact on the type of clouds produced; L2L bands formed in both boundary layer schemes and closely resembled each other. These two verification processes show that L2L bands did form around 18 UTC on 15 February 2007 in both the model and the observations. This will be considered in the comparison of the two events.

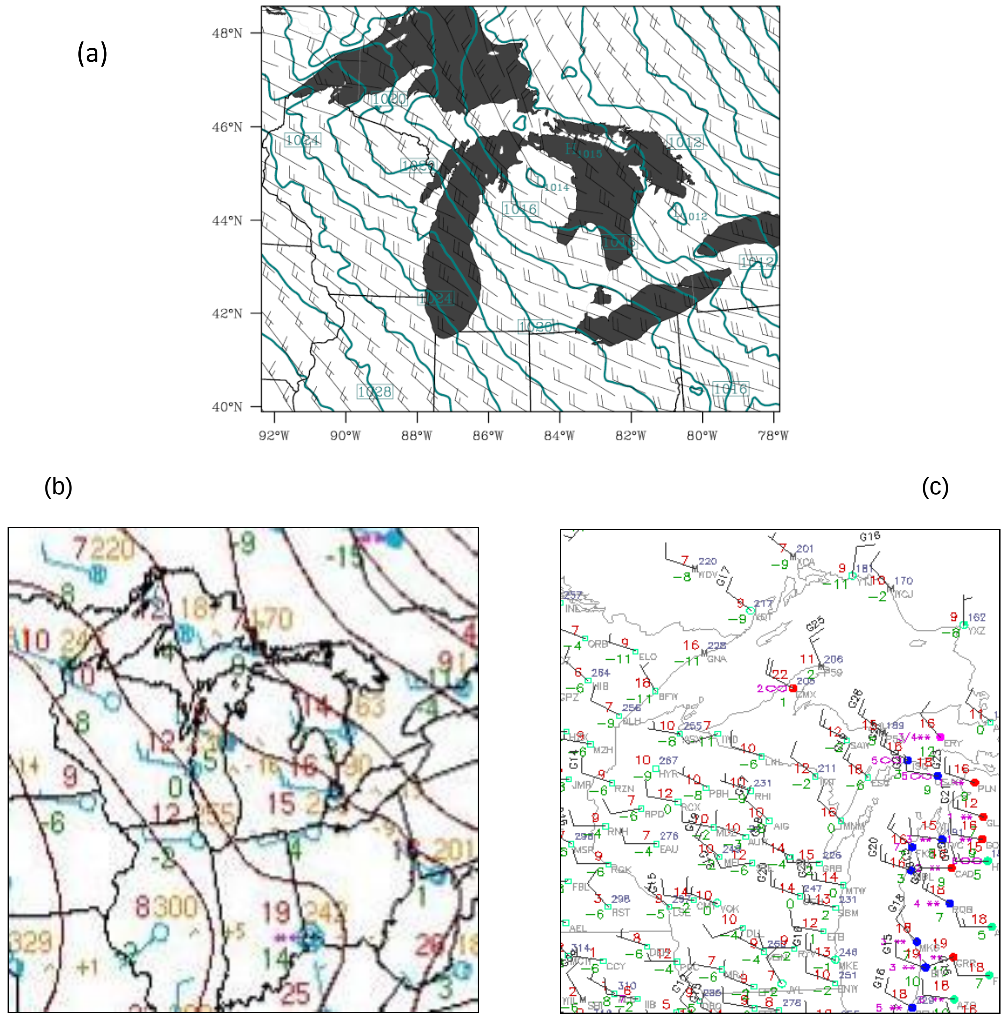


Figure 4.4. As in Fig. 4.1, but for (a) and (b) at 18 UTC and (c) UCAR-observed surface wind values at 1816 UTC on 15 February 2007.

(a) (b)

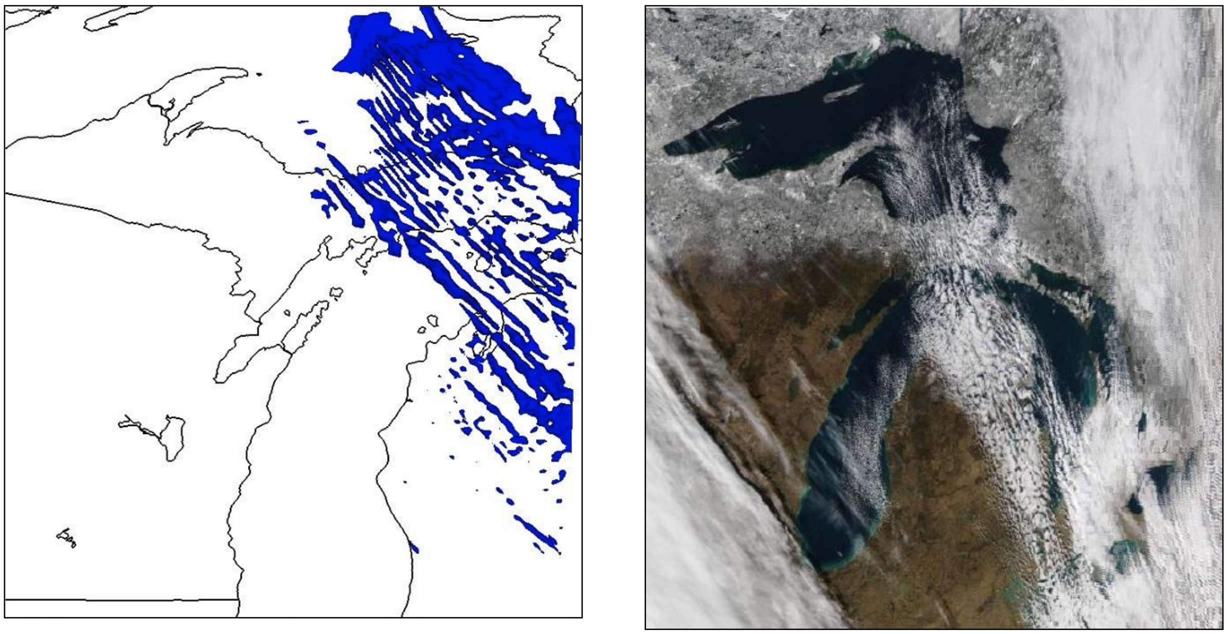


Figure 4.5. Cloud comparison of (a) model output for domain 3 at 17 UTC and (b) Terra MODIS satellite observations at 1715 UTC on 02 December 2003. The cloud isosurface in (a) shows a cloud water mixing ratio of 0.05 g kg^{-1}

(a) (b)

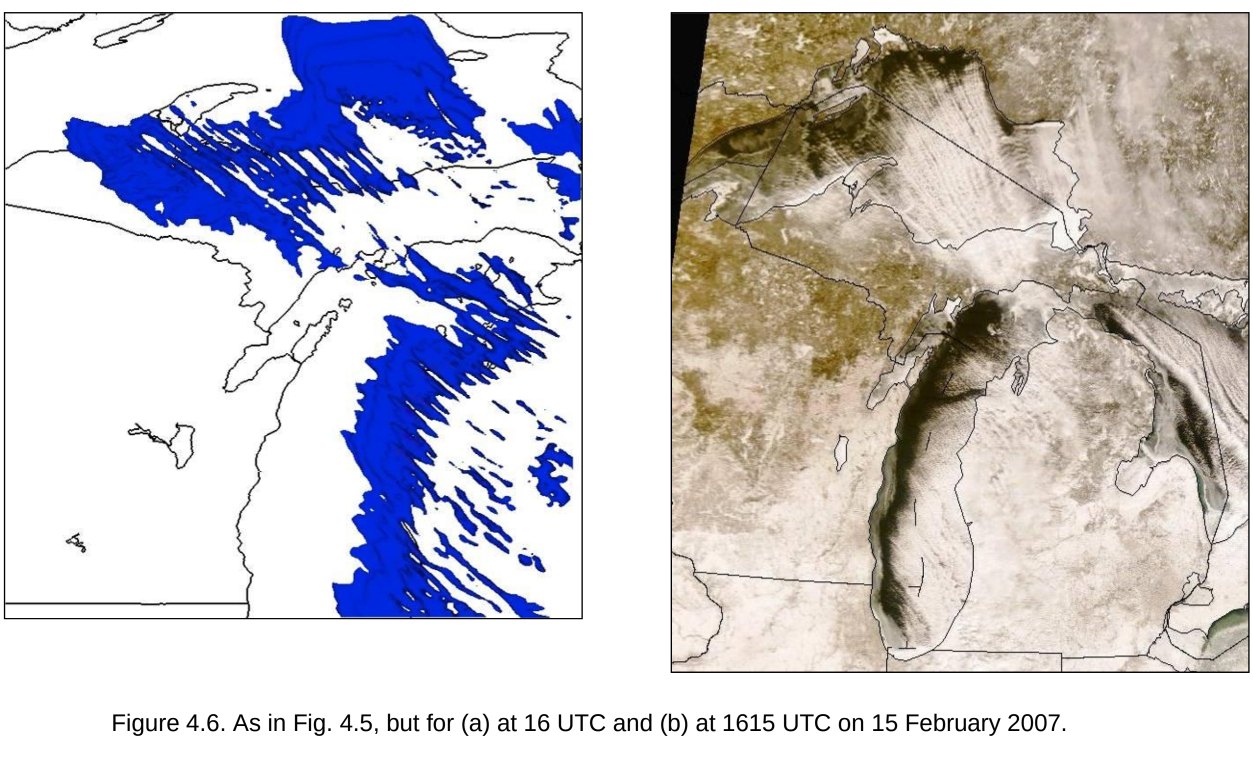


Figure 4.6. As in Fig. 4.5, but for (a) at 16 UTC and (b) at 1615 UTC on 15 February 2007.

(a) (b)

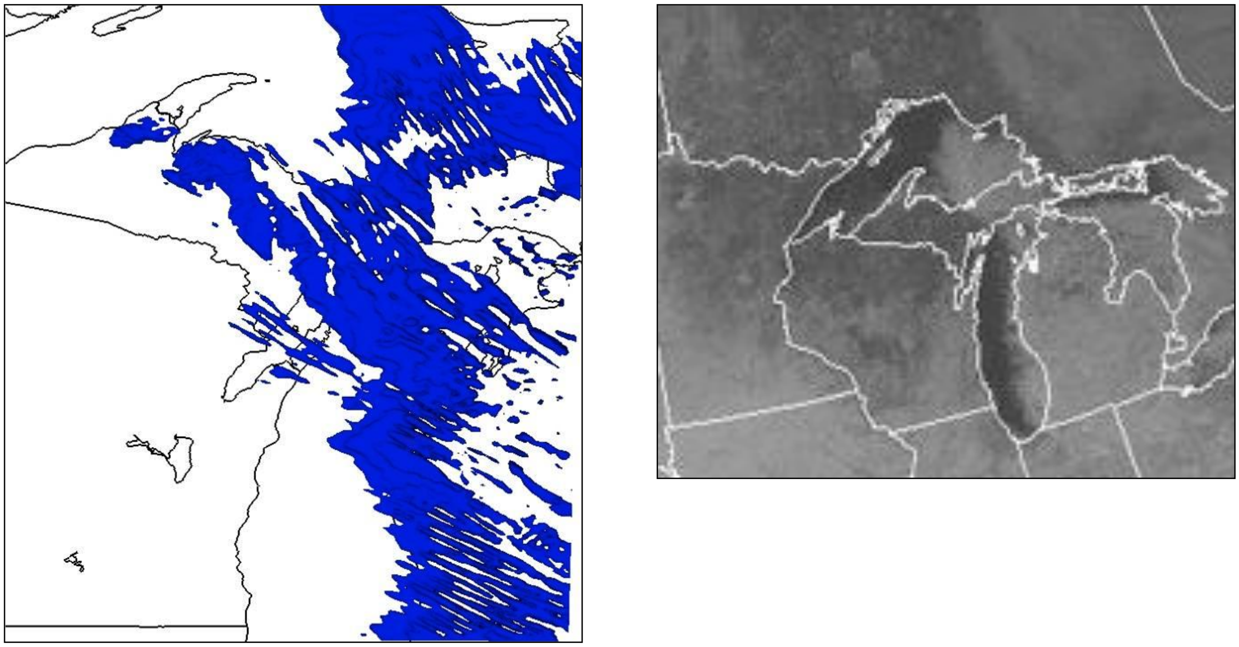


Figure 4.7. Cloud comparison of (a) model output for domain 3 at 21 UTC and (b) UCAR visible satellite observations at 2115 UTC on 15 February 2007. The cloud isosurface in (a) shows a cloud water mixing ratio of 0.05 g kg^{-1}.

4.2 ***Simulations for 02 December 2003***

4.2.1 Base Simulation for 02 December 2003

The base simulation for the L2L case of 02 December 2003 was examined in order to determine differences between the boundary layer upwind of Lake Superior versus downwind of Lake Superior.

The L2L case of 02 December 2003 started with lake-effect clouds developing within the northwesterly flow ahead of a surface high pressure system. Wind speeds within this flow were averaging 15 m s^{-1} (30 knots). Based on cloud water mixing ratio isosurface images from domain 3 of the model analysis, L2L bands began to form from Lake Superior to Lake Michigan around 00 UTC on 02 December. Earlier in the event, at 21 UTC 01 December, northwesterly winds across the Great Lakes region brought cold air across Lake Superior and preconditioned the boundary layer air downwind of the lake, or over the Upper Peninsula of Michigan and onward to Lake Michigan and Lower Michigan. In Fig. 4.8, several key features can be noted that reveal processes in the boundary layer. In this figure, and as reference for all remaining figures of cross-sections taken along the bands, grid points 0 to 90 represent northeastern Lake Superior to the Upper Peninsula of Michigan, grid points 90 to 140 represent the pass over the Upper Peninsula, grid points 140 to 200 represent northern Lake Michigan, and grid points 200 to the end represent northern Lower Michigan. Potential temperature is plotted in 1°K incremental black contours and relative humidity is plotted every 10% in color-filled contours. Prior to band formation (not shown), the top of the well-mixed boundary layer grew to approximately 2 km. Within the well-mixed boundary layer, a cold-to-warm gradient existed from Lake Superior to Lower Michigan as cold northwesterly wind flowed across the region. As a result of heating of near-surface air over Lake Superior, a superadiabatic lapse rate developed. Less than saturated conditions existed in the region downwind of the Upper Peninsula, and a much shallower cloud layer occurred over Lake Michigan than over Lake Superior. Thus the intense convective mixing and cloud layer over Lake Superior was not maintained as it crossed Upper Michigan. In the stable air above the boundary layer, gravity waves were present throughout the two cases and their respective sensitivity simulations.

AT 00 UTC 02 December, as the L2L bands were developing, boundary layer heights were slightly higher, especially over northern Lower Michigan (see Fig. 4.8a). L2L bands were present over the western Upper Peninsula—mainly from the Keeweenah Peninsula—down to northern Lake Michigan and Lower Michigan. Lake-effect bands were located over northeastern Lakes Superior and Michigan, but these were discontinuous over the land between them. Potential temperatures of boundary layer air reaching the Upper Peninsula and Lower Michigan were cooler at this time by as much as 3°K. The lapse rate was still superadiabatic over Lakes Superior and Michigan, with an even stronger superadiabatic lapse rate over Lake Superior. Wind speeds in the western Great Lakes region were about 15 m s^{-1} (30 knots) from the northwest. In Fig. 4.8(a), there was a slot of drier air in the cloud layer over the Upper Peninsula as the cross-section passed through a cloud-free section. However, a much deeper layer of saturated air and clouds continued across the land and over Lake Michigan. Close to the surface of the Upper Peninsula, lower valued contours of potential temperature marked colder and more stable air from the colder land surface.

Overnight on 02 December, the lake-effect bands became oriented northwest to southeast, instead of the west-northwest to east-southeast direction of the bands at the beginning of the event. As this occurred, the L2L bands began forming farther eastward over southeastern Lake Superior and the eastern Upper Peninsula to northeastern Lake Michigan and northern Lower Michigan. While the number of bands decreased in coverage, L2L bands still formed, along with lake-effect clouds forming over the western shore of Lower Michigan down towards the border with Indiana. During this time, the height of the boundary layer decreased to about 1.6 km as it was suppressed by the lowering subsidence inversion. Nearly constant temperature cool air was seen in the lower boundary layer from the Upper Peninsula down to northern Lower Michigan. Surface wind speeds decreased slightly over land to about 3-8 m s^{-1} (5-15 knots), but remained about 10-15 m s^{-1} (20-30 knots) over the open water. Figure 4.8(b) is representative of the overnight boundary layer. Here, even at 12 UTC 02 December, despite having seen a superadiabatic lapse rate and continued mixing in the lower boundary layer overnight, the cloud layer remained shallow. A break in the clouds occurred over the northernmost part of Lake Michigan. Increased stability aloft was noted as the potential temperature contour gradient increased. This suggests that the subsidence inversion was lowering because of the impeding high pressure system, which compressed the boundary layer. Cooling of the air by the land-surface of the Upper Peninsula was noted as well from the potential temperature contours downwind of that region. The height of the boundary layer also appeared higher at the intersection of the Upper Peninsula of Michigan, as noted by the upside-down "U" shape in the tightly packed potential temperature contours in Fig. 4.8(b). Following 12 UTC 02 December, surface winds over the Great Lakes region began to weaken significantly to 8 m s^{-1} (15 knots) and less; they remained weak for the remainder of the event.

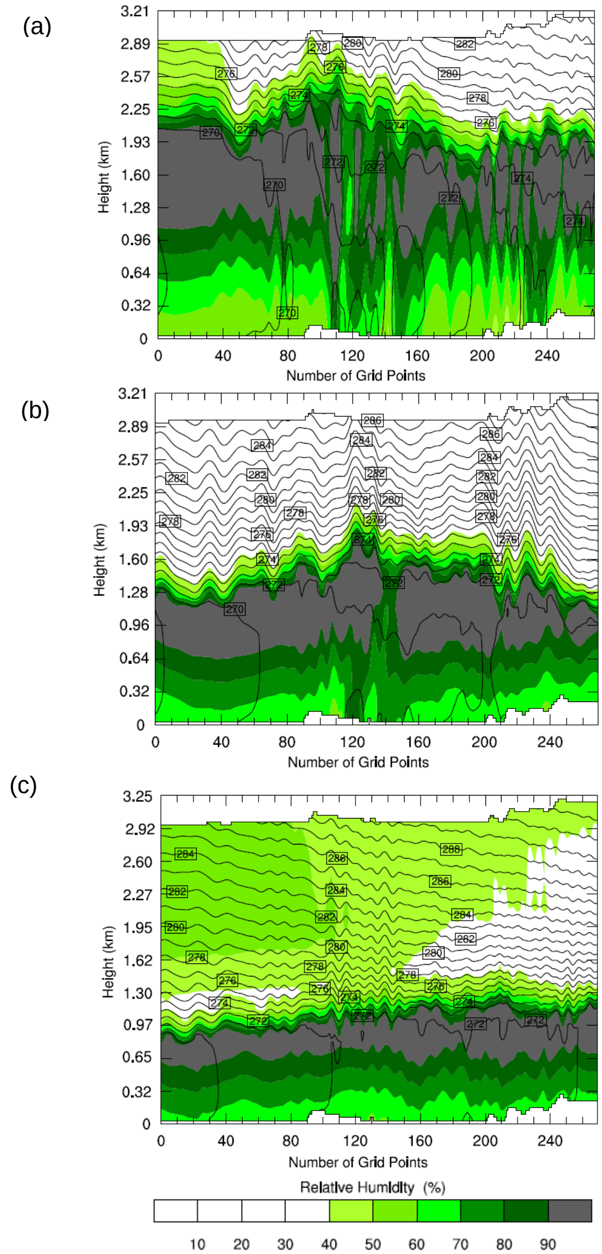


Figure 4.8. Along-band cross-sections of potential temperature (solid black lines, every 1°K) and relative humidity for (a) 00 UTC 02 December 2003, (b) 12 UTC 02 December 2003, and (c) 21 UTC 02 December 2003. Image (a) represents conditions at time of formation of L2L bands, image (b) shows the boundary layer conditions that were typical of the overnight hours from 01 December to 02 December, and image (c) shows boundary layer conditions typical until the end of the event at 06 UTC 03 December 2003. For all along-band cross-sections, grid points 0 to 90 represent Lake Superior, grid points 90 to 140 represent the Upper Peninsula of Michigan, grid points 140 to 200 represent northern Lake Michigan, and points 200 to the end represent northern Lower Michigan.

As the day progressed on 02 December, more L2L bands were produced over the eastern Upper Peninsula, northern Lake Michigan, and farther inland over north-central Lower Michigan. Figure 4.8(c) shows the boundary layer cross-section at 21 UTC 02 December. The boundary layer conditions seen at this time remained similar until the end of the event at 06 UTC on 03 December. Despite the increase in L2L bands after the overnight hours, the well-mixed boundary layer decreased in height and the clouds were shallower than earlier in the event. This was likely due to light winds and subsidence aloft from the surface high pressure system, which was located in the vicinity of Chicago, Illinois, at 21 UTC 02 December. Overall, the near-surface air that reached Lower Michigan was 5°K cooler than it had been at the start of the event at 21 UTC 01 December. Superadiabatic lapse rates existed until 06 UTC 03 December, when the L2L event ended with the presence of the surface high pressure system over western Lower Michigan.

Prior to and throughout this event, the mixing in the boundary layer was enhanced by Lake Superior and was sustained over the Upper Peninsula of Michigan and beyond to Lake Michigan and Lower Michigan. In addition to the boundary layer growth seen in the cross-sections along the L2L bands, growth can be visually noted in the cloud band structure. In the satellite photo shown in Fig. 4.5, shallow and narrow cloud bands are seen over Lake Superior. As the bands evolved downwind, the clouds thickened and the bands increased in wavelength. This is especially visible over northern Lower Michigan.

Cross-sections across the bands before and after crossing the Upper Peninsula provided further evidence that the boundary layer modifications by Lake Superior were not interrupted by the Upper Peninsula (George, 2008). The boundary layer held its mixing, moisture, and heat from the influences of Lake Superior as air passed over the Upper Peninsula. Although fewer cloud bands were noted downshore of the Upper Peninsula, relative humidity values remained high and potential temperatures remained near constant in the lower boundary layer. Only a slight decrease in the boundary layer height occurred as the air passed over the Upper Peninsula.

4.2.2 Sensitivity Studies for 02 December 2003

The first sensitivity study was to examine the effect of snow cover on the land-surface roughness length. As discussed in Section 3, snow cover on a land-surface in the WRF model significantly lowers the roughness length value, so much that even forested or urban regions appear as though they are not there. The change incorporated for this study was to maintain the true roughness length of the appropriate landuse region in the model. This is important since the majority of the Upper Peninsula and northern Lower Michigan is forested. This case had little snow cover at initialization and only a small amount of snow cover was created by the L2L bands. Thus only small differences in roughness were indicated and virtually no effect was visible in the L2L event.

The second sensitivity study to be discussed is the increase or decrease of 5°K to the surface temperatures of the western Great Lakes in domain 3 of the simulation. These simulations will hereafter be referred to as Plus-5 and Minus-5. Following the temperatures in Fig. 2.2, the 5°K increase corresponded to an approximate increase from 277°K to 282°K over the majority of Lake Superior and an increase from 278°K to 283°K over the majority of Lake Michigan. The Minus-5 case corresponded to changes from about 277°K to 272°K over most of Lake Superior and 278°K to 273°K over the majority of Lake Michigan. Since the WRF-ARW version 2.2 model uses a water temperature of 271°K to delineate ice, the temperature decreases created in the Minus-5 run did not create ice over the lakes.

The most significant response to the Plus-5 temperature change was noted on around 18 UTC on 02 December and through the rest of the event, as band formation increased southward along the western shore of Lower Michigan as shown on the left in Fig. 4.9. It is important to note, however, that these new bands were not L2L bands. Within the L2L bands, cloud bands (and depths) were thicker. For the Minus-5 simulation, the cloud bands appeared very similar to those from the base simulation for the first twelve hours of the simulation. Notable changes first occurred after 12 UTC on 02 December. The right panel of Figure 4.9 shows the cloud field for the Minus-5 simulation. In comparison with the base simulation, a reduction in the number of bands over land masses is apparent. Locations of the bands remained similar, except the western-most L2L bands over northern Lake Michigan that appeared in the base simulation became broken apart in the Minus-5 simulation. The L2L bands diminished more quickly as well, at 00 UTC on 03 December instead of 06 UTC for the base simulation.

Figure 4.10 shows an example of the changes to the boundary layer that occurred throughout the event with the warmer and colder lake surfaces. The cross-section on the left is the Plus-5 simulation, while the cross-section on the right is the base simulation. The 5°K increase to lake surface temperatures caused more moisture and heat to be generated into the near-surface air over the lake, which in turn was carried downwind over land. Along with thicker cloud bands for the Plus-5 case, the well-mixed lower boundary layer extended higher, especially at the intersections with the Upper Peninsula. Also at the Upper Peninsula, cold and stable air was noted near the surface. Since this was more stable than the base simulation over the Upper Peninsula, there was a larger effect on circulations in this region. The superadiabatic lapse rate over the Lakes was stronger for the Plus-5 case than for the base case. The cross-section for the minus-5 simulation is shown in the lower panel. Within the boundary layer, it was noted throughout the period that colder surface air temperatures reached Lower Michigan and the well-mixed layer was shallower. The cloud layer appeared shallower than the base case cloud layer as well after 18 UTC 02 December. Not only was the surface air 2°K colder than the base temperatures over Lower Michigan, but the near-surface air temperature was about 4°K cooler than at the start of the event. In general, the cooler lake temperatures did not have much of a negative effect on moisture availability from the lake but did have a negative effect on thermal fluxes from the lake, which in turn decreased boundary layer heights and cloud thickness.

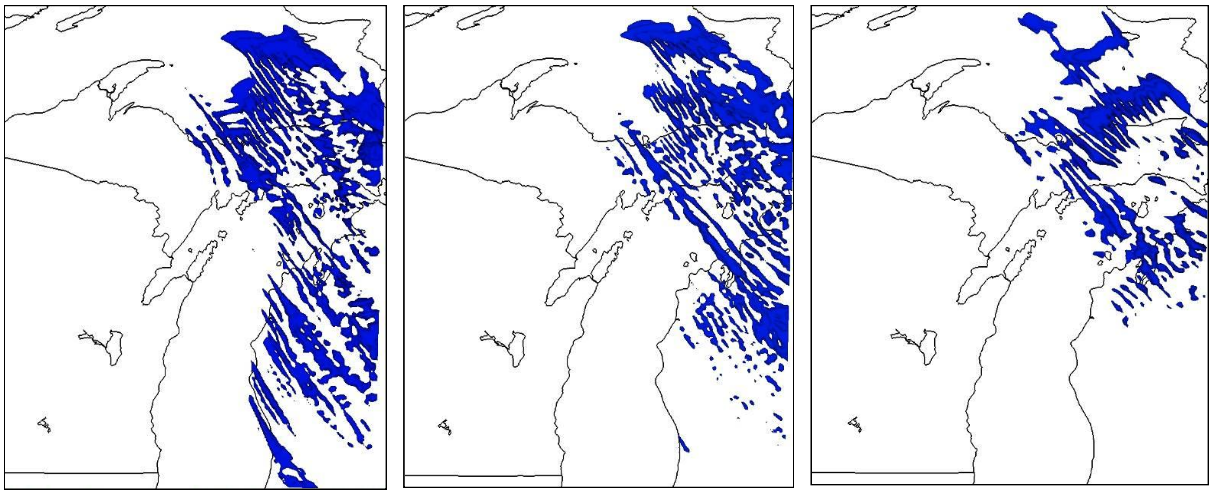
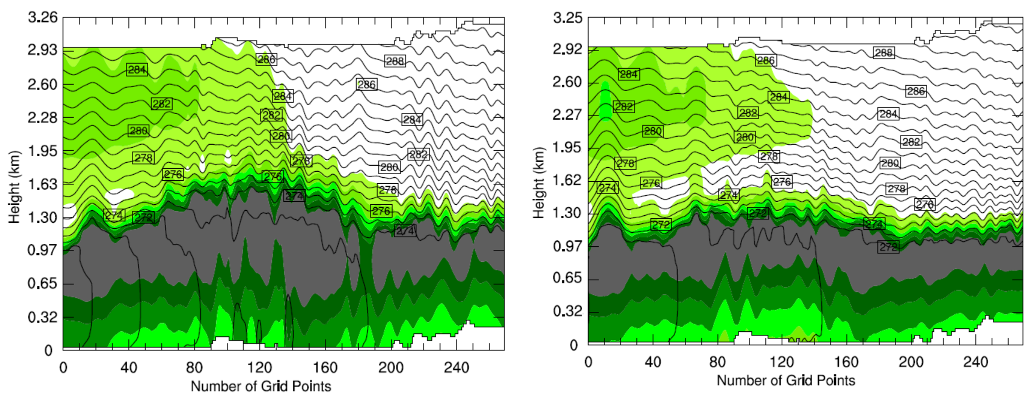


Figure 4.9. Comparison of cloud model analysis for the December Plus-5 simulation (left), base simulation (center), and Minus-5 simulation (right) at 18 UTC on 02 December 2003. Cloud isosurfaces have a mixing ratio of 0.05 g kg^{-1}.



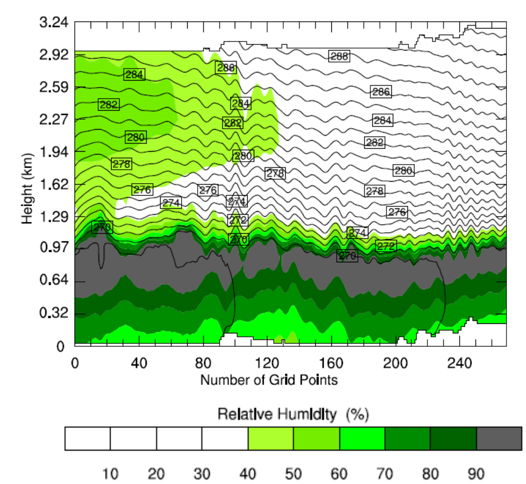


Figure 4.10. Boundary layer impacts at 18 UTC, 02 December 2003 due to Plus-5 and Minus-5 lake surface temperature change. Along-band cross-sections of potential temperatures (solid black lines, every 1°K) and relative humidity are shown. The Plus-5 case is to the above left, the base case is to the above right, and the Minus-5 case lower.

4.3 *Simulations for 15 February 2007*

4.3.1 Base Simulation for 15 February 2007

The second event of 15 February 2007 was chosen as a non-L2L event to compare with the L2L case. In verification of cloud field analysis from the model with satellite observations of clouds, it was discovered that this event began with typical lake-effect convection and then developed L2L bands late in the case. In response to this discovery, the base simulation for February was studied to determine the meteorological changes that occurred when L2L band formation began. This case was also compared against the L2L December case to determine differences between the entire December L2L case and the non-L2L part of this event along with the L2L part of this event, at their respective times. The same cross-sections shown in Fig. 3.3 were used.

At the start of the simulation, lake-effect clouds had already formed over southern Lake Superior and the majority of Lake Michigan in the northwesterly flow. Wind speeds around the western Great Lakes region were generally at 10 m s^{-1} (20 knots). The boundary layer conditions remained quite similar through the beginning hours of the event as illustrated in Fig. 4.11(a) at 21 UTC on 14 February 2007. The warmth of the lake and the cool northwesterly flow created superadiabatic lapse rates over Lakes Superior and Michigan. A deep well-mixed layer existed up to about 1.6 km, though a decrease was noted over inland Lower Michigan. No bands were seen over the Upper Peninsula of Michigan; therefore the colder potential temperature contours seen over this region were a result of the cold land mass. Cold air was noted flowing in from the northwest, as evident in the cold-to-warm potential temperature gradient in the boundary layer from Lake Superior to Lower Michigan.

The overnight hours of 15 February caused a large reduction in the lake-effect clouds seen over Lake Michigan. The height of the boundary layer did not diminish at all, as seen at 12 UTC on 15 February in Fig. 4.11(b). Except for cooler air associated with the Upper Peninsula, temperatures were relatively constant in the lower boundary layer over the cross-section region. Wind speeds also remained comparable to those seen in the beginning of the event. Lapse rates over the western lakes remained superadiabatic as well. Given that the clouds diminished while boundary layer conditions remained favorable for further cloud production, geostationary infrared satellite images from UCAR were used to verify the reduction in clouds (not shown). No evidence was found to contradict the model cloud analysis.

At approximately 18 UTC on 15 February, wind speeds over Lakes Superior and Michigan increased to a range of 13-15 m s^{-1} (25-30 knots) and L2L bands developed. The greatest L2L band development occurred around 21 UTC, as seen in Fig. 4.7. Figure 4.12 shows an along-band cross-section of potential temperature and relative humidity for 21 UTC on 15 February. This figure demonstrates that during the L2L part of the event, boundary layer heights were lower over the lakes. In the lower boundary layer, or in the well-mixed layer, the temperature that reached Lower Michigan from upwind Lake Superior was about 1°K warmer than at the beginning of the case. Very high relative humidity values were located near the surface during this time as well. Definite boundary layer growth was noticeable over Lake Superior downwind to Lower Michigan. A stronger superadiabatic lapse rate was noticeable too at this time in the figure. Cloud depths, however, remained relatively consistent throughout the entire event. L2L bands continued until 06 UTC 16 February, when all lake-effect convection ended. Up until this time, the depth of the mixed boundary layer became shallower, to about 0.9 km, even though winds continued to flow westerly at 10-15 m s^{-1} (20-30 knots) over the region. The surface high pressure system was still located well to the southwest of the Great Lakes, over the proximity of Kansas.

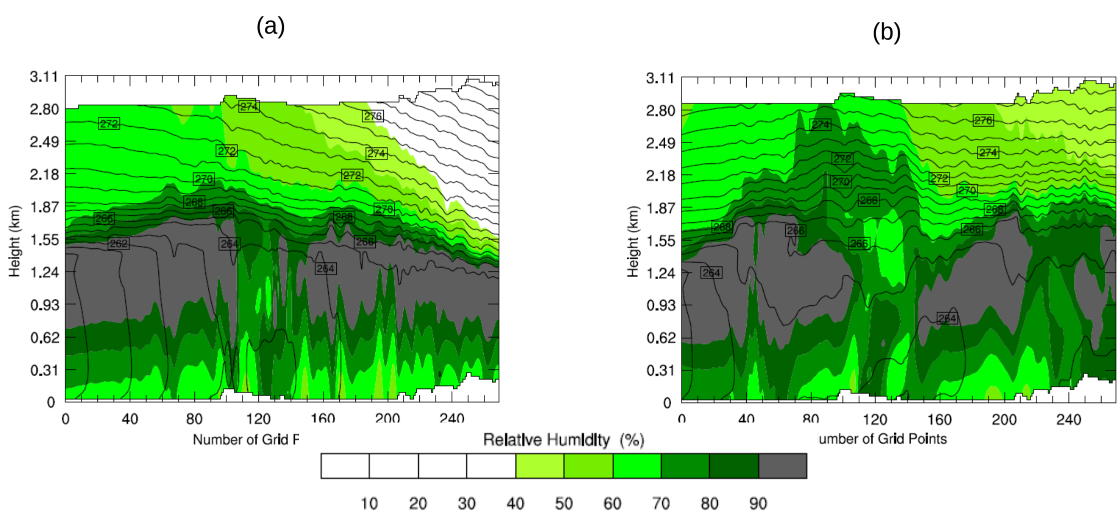


Fig. 4.11. As in Fig. 4.8, but for (a) 21 UTC 14 February 2007 and (b) 12 UTC 15 February 2007. Image (a) shows the boundary layer conditions at the beginning of the event, when typical lake-effect clouds were in place. Image (b) shows the boundary layer conditions that were typical of the overnight hours from 14 February to 15 February.

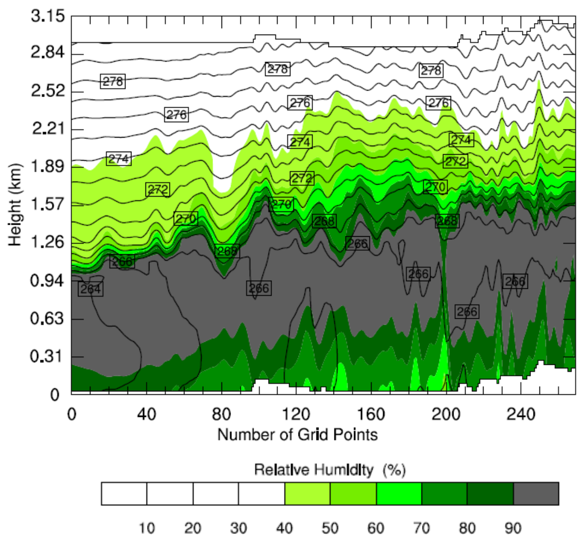


Figure 4.12. Along-band cross-section of potential temperature (solid black lines, every 1 K) and relative humidity for 21 UTC 15 February 2007. This is representative of the L2L portion of the event.

Cross-sections of potential temperature and relative humidity along the north and south borders of the Upper Peninsula presented by George (2008) show relatively continuous moisture and heat transport across the Upper Peninsula. The boundary layer height was suppressed slightly by the passage over the landmass. However, these results further indicated that the modifications to the boundary layer by Lake Superior were able to continue across land to downwind Lake Michigan and Lower Michigan with little interruption.

4.3.2 Sensitivity Studies for 15 February 2007

The same three sensitivity tests utilized for the December case were applied to the February case as well. This case had snow on the ground prior to the start of the simulation, so much more snow was present to affect the roughness length variable in this case than in the December case. Again, the roughness length variable was compared to the roughness length in the base simulation. The roughness lengths in the western Upper Peninsula were decreased in the base simulation by 50 percent or more. Snow cover even impacted cities, with the largest heights in the domain, with a very large underestimation of the roughness in the base simulation by about 0.46 m in the Milwaukee metropolitan area, located farther down the southeastern shoreline of Wisconsin. However, the reduction to roughness length had little impact on the formation of L2L bands. The cloud fields of the sensitivity run were virtually identical to the base simulation at all times. Cross-sections of potential temperature and relative humidity in the boundary layer for the region (not shown) also proved that little change occurred to the boundary layer as a result of the changes in roughness length.

The effect of changes in lake temperature on the simulated clouds are shown in Fig. 4.13. The 5°K increase to lake surface temperatures did not have a significant impact on the L2L bands. Lake surface temperatures were increased from 275°K to 280°K over most of Lake Superior and from 276°K to 281°K on average for Lake Michigan. Overall not many differences were determined to exist between the base simulation and this Plus-5 simulation cloud field analysis. Small differences were first noted at 00 UTC on 15 February, with a decrease in the Plus-5 case. This decrease took place mainly over Lake Michigan and the shoreline of Lower Michigan. Similar decreases for these locations occurred through 12 UTC on 15 February. At the locations where L2L bands formed at 18 UTC 15 February from south-central Lake Superior to northern Lake Michigan, L2L bands did not form in the Plus-5 case. L2L bands formed three hours later at 21 UTC. Fewer bands, however, were noticeable over northern Lake Michigan at this time for the Plus-5 sensitivity run compared to the base run. Thus, the increased temperature difference appears to hinder L2L bands. For the Minus-5 simulation, lake surface temperatures were decreased from 275°K to 270°K over central Lake Superior and from 276°K to 271°K over most of Lake Michigan. This is about 2°K less than the Minus-5 simulation for December. These temperatures are also right at the lake freezing temperature threshold of 271°K for the model, though additional ice was not allowed to form in this simulation. Near the beginning of the model run, slightly fewer cloud bands appeared over the western Great Lakes. Figure 4.13 shows an example of this at 12 UTC on 15 February 2007. This continued until approximately 18 UTC on 15 February, when the base cloud field and Minus-5 cloud field became similar. L2L band locations were the same for each case. Bands diminished over the waters once again at 00 UTC on 16 February and fewer clouds remained in the Minus-5 case over both the lakes and the land by the end of the simulation.

Although there was a tendency for fewer bands in the Plus-5 simulation, potential temperature and relative humidity cross-sections along the bands revealed that the bands were still thicker than the base simulation clouds. This was seen beginning at 00 UTC 15 February. Temperatures were comparable between the two cases until 12 UTC on 15 February, when warmer temperatures aloft were seen over Lake Michigan and northern Lower Michigan for the Plus-5 sensitivity case. This was also when the wind speed began to increase. As seen in Fig. 4.14, temperatures were only warmer by a degree or two. Temperatures became similar again around 00 UTC 16 February. Figure 4.14 also depicts a thicker cloud layer and similar mixed boundary layer heights. The Plus-5 simulation showed superadiabatic lapse rates consistent with the base simulation for the duration of the simulations. For the Minus-5 simulation, despite the similarities in cloud band structure, potential temperature and relative humidity cross-sections did reveal that boundary layer modification had occurred. Although a thinner cloud layer was not evident, the well-mixed boundary layer was not as deep for the Minus-5 run as it was for the base run throughout the period.

This was seen for both the non-L2L period, shown in 4.14 (lower) and the L2L part of the event, as shown in Fig. 4.15. During the non-L2L period, as shown in Fig. 4.14, cooler surface-to-850 hPa potential temperatures coincided with the lower inversion height in the Minus-5 case. The temperatures at this time were not only colder than the base simulation temperatures, but were also at least 2°K cooler than earlier in the period. Likewise, during the L2L part of the event at 21 UTC on 15 February (shown in Fig. 4.15), boundary layer mixing extended to lower heights for the Minus-5 case (shown on the left) than for the base case (on the right in Fig. 4.14). In the Minus-5 cross-section at 21 UTC, the dry slot prior to the passage over the Upper Peninsula corresponds to a small break in the clouds over this region. Near-surface temperatures at this time in each of the two simulations warmed slightly compared to 12 UTC, especially in the right half of the cross-sections over the Upper Peninsula and Lake Michigan to Lower Michigan, as the wind shifted to a more westerly component. Over Lake Superior, the colder lake surface temperatures were indicated by the lower potential temperature contours at this location. Superadiabatic lapse rates were also seen over Lake Superior for this sensitivity study. Following 21 UTC on 15 February, the boundary layer heights decreased by equal amounts in each of the simulations.

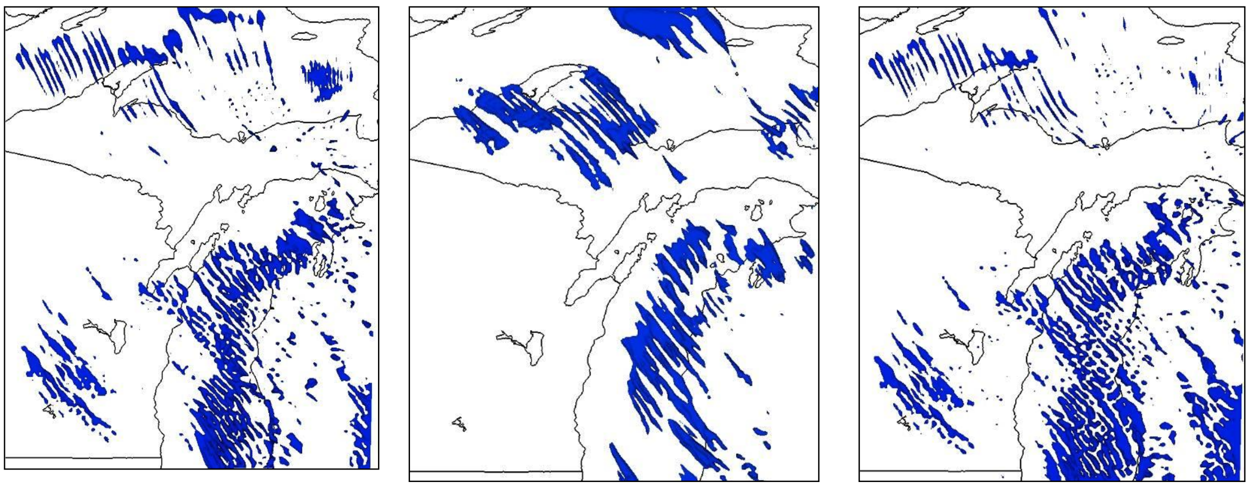
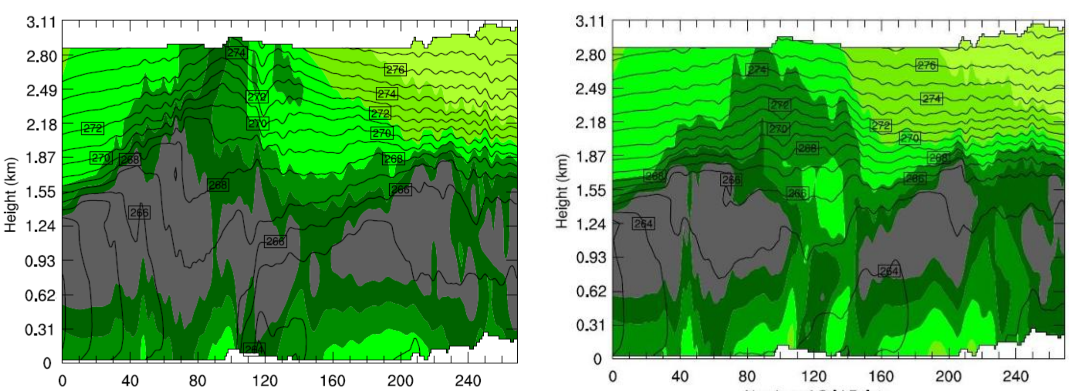
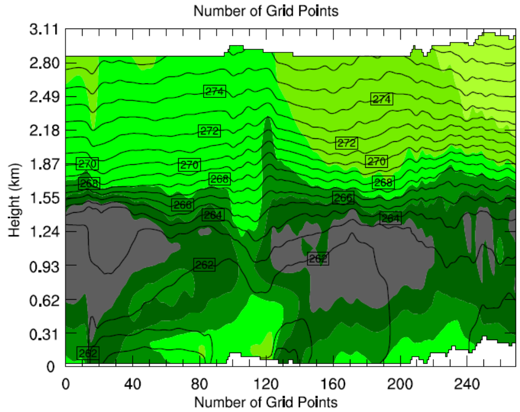


Figure 4.13. As in Fig. 4.9, but for the February Plus-5 simulation (left), base simulation (center), and Minus-5 simulation (right) at 00 UTC on 15 February 2007.





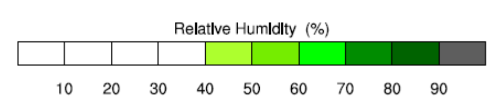


Figure 4.14. As in Fig. 4.10, but at 12 UTC on 15 February 2007.

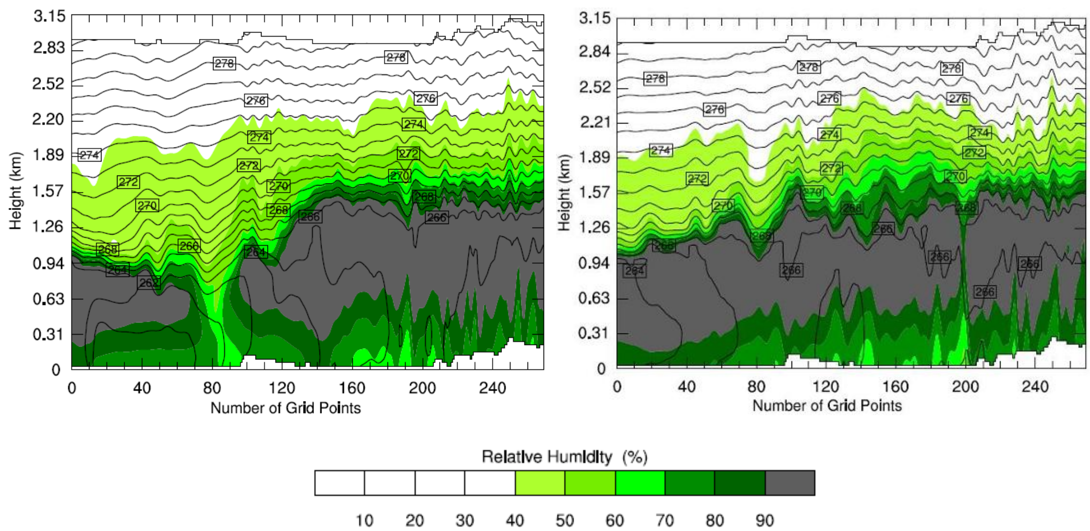


Figure 4.15. Boundary layer structure for Minus-5 and base simulations at 21 UTC on 15 February 2007 corresponding to the L2L part of the event. The Minus-5 case is to the left while the base case is to the right.

5. DISCUSSION

Two lake-effect events, 02 December 2003 and 15 February 2007, have been simulated to examine the environmental conditions conducive to L2L bands. The December case contained L2L bands throughout the event, while the February case started out as a typical lake-effect event and then evolved into an L2L case. Comparison of observations with plan-view images of model output for each case confirmed that the model accurately simulated the L2L bands and the environmental conditions which accompanied them. Only subtle differences existed between the observations and the simulations, and these differences were largely due to resolution differences. Sensitivity studies were performed to examine the effects of the strength of lake-land temperature differences and surface roughness over land (especially due to changes from snow cover).

Analysis of environmental conditions, presented in more detail in George (2008), indicated that for both cases, values of the most important variables for lake-effect snow: lake surface-850 hPa temperature difference, wind speed and direction, low-level inversion height, surface roughness, are consistent with both forecasting rules (Niziol, 1987, 1995) and theory (Hjelmfelt, 1990).

In the Noah land use scheme, surface roughness was decreased in proportion to snow cover even for urban and forest land use classes. To examine whether roughness and snow cover might play an important role in the ability of lake bands to persist while crossing land, this reduction of land use was eliminated. In the December case, snow cover was small, but in the February case, snow cover was more substantial over Upper Michigan. However in neither case was a significant change in the lake-effect bands observed. Thus, this study suggests that the roughness of the land surface probably does not play a critical role in L2L bands.

In the December 2003 L2L case, temperature differences over the western Great Lakes were about 18°C when the L2L bands began. Subtracting 5°K from the lake surfaces created a smaller temperature difference yet did not alter the time of L2L formation from the base simulation. Temperature differences as small as 8°C and as large as 25°C (as a result of adding 5°K to the lake surface temperature) were noted during the remainder of the L2L event. The February 2007 case consisted of higher temperature differences overall when comparing the L2L part of this case to the December case. L2L bands formed anywhere from 15°C (for the Minus-5 simulation) to 24°C (for the Plus-5 simulation), but were found to only last for a few hours when 5°K was added to the lake surfaces. This created a larger temperature difference of 24-27°C that tended to produce lake-effect snow bands only.

Inversion height, or the depth of the well-mixed boundary layer, was comparable through time between the two base simulations. The initial preconditioning of the lakes and land downstream of Lake Superior from the northwesterly flow led to boundary layer heights up to 2 km early in each case. As the high pressure systems approached the Great Lakes region, subsidence aloft lowered the inversion height until each of the two events ended with inversion heights less than 1 km. In each case, raising the lake surface temperature by 5°K caused an expansion in the depth of the boundary layer as more heat and moisture became available from the lake surface, which created more intense circulations. Within this boundary layer cloud bands were typically deeper than the base case counterparts. Lowering the lake surface temperature by 5°K had the opposite effect; that is, the decrease in the lake temperature caused shallower mixed boundary layers and cloud depths.

Wind direction was relatively consistent from the northwest for the two events. Surface wind speed over the western Great Lakes region during the start of L2L convection in both the December and February events was similar at 15 m s^{-1}. In the February case, this represented an increase in wind speed from somewhat lower values earlier with lake-effect snow, but no L2L bands. Even though surface level wind speed subsequently diminished gradually until the end of each event, L2L bands continued to form until wind direction shifted from the northwest to a more westerly direction over Lake Superior. The sensitivity studies did not affect wind speed or direction in either case.

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