Numerical Study of the 10 January 1998 Lake-Effect Bands Observed during Lake-ICE

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ABSTRACT

This paper presents the results of a series of idealized cloud resolving simulations of the evolution of moist roll convection observed as part of the Lake-Induced Convection Experiment (Lake-ICE) that took place during the 1997/98 winter over central Lake Michigan. Satellite and radar observations of the roll convection depict striking linear rolls stretching from 10 km off the western shore of the lake, across to the eastern shore, and then continuing across Michigan. The spacing of the primary rolls was observed to be 6 km, giving a ratio of spacing to depth of about 5:1, which is consistent with theory. In addition, a longer wavelength (13 km) of stationary banding was observed parallel to the shoreline.

In an earlier study of this case, multiply nested simulations of the convective rolls based on real data variable initialization were successful in producing banded structures with similar spacing and location over the water to those observed using fine grid resolution of about 500 m. Unfortunately, the initial locations of simulated bands were organized primarily by numerical effects of grid interpolation. This suggested that the spacing of the bands was robust, but that their initial location was highly sensitive to subtle systematic forcings. In this paper, a set of idealized model experiments, designed to isolate the role that physically realistic local forcing plays in the organization of the rolls, was performed. Because externally generated upstream turbulence was suppressed in these tests so as not to bias the result, the generation of rolls was delayed until 20–30 km downwind of the observed location and the location simulated in the previous grid nesting experiments. It was shown that the subtle effects of the shoreline geometry were sufficient to spawn a near-surface streamwise vorticity that became the primary seed for roll development at the most efficient mode of roll convection. These results suggest that previous structures evolved in the upstream shear-driven land-based mixed layer were likely also important in determining where the nonlocal overturning was first triggered. It is not clear from these results whether the shear-driven structures that evolved over the land also played a significant role in organizing the structural geometry of the lake rolls. Results also suggested that the shore parallel bands were a robust feature of the atmospheric structure resulting from resonant gravity wave trapping in the frontal layer.

1. Introduction

Banded convective structures in the planetary boundary layer are common, especially following cold air outbreaks over relatively warm bodies of water. Under certain conditions, persistent convective bands can result in significant precipitation such as occurs with lake effect storms in the Great Lakes region of the United States. Over the past three decades, a great deal of work has been published attempting to explain convective bands. Kristovich (1991) grouped the theories into two categories. “Classical” theories, such as Lilly (1966), Kuettner (1971), Asai (1970), Brown (1972), and Sykes and Henn (1989), relate bands to convective rolls due to the combined effects of wind shear and weak or unstable thermal stratification. In the case of weak thermal stratification, Brown (1972) described a cross-roll shear effect. For unstable thermal stratification, Kuettner (1971) described along-roll shear in the presence of Rayleigh–Benard instability. Roll convection has been theorized to benefit from the presence of vertical wind shear generated in a rotating frame of reference by Ekman turning for environments ranging from neutral to weakly stable thermal stratification (Brown 1980).

The second category of theories proposed by Kristovich (1991) consists of the “nonclassical” theories. These theories, proposed by Clark et al. (1986) and later advanced by Balaji and Clark (1988), Hauf and
Clark (1989), Balaji et al. (1993) and Lane and Clark (2002), relate banded convective structures in the boundary layer to a resonant forcing by gravity waves trapped in the stable layer topping the boundary layer. The theory explains certain relatively large banded structures of boundary layer convection, oriented with a component of their longitudinal axis perpendicular to the flow. Recently, observational evidence for the existence small, 6-km wavelength gravity waves over Lake Michigan during the Lake-Induced Convection Experiment (Lake-ICE) was also provided from synthetic radar observations of the water surface by Winstead et al. (2002).

Most observational studies of roll convection during cold air outbreaks and over relatively warm lakes or an ocean, suggest a major role is played by unstable stratification, that is, Rayleigh–Benard Instability (RBI), in the formation of roll convection. Predictions of roll convection over cellular convection have been attempted on the basis of determining the most unstable mode of RBI. Nevertheless, both classical and nonclassical approaches to explaining the bands may be relevant.

The aspect ratio of the convection, given by the ratio of the width of the band to the depth of the convection, together with the wind shear characteristics may provide some guidance to determining the underlying mechanism. Brown (1980) points out that theories of inflection point instability of the shear profile, initially proposed by Lilly (1966) for weakly statically stable environments, suggest aspect ratios of about 4:1 with roll orientation near the direction of the wind. Under conditions of static instability, RBI theory and numerical results of Mason and Sykes (1980) suggest smaller aspect ratios of 1–2:1, while calculations by Houze (1993) also suggest smaller aspect ratios of 3:1 as being the most unstable mode of RBI overturning. Observational evidence discussed by Miura (1986), however, suggests much larger aspect ratios are present in the real atmosphere, some as high as 10:1, behind cold air outbreaks. Sykes et al. (1988) made numerical calculations to demonstrate that large roll spacing of about 5:1 is produced in the presence of cloud top entrainment. They found that the suppression of convection by the entrainment of very stable air from above the planetary boundary layer (PBL) suppresses convection between streets, creating a wide spacing dependent on the degree of stability produced by the entrainment.

Although theories for the existence and spacing of observed roll structures in the generic CBL can be found, the formation and evolution of coherent structures in real observations is less well understood. Observations often indicate a mixture of coherent structural regimes in a particular region, suggesting that environmental factors likely play at least some role in determining what coherent structures ultimately will evolve. This is indirectly confirmed by LES simulations of coherent structures that sometimes exhibit multiple structural organization equilibria within a single environment (Andren 1995). Hence it is reasonable to expect that factors locally influencing structure may involve the history of events (hysteresis) within the flow that lead to particular structures observed.

The Lake-ICE experiment was conducted from 4 December 1997 to 19 January 1998 to study the structure and evolution of roll convection in the unstable boundary layer occurring during wind-parallel lake effect convection. Central Lake Michigan was chosen as the experiment site because it represented an ideal laboratory to study roll convection in the unstable boundary layer since 1) cold arctic air approaching the central portion of the lake tends to be relatively free of previous boundary modification from upstream lakes; 2) the central lake region is sufficiently far from congested air space, enabling the free operation of research aircraft; 3) the topography along the lake shores is modest and thought not to be a major factor in mesoscale organizations; 4) the shoreline is relatively simple, limiting its influence on mesoscale organization; and 5) cold air outbreaks resulting in wind-parallel roll convection occur predictably virtually every season. As part of this study, the University of Wisconsin (UW) set up its Volume Imaging Lidar (VIL) on the windward shore to observe the earliest stages of development of the unstable boundary layer within the frictionally driven boundary layer moving off the land.

In this paper, wind-parallel bands observed during the 10 January 1998 Intensive Observation Period (IOP) of Lake-ICE will be studied. In the next section selected satellite and lidar observations of the bands will be presented. In section 3, the design of an idealized numerical experiment designed to isolate important processes forming the bands will be discussed. In section 4 the results of the control experiment will be presented. In section 5, several sensitivity experiments will be presented, aimed at clarifying the results. Finally, in section 6, conclusions will be made.

2. Observations

Examination of the Geostationary Operational Environmental Satellite (GOES) superrapid scan visible imagery of the 10 January 1998 IOP 6a of Lake-ICE, seen in Fig. 1a, shows apparent ties of the dominant lines of clouds to shoreline features and strong continuity of structure and band size along particular bands.
A particular example is the convex shoreline feature near Two Rivers, Wisconsin (about one-third of the way down the lake from the top of the figure), that is associated with a dark line across the lake, where cloudiness is suppressed. The bands do not appear to markedly evolve in their spacing from the windward shore to the leeward shore and even subsequently across the land areas of central Michigan. In fact, one typical cloud band is observed to first appear at its full mature size in the middle of Green Bay, which is only about 10 km east of the windward shoreline. At an average 20 m s$^{-1}$ in the upper PBL, it would only take the flow less than 500 s to move from the shore to where the band is first visible as condensed cloud. This is well short of the time scale, expected for turbulent unstable overturning to develop such a large mature coherent structure. It may suggest the existence of pre-existing coherent structure in the flow impinging on the lakeshore from the west or a strong organizing influence of the shoreline geometry or topography in forcing the band.

Figure 1b depicts three major types of stationary banded structures apparent in Fig. 1a, and confirmed to be stationary relative to the ground and robust when the image of Fig. 1a is animated (see http://www.ssec.wisc.edu/data/rtlake/lake.html). The most striking organization are the type "A" bands, which begin in the visible cloud field about 10 km off the windward shore of the northern and central portions of the lake and then persist onto the lee shore all of the way across the state of Michigan. These bands exhibit a spacing of about 6 km both over the land and water. The bands are stationary but the cloud elements within move from the west to the east along the bands. Over the water, these bands tend to be less distinctive on the lee side of the lake, apparently as cloudy matter fills in the region between the bands. The bands then emerge to be more distinctive over Michigan, where the excessive cloudy matter between them disappears. These A bands seem to be nearly parallel to or at a small angle to the wind in the PBL observed by the Lake-ICE soundings.

Discriminating examination of the type A bands shows that some of the bands seem to be more well defined than others. Such "strong" type A bands are identified to be those having a greater degree of clearing between bands, which is evidence of greater amplitude of subsidence and cold/dry air entrainment between bands. Note that these strong type A bands tend to be downstream of the most severe windward shoreline bends, suggesting a possible influence of shoreline geometry.

The second most striking band type is the type "B" band. These stationary bands are oriented nearly perpendicular to the near-westerly inversion height wind and spaced approximately 10–15 km apart. They are strongest off of the windward shore and seem to be less defined on the east side of the lake.

Finally, the third band type is the type "C" bands oriented about 30° clockwise from the inversion level winds. These bands are the least common of the three types and occur primarily along the lee shore line of the widest sections of the lake. They are nonstationary and move toward the east-northeast normal to the bands. Their propagations to the east-northeast are consistent with a topographically induced trapped wave normal to the low-level east-southeasterly flow.

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1 NOAA took “superrapid scan” imagery on 10 January 1998 in support of Lake-ICE.
Examination of volume imaging lidar (VIL) observations near Sheboygan provides additional insight into the role of shoreline geometry on coherent cloud structures. Figure 2 depicts a calculation of wind velocity and divergence from the cross-correlation between successive near-surface PPI scans by the VIL on both the 10 and 13 January case (Mayor and Eloranta 2001). Since the 10 January surface winds are from the southwest and the 13 January winds are from the northwest, these two wind directions together elucidate differing shoreline effects on divergence within the scan area of the VIL. Figures 2a and 2b suggest the existence of plumes of convergence downwind of convex shorelines. By continuity, these near-surface plumes are suggestive of enhanced streamwise vorticity forming near-wind-parallel rolls, with downward motion predominating downwind of a convex shoreline and upward motion predominating downwind of a concave shoreline.

Just offshore, the UW–VIL (no observations were made onshore for eye safety reasons) observations explicitly depicted the existence of nonlocal vertical transport to the top of the boundary layer moving off shore from the land. Animations of RHI sections clearly depicted shallow thermals being swept up to the 1-km height of the boundary layer immediately as the flow passed off shore. However it was unclear if the shear-driven structures onshore were also nonlocal, or if the nonlocal thermal bursts only occurred as the warm lake air became involved.

Detailed examination of the weak clear air returns of the NEXRAD level II data from Green Bay suggested that the coherent structures over land to the west may have indeed been roll like, but of relatively small scales compared to the type A rolls observed in the clouds over the lake via satellite. The role of these preexisting structures is interesting and will be revisited in a future study.

The vertical atmospheric structure observed by Lake-ICE at Sheboygan is given by the sounding in Fig. 3a and the hodograph in Fig. 3b. This post-frontal sounding is taken by Lake-ICE at 1330 UTC just upwind of the windward shoreline. Trajectories passing over this point had not passed over an unfrozen lake in their recent history.

Note that there are four tropospheric layers of interest, beginning with 1.3-km deep PBL characterized by shear-driven mixing and an Ekman veering wind profile. The PBL is topped by a 1-km-deep frontal inversion layer (frontal zone) of very high stability and a backing wind profile characteristic of active cold-air advection. Above the frontal inversion layer lies the frontal layer extending to 3 km AGL, which is typified by weaker stability conditions and correspondingly weak wind shear. Between 3 km and the tropopause is the prefrontal troposphere characterized by weak stability.
3. Experiment design

To study the salient processes in creating the observed banded structures, a cloud-resolving numerical simulation of the event was performed. Preliminary cloud-resolving simulations (see Tripoli et al. 1999) were carried out using the UW–Nonhydrostatic Modeling System (NMS) configured in five two-way nested grids beginning with a continental scale grid having 60-km resolution and using five nests of ratios as high as 5:1 to grid down to 500-m spacing in the vicinity of Sheboygan and across the lake. These simulations successfully reproduced the observed type A and type B band structures to a remarkable degree. This included both band spacing and the regions of initiation. Investigation of the process by which bands were initiated revealed that the bands formed along boundaries of the parent nest’s grid box interfaces even when relative strong random turbulence was imposed. This suggested that the initial organization of rolls was sensitive to subtle but systematic forcings. In this study, we will perform a set of idealized numerical experiments designed to determine, in the absence of upstream forcing or numerical biases, 1) if the observed A and B bands are related to local instabilities and if so what instabilities and 2) if there is a physical basis for the apparent, but subtle, relationship between the observed bands and shoreline geometry.

In this section the UW–NMS used to perform the numerical integrations will be briefly described and the design of the idealized control experiment will be presented. The sensitivity experiments to the control will be described in subsequent sections.

a. Numerical model

The UW–NMS, described by Tripoli (1992) and Tripoli (2003, manuscript submitted to Mon. Wea. Rev.) is a quasi-compressible nonhydrostatic atmospheric mesoscale/cloud model. The current implementation included bulk microphysics prediction of the formation of pristine ice crystals, aggregated crystals, and rimed crystals. Rain and graupel were ignored for this cold cloud case. Cloud water, vapor, temperature, potential temperature, and density were diagnosed.

A local turbulence parameterization was employed based on a 1.5-level eddy viscosity formulation, where the eddy viscosity is diagnosed from a predicted turbulent kinetic energy. Surface fluxes were supplied from a surface-layer parameterization scheme described by Businger et al. (1971) coupled to a soil model described by Tremback and Kessler (1985) and modified to simulate effects of snow cover rather than soil. A two-stream long/shortwave cloud-active radiation scheme was used based on Ackerman and Stephens (1987).

b. Control experiment design

It was attempted to idealize the case by designing a single grid system with periodic boundary conditions in the alongshore (meridional) direction and open boundary conditions in the cross-shore (zonal) direction. This effectively idealized the lake as infinitely long. The narrow periodic domain in the meridional direction creates some restriction to possible band geometries that will be explored later. In the zonal direction, the upwind boundary was placed 30 km west of the lake and the inflow from the west was fixed. Development of explicit turbulence in the model between the lake and west boundary was suppressed by the imposition of a horizontal “sponge” absorbing layer, creating an idealized laminar flow incident on the west lake shore. Although observations by the UW–VIL did suggest that shear-driven turbulence was present in the flow incident on the lake’s western shore, the simulation of land-based coherent turbulence structures would have required a much larger upstream land area and probably very much higher resolution. To keep the scope of this study limited, this experiment was therefore designed to isolate local effects on the generation of bands and leave...
the issue of hysteresis from land-based coherent structures for a future investigation.

A cloud-resolving grid of 400-m spacing was chosen because it could resolve roll spacing considerably smaller than the observed 6 km, while maintaining computational affordability. One would not expect this simulation design to be able to capture the upscale organization of turbulence to the cloud scale as some have proposed (see Mayor et al. 2003), and this should be considered when interpreting the results. Although the 400-km horizontal and 100-km vertical resolution is sufficient to capture the A bands, it is marginal to capture much smaller large-eddy circulations likely having roll circulation scales of a couple of hundred meters as sensed by lidar (Mayor et al. 2003). To the extent that the growth of the A bands was an upscale process from the small large-eddy scale to the 6-km A-band scale, the present resolution would be insufficient to capture their formation. On the other hand, if there is instability on the 6-km A-band spacing that favors structures on that scale, this experimental design can capture their formation. For this reason, we term this experimental design a “cloud-resolving” model simulation rather than a large-eddy simulation (LES).

A minimally sized domain of 90 grid boxes in the meridional direction and 500 grid boxes in the zonal direction was chosen as a computational compromise to capture both the bands over the water and over Michigan and a small region over the land upwind of the lake. The periodic alongshore structure of the grid is identical to assuming an infinitely long Lake Michigan. This was deemed appropriate since the goal was to isolate local influences and the sounding structure itself on the organization of the bands.

To capture the effects of topography and shoreline geometry while preserving periodicity in the alongshore direction, the actual topography (from a 200-m resolution dataset) at the Sheboygan latitude was used for the

![Fig. 3b. Hodograph of observed winds at Sheboygan at 1330 UTC. The thin line is the smoothed observed wind profile and the thick line is the initial wind profile used to initiate the model. Within the boundary layer, there are two model initial profiles: one following the actual observations, and the second following a gradient wind profile determined empirically. The colors of the derived profile correspond to the height labels colored similarly.](image)
north half of the domain. For the south half, the topography of the north half was simply folded over to the south giving a reflective symmetry in both the shoreline and the topography. This somewhat matched the bay-like structure of the actual terrain on the east and west shoreline in the Sheboygan region, in effect idealizing a natural shoreline variation.

A 100-m vertical spacing was assigned to the lowest 1.2 km of the simulation domain and then stretched slowly above to 750-m spacing by 5-km AGL. The model top was placed high enough (12 km) so that the gravity wave trapping, seemingly causing the type B bands in the nesting simulations, could be represented.

The eastern outflow boundary employed the Durran (1981) gravity wave radiation condition. As mentioned above, the western inflow was held constant. A 50-point (20 km) horizontal sponge layer was specified adjacent to the boundary to absorb the westward growth of the type B waves and maintain a laminar flow. This meant that there were just 20 “free” grid intervals west of the western lakeshore. One consequence was that the predicted TKE incident on the western lakeshore, was about 50% lower than it would have been with out the sponge zone. This was because the static destabilizing and TKE generating effects of frictionally induced differential cold air advection were eliminated. Tests with a larger upwind free domain (not shown) demonstrated the effects to lead to the development of more vertical mixing on the windward shore as a result but no significant alteration to predicted roll convection over the lake.

The model was initialized horizontally homogeneous with the adjusted observed 1330 UTC Lake-ICE sounding taken at Sheboygan and the empirically derived gradient wind (Fig. 3). It was determined that 6 hours of simulation beginning at 0700 UTC was sufficient to reach an equilibrium structure across the 200-km domain. This simulation design was labeled the “control” experiment. Following a discussion of the results of the control, the design and results of sensitivity studies to the control experiment will be presented.

Time scales were approximately 3 h for flow at the top of the boundary layer to pass across the entire domain. The effects of geostrophic and gradient wind adjustment therefore could not be neglected in the simulation dynamics or in the initial conditions. As a result, the effects of the gradient wind adjustment had to be included in order to capture the accelerations of the wind in the boundary layer. Such a gradient wind was found using one-dimensional simulations to reproduce the observed Ekman spiral with the model eddy viscosity. The resultant wind profile is also given in Fig. 3b.

4. Results of control simulation

Figure 4 depicts the simulated cloud and accumulated precipitation field at 6 h. The precise amount of precipitation could not be evaluated since the pattern evolves from a fictitious initial state and observations on such a fine scale were not taken over Michigan. Nevertheless, the peak amounts of 7–8-mm liquid equivalent seem reasonable.

The simulated core cloud field depicted in Fig. 4 shows five–six separate cloud bands across the lake, having an average spacing of 6 km, which is consistent with the observations of type A bands. The spacing represents an approximately 5:1 ratio to the PBL height, which is again consistent with the findings of Sykes et al. (1988) for a CBL. The cloud bands extend
over the land region, although they weaken somewhat. For the surface threshold displayed (0.05 g m$^{-2}$), the southernmost bands are too weak to show up, indicating a meridional wavenumber 1 (36 km) oscillation is present in the roll amplitude. This is likely due to either the coastline or topography variation in the meridional direction. Note also a finer spacing in the bands is weakly present at their western extremity, but a wider spacing is ultimately selected. The simulated bands are oriented almost zonally except over the center of the lake where they are tilted slightly clockwise from the zonal direction.

The bands begin to appear about one-third of the way (30 km) across the lake in the simulation, which is considerably farther east than observations shown in Fig. 1a, where bands begin within 10 km of the western shore line. The preliminary multiply nested real-data simulations performed (Tripoli et al. 1999) produced bands considerably closer to the western shoreline than this test. The property of the control simulation that most delayed band formation was the imposition of the upstream absorbing layer that produced laminar flow impinging on the windward lakeshore. As mentioned earlier, VIL observations depicted the immediate eruption of nonlocal turbulence just off shore, which took nearly one-third of the lake width to develop in the simulation. In the case of Tripoli et al. (1999) the type A bands formed on what were most likely numerically generated subtle structures from grid interpolation entering the western boundary of the fine grid.

The orientation of the simulated type A bands is approximately east–west compared to the observed type A bands, which were west-southwest–east-northeast. There were several reasons found for this. One, demonstrated in the next section with sensitivity tests, was that restrictions of the narrow periodic domain scale created some resistance against acquiring a more positive tilt. Second, in the real atmosphere, the base state wind was rounding a trough whose axis was near the eastern shore of the lake. Hence, the observed winds veered considerably in moving from western Lake Michigan to eastern Michigan. This veering was not represented in the idealization.

If one compares the simulated type A band orientation with the local winds and wind shear, it becomes readily apparent that the simulated bands are oriented parallel to the inversion height winds. This appears to be consistent with the observations where the veering winds would explain the increasingly positive tilt moving eastward toward eastern Michigan. It is interesting that the bands are at a small angle to the surface winds and a large angle to the northwesterly shear across the boundary layer in the Sheboygan sounding.

The existence of type B bands in the simulation was profound. Depicted in the meridionally averaged vertical motion field cross section (Fig. 5), the simulated B bands are related to vertical motion fluctuations that extend throughout the deep troposphere but are strongly enhanced in the frontal layers between 1 and 3 km AGL, suggesting trapping. These waves were stationary relative to the surface and had a wavelength of 13 km and had wave fronts oriented north–south, all
consistent with the observed type B waves. Note that these waves are vertical with height and have strongest amplitude in the middle frontal layer and are stacked vertically. Corresponding thermal fluctuations are phase shifted 90°. Because the waves appeared to continue at low amplitude up to the model top at 12 km, it is suggested that, if they are ducted, the duct is “leaky.”

A wave ducting analysis procedure, outlined in the appendix, was performed on the horizontally averaged thermodynamic and horizontal wind profile across the lake for stationary gravity waves with wavelengths 8, 10, 20, and 40 km. Results shown in Fig. 6 confirm that a wavelength of 10–20 km is indeed susceptible to ducting in the frontal layer, confirming the hypothesis that the type B bands resulted from a ducted gravity wave. A separate sensitivity test was also performed (not shown) whereby the duct was eliminated by extending the temperature and wind velocity observed at the base of the inversion (850 hPa) to the model top, in effect making the Scorer parameter constant with height. In that test, the simulated B bands were eliminated, although the predicted A bands remained. It was interesting that, in that case, the predicted A bands were considerably more coherent and less fragmented, suggesting that the simulated B bands may be somewhat destructive to the coherent A structures.

Model-predicted horizontal wind direction and speed, vertical motion, potential temperature, and humidity are compared to actual aircraft observations (actually taken about 50 km south of the approximate model domain) as shown in Fig. 7. The comparisons are reasonable overall. Also, prior to the initiation of convection approximately one-third of the way across the lake, model vertical velocity is too weak. After this point, however, the magnitudes compare well. The simulated temperature and humidity tends to be somewhat higher than expected. This can be attributed to the vertical resolution of the model, which places the 200-m flight level only two grid intervals above the model surface. This is effectively in the “numerical friction layer” of the model, where the effects of surface forcing are still being “smeared” into a numerically resolvable feature. A bias toward the southerly direction with slower speeds is also consistent with this effect.

Figure 8 depicts the meridionally averaged mean state of the boundary layer across the domain. The top two parts of the figure display the square of the Brunt–Väisälä frequency defined as

$$N^2 = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial z},$$

where $\theta_v$ is the virtual potential temperature, $g$ is the gravity, and $z$ is the height coordinate. In regions where $N^2$ is large, stability is high and, where it is small, stability is low. Where $N^2$ is negative, the lapse rate is superadiabatic and the atmosphere is absolutely unstable. The deep blue shaded region depicts a relatively stable layer, found in the middle (inversion) portion of the frontal layer initially between 1 and 2 km AGL. Note that the layer drops (due to offshore wind acceleration) just offshore and then rises slightly to above its initial elevation on the eastern side of the domain. Note also the wave structures resident in that layer.

Fig. 6. Wave duct analysis for average conditions across simulated lake. Vertical axis is height and horizontal axis is wave number labeled by the corresponding vertical wavelength. Four curves are drawn for four horizontal wavelengths as labeled. The arrows indicate the minimum vertical depth range of the wave duct for a viable trapped wave to exist in a layer corresponding to the vertical wavelength labeled along the abscissa. The gray shaded region is the forbidden region where the vertical wave number is 0 or less. Five layers of the atmosphere are depicted by green (upstream shear driven boundary layer), orange (internal convective boundary layer), yellow (shear driven internal boundary layer), dark blue (stable frontal layer) and blue–green (upper tropospheric layer).
The second panel of Fig. 8 depicts the Richardson number given by

$$Ri = \frac{N^2}{(\partial u_i/\partial x_j + \partial u_j/\partial x_i)^2},$$

where $i, j, k$ are the three Cartesian directions; $x_i$ is the coordinate tensor; and $u_i$ is the velocity tensor. The value of $Ri$ is the ratio of the production of turbulence by shear to the stabilizing influence of stability. When $Ri$ is large and positive, the flow is laminar and, when $Ri$ becomes less than the critical value of about 0.25, the flow will break into turbulence. When $Ri$ is negative, the flow is turbulent and convective.

Figure 8 shows regions featuring all three situations of $Ri$ magnitude. The flow entering the domain on the western side shows a laminar layer in the frontal region overriding a turbulent shear-driven boundary layer (SBL “A”). After the flow moves off shore, heating from below begins to form a superadiabatic layer, driving $Ri$ to be negative and forming a convective boundary layer (CBL “A”) within the SBL “A.” This is a classical example of an internal boundary layer (IBL). CBL “A” is simulated to deepen to nearly 1 km before explicit convection develops about one-third of the way across the lake. This deepening must take place through local subgrid-scale downgradient mixing, which requires the thermal profile to be superadiabatic. In the real atmosphere, turbulent transport can be nonlocal, so such deep superadiabatic layers are not observed.

Once explicit convection erupts, CBL “A” is replaced by the much more shallow CBL “B,” where the superadiabatic layer is considerably more shallow and the zonal wind shear is concentrated in the lowest few meters. This is because the explicit turbulence is nonlocal, and consumes the superadiabatic layer. This is clearly more realistic. This concentration of shear tends to make $Ri$ more negative in CBL “B” than CBL “A.” Above CBL “B,” the $Ri$ within becomes supercritical.
the Brunt–Väisälä frequency is weak, and the shear is weak. There, nonlocal turbulence in the form of buoyant convective plumes enters from below and drives upward to the top of the PBL where it is turned back from the middle stable frontal layer. The top of this CBL “C” rises in response to the convective plumes from about 1 to 1.3 km by the lee shore, as was observed. At the top of CBL “C” strong shear between convective plumes, carrying low momentum from below and strong geostrophic winds above, produces a weakness in Ri clearly seen as a lighter blue region labeled as the “top of CBL” in Fig. 8.

To the east of CBL “B” surface heating shuts down and the superadiabatic layer disappears. As a result, Ri returns to positive values. The loss of penetrating convective plumes drastically reduces the scale length of turbulence near the surface, now generated entirely by shear. This produces an accumulation of frictional slowing in the lowest 100–200 m near the surface, greatly enhancing the shear in that region and reducing it aloft. A new IBL based on shear, labeled SBL “B,” grows upward in response to a deepening shear layer. Above SBL “B” the near-neutral PBL effectively decouples from the surface. The layer is turbulent, composed of turbulence in the form of passive type A rolls advected from over the lake. Because the layer is decoupled from the surface friction, the type A rolls can be maintained for a long distance or until the SBL “B” grows deeper to engulf the rolls.

The hodographs and vertical profiles of momentum, $\theta_e$, and $q_v$ given are also depicted in the lower portion of Fig. 8. The hodographs show the maintenance of an
“Ekman” layer through the formation of CBL “A,” which is destroyed in CBL “B.” Hence it is apparent that the simulation suggests that Ekman turning occurs only in the presence of small scale lengths and down-gradient turbulence. Hence, the existence of an Ekman layer on the west side of the lake in the control simulation is likely a result of the failure of the simulation to produce nonlocal turbulence on the resolvable scale. Note that the deep shear layer in CBL “A” is replaced by a very shallow shear layer in CBL “B.” Also evident is the fact that, in all regions, the lapse rate is very close to neutral and the superadiabatic departure from neutrality is only on the order of 1°C. The CBL wind becomes almost pure westerly once explicit convection commences. This is because the v component of the flow upstream was produced purely by a slow local vertical mixing process in the presence of Coriolis, which occurs on much too long a time scale to be maintained once explicit rapid nonlocal turbulence begins and transports momentum throughout the layer quickly. This has implications for nonlocal transport models, which usually do not take into account momentum transport and continue to maintain an Ekman profile.

At the eastern end of CBL “A” the turbulence that forms is much too strong as a result of the “nonphysical” deep superadiabatic layer. Clearly the merging of an explicit CBL with a parameterized CBL cannot occur smoothly with local turbulence closure in regions where buoyant plumes are unresolved.

Figure 9 shows the averaged vertical heat, moisture, and momentum transport over five regions across the domain. Here one can see that heat is transported in the net upward below the inversion but either upward or downward above the inversion. In regions on the eastern half of the lake and just west of the eastern shore line, where heat is actively entering the system from the surface or surface layer, the transport above the base of the inversion is negative, while in regions west and east of the lakeshore, the transport above the base of the inversion is positive. These regions of positive transport suggest that the disturbance is transporting cold air from above downward. Over the lake, the disturbances are dominated by overshooting warm thermals, which become negatively buoyant above the inversion but continue to be driven upward by their momentum.

Interestingly, the moisture fluxes effectively disappear above the inversion. This is because there is no large vertical gradient of moisture above, so the propagation of a gravity wave into the frontal layer will not produce significant local downward transports of moisture in that region. It is also noteworthy that the surface moisture flux reaches its maximum value on the western side of the lake, while the moisture flux at 200 m reaches its maximum near the eastern side of the lake. This can be attributed to the moisture being deposited into the low layers early, when the difference between the air and water humidity is the greatest, but being transported upward strongest later when the nonlocal convective thermals become stronger.

The vertical momentum flux shows some similar characteristics to moisture in that the surface flux of zonal momentum is strong initially, but the 200-m elevation flux is strongest as the thermals become stronger farther downwind. This effectively causes the wind to slow in the lower boundary layer, and then speed back up as the deeper transport begins replenishing momentum toward the eastern side of the lake. The
momentum flux off the water is significantly higher once the thermals come alive.

It is also noted, that, since the surface wind was near zonal, the $u$-component friction was also small. Nevertheless a strong upward flux of $v$ momentum (strong negative downward flux) occurred over the middle of the lake. This is a result of the penetration of cold northerly flow dominated thermals from the top of the boundary layer downward. Since the mean flow never becomes strong northerly sheared, it seems this transport was occurring nonlocally in negatively buoyant thermal bursts from above.

Based on the definition of a Rayleigh number by Houze (1993) and using the modeled eddy viscosities for heat and momentum in place of molecular viscosity, the Rayleigh numbers and critical Rayleigh numbers of the modeled flow for cellular and roll convection are presented in Fig. 10. By comparing the Rayleigh number of the modeled flow to the critical Rayleigh number for either linear or cellular convective rolls, we can determine for which organization, if any, either dry or moist convection is unstable. We find that the flow is unstable for both roll and cellular convection on the windward side of the lake prior to the onset of explicit convective rolls. Lee of the explicit roll development, the simulation is unstable only with respect to moist convection. Note that the most unstable modes are for cellular convection, despite the proliferation of roll convection in the model simulation. Hence the model results seem to confirm the notion that it is not the most unstable modes that are selected for growth.

The successes of the control simulation can be considered quite remarkable given the complexity of the observed structure when compared to the simplicity of the idealized simulation. Of particular note is the fact that the simulation featured explicitly laminar flow approaching the western lakeshore and yet developed reasonable looking bands, albeit delayed, over the lake. Grid resolution restrictions alone would prevent the growth of mature explicit turbulence over the western side of the lake. This would suggest that the simulated bands grew from local effects rather than as an upscale progression of maturing stochastic turbulence. In the next section several sensitivity tests will be presented, designed to uncover what the salient processes were.

5. Sensitivity experiments

a. Grid design

The restrictions of the idealized grid design, including resolution and domain size, may have played a role in determining the coherent structures that were simulated. To better understand the roles played by these
considerations, two sensitivity experiments were conducted.

• A1: The simulation was rerun with the same grid dimensions and covering the same region but with 200-m grid spacing instead of 400-m spacing. This will help determine to what extent the simulated band geometries and sizes were a result of grid resolution.

• A2: The simulation was rerun with the double the domain size in the meridional direction, by repeating the land and topography function one more cycle in the north. This experiment was designed to determine to what extent restrictions imposed by the narrow periodic domain in the meridional direction influenced band geometries.

The results of these tests are given in Fig. 11. In both A1 and A2 the bands begin slightly farther west than the control. This is likely because explicit overturning begins sooner with higher resolution for A1. The reason that occurs in A2 is likely because there was some suppression of the eruption of turbulence by the periodic nature of the domain.

The results of A1 show a tendency to at first form more numerous structures on the western extremity of the cloud mass but to evolve five–seven structures similar to the control by the eastern shore. Over the land to the east, it seems that there is a greater tendency for the band strength to weaken. There seems to be five–six bands remaining over the land as in the control, suggesting that higher resolution did not increase the number of bands that eventually survived.

The results of A2 are quite interesting in how markedly the west-southwest–east-northeast slope of the bands increased by making the domain larger in the meridional direction. This simulated northeast to southwest band tilt is somewhat closer to that observed within the type A bands shown in Fig. 1. In addition, there are six–seven bands in A2 versus five–six bands in the control.

This test would seem to suggest that the control domain size in the meridional direction was placing some restriction on the band orientation, winds, and the exact spacing of the bands. Nevertheless, it appears that the essence of the coherent structures simulated in control is relatively robust under some major alterations of the domain setup.

b. Sensitivity to surface features

It is also of interest to determine how much the model physics influence the simulated band structures. These physics tests include

• B1: Eliminate Coriolis and $\nabla$ component of wind to isolate the effects of the $\nabla$ component of the wind on the coherent structures. This will eliminate the effects of geostrophically induced streamwise vorticity.

• B2: B1 plus no zonal shoreline variations, no topography to see what effects the combined variations of topography and shoreline have on the formation coherent structures.

• B3: B1 plus no topography to isolate the effect of the variation of topography on coherent structure.
- **B4: B1 plus no shoreline variations** to isolate the effect of the variation of the shoreline on coherent structure.

- **B5: B2 plus a smooth sinusoidal function for shoreline variation** to isolate the effect of very small scale shoreline structures. This tests to see whether the wavenumber 1 (36 km) scale shoreline variation is sufficient to form 6-km bands or whether the 6-km scale shoreline variations are key to forcing coherent structures.

The results of experiment B1 (Fig. 12) suggest that the existence of the bands was not dependent on the meridional wind or on the curvature of the hodograph.
The strength of the bands does decrease more over the land in the case of no meridional wind, however, suggesting that the meridional wind may play a role in the band maintenance downwind of the convective boundary layer.

Experiment B2 reveals that, when no shoreline or topography variations are specified, the lake convection is cellular and effectively absent of bands entirely. Moreover, the existing convections disappeared at the 0.5 g m$^{-3}$ condensate over the eastern simulated land area. Apparently, the meridional surface variations have a critical effect on band formation. Experiments B3 and B4 suggest that the banding is almost entirely due to the shoreline variation rather than the meridional topography variation.

It was found that there was a significant link between the individual type A bands and shoreline features through the $x$ component of (streamwise) vorticity near the surface, given by the color shading in Fig. 12. This represents the intensity of an $x$-oriented roll motion and it is apparently initiated at very low values along the shoreline. The amplitude of the vorticity variation was about $\pm 60$ s$^{-1}$ and was only about 100–200 m deep. It was initiated baroclinically in regions of shoreline variation as a result of meridional heating gradients as flow within a concave region of shoreline moved offshore ahead of flow where land projected out farther.

Two scales of vorticity production are evident: the scale of small shoreline variations and that of the wavenumber 1 variation. The weakening of bands in the northern and southern part of the domain over land in the control was apparently linked to this wavenumber 1 variation. The question arises: Was it really the small-scale shoreline variations, or was it the overall baroclinic production of streamwise vorticity by the wavenumber 1 shoreline variation that resulted in the type A bands? Experiment B5 was designed to answer this question by composing an “idealized” shoreline from a sine function that closely approximates the control shoreline, but without the small-scale fluctuations. The result, also shown in Fig. 12, was a more dominant wavenumber 1 effect but not the elimination of type A rolls. Interestingly, the rolls that did form on the upwind side of the cloud mass at a much smaller spacing of 6–8 $\lambda$. It is noteworthy that the rolls on the eastern lakeshore again settled out at the observed type A spacing, again suggesting a preference for a 6 km (5:1) spacing consistent with Sykes and Henn (1989).

Together, these results suggest that for roll convection to form, it is important that streamwise vortices be initiated independently and sufficiently, in this case by shoreline variations. If the 5:1 (6 km) rolls appear initially, they will rapidly grow relative to other scales and dominate. Otherwise, rolls can be initiated on larger or smaller scales (experiment A1) but will eventually evolve to the preferred 6-km scale. What constitutes “sufficiently” is not well answered here. Certainly one would expect a significant influence of any preexisting upstream roll convection in initiating rolls over the lake if it existed. The relative importance of upstream coherent structures to shoreline variations on the formation of rolls is unknown. It would likely depend on the horizontal and vertical scales of any preexisting structures. Observations provide evidence that both effects may be important.

6. Summary and conclusions

In this paper, the 10 January 1998 Lake Michigan convective boundary layer and its evolution downwind of the lake were studied using idealized numerical simulation of the event where environmental factors influencing the event could be varied. Results were compared to observations taken during the Lake-ICE field campaign.

Satellite and radar observations indicated that there were three categories of ground-stationary banded structures labeled in this study to be 1) type “A” bands, oriented west-southwest to east-southeast and turned at a small angle counterclockwise from the shear, spaced 6 km, and spanning from near the western lake shore across the entire state of Michigan; 2) type “B” bands, oriented approximately parallel to the lake shore, nearly perpendicular to the wind, spaced 13 km, and strongest on the windward side of the lake; and 3) type “C” bands which were oriented northwest to southeast, spaced 6 km, and confined to limited regions on the lee side of the lake, only over the wider portions of the lake.

The cloud-resolving simulations were successful in simulating both the types A and B bands and the persistence of the type A bands downwind of the lake within the limitations of the idealized domain. The type simulated A bands, however, initiated about 20–30 km too far offshore. Simulations, not shown, with the duct removed were conclusive in demonstrating that the type B bands resulted from resonant trapped internal gravity waves in the lower frontal layer fueled by penetrating moist convection over the lake. The 13-km spacing represented the dominant trapping scale for the frontal structure present and for flow normal to the convection.

These results suggest that the type A bands formed in the presence of a “seed” that initiated a weak roll type circulation. Sensitivity experiments suggested that, if the undulating shoreline geometry included sufficient amplitude variations near the 6-km scale, 6-km bands

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would be seeded by the baroclinic generation of streamwise vorticity along the shoreline.

An implication of this is that one would expect the roll formation to also be sensitive to coherent structures within the upstream shear-driven boundary layer. This implies that, in PBLs where Rayleigh numbers are very large and all modes are unstable, the growth of a particular coherent structural regime is likely more dependent on the existence of initiating structures in the flow than on what the most unstable growth mode is. Our experience with the nonidealized variable initialization of this case in a previous study suggests that the delayed initiation of convective overturning in this idealized experiment was likely related to the lack of initial upstream structures. It is not obvious, however, if the upstream shear driven structures would be of the scale or geometry necessary to dominate the organization over the subtle but systematic effects of shoreline geometry. We plan to investigate this in a future study where we will attempt to simultaneously simulate a mature upstream boundary layer at very high resolution.

Finally, the striking persistence of the bands across Michigan resulted from the development of an internal shallow shear-driven internal boundary layer on the lee shore eastward. This effectively confined vertical turbulent mixing to a shallow layer near the surface and decoupled the cloud layer from surface friction generated turbulence. The upper PBL became laminar and neutral and the cloud bands could persist in the absence of destructive dissipation.

Future work should be focused on better defining the role of hysteresis in determining roll placement and structure. The Lake Michigan laboratory has proved to be a valuable tool. The previously unexpected importance of upstream structure requires that, in the future, additional field campaigns featuring lidar observations pointed both off- and onshore are needed to better quantify these apparent effects.

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APPENDIX

Wave Trapping Analysis

In this appendix a procedure for the quantitative analysis of the wave trapping in the atmosphere based on the Lindzen and Tung (1978) discussion of wave ducting is presented. Beginning with the set of two-dimensional inviscid, linearized equations of motion in Cartesian coordinates, applying a wave solution in the form, \( A(x, z) = \hat{A}e^{(i(kx + cz)} \), and combining equations yields an approximate vertical structure equation given by

\[
 w_{zz} + (l^2 - k^2)w_{zz} = 0, 
\]

where \( x \) is the horizontal distance, \( z \) is the vertical distance, \( t \) is time, \( w = dz/dt \) is the vertical motion, \( k = 2\pi/L_z \) is the horizontal wavenumber, \( L_z \) is the horizontal wavelength, and \( c \) is the gravity wave phase speed. The Scorer parameter \( (A1) \) is given by

\[
 l^2 = \frac{N^2}{U^2} - \frac{\partial^2}{\partial z^2} + \epsilon, 
\]

where \( U \) is the mean horizontal velocity, \( \epsilon \) represents the remaining terms that are neglected, and \( N \) is the Brunt–Vaisälä frequency given by

\[
 N^2 = \frac{g}{\theta \frac{\partial \theta}{\partial z}}, 
\]

where \( \theta \) is the potential temperature and \( g \) is gravity. For this formulation, it is assumed that the wave is stationary relative to the surface so propagates with a phase speed equal and opposite to \( U \). The vertical wavenumber \( (k_z) \) for a gravity wave is given by

\[
 k_z = \frac{2\pi}{L_z} = (l^2 - k^2)^{1/2}. 
\]

Gravity wave solutions exist only for the case \( l^2 > k^2 \). As \( l^2 \rightarrow k^2, L_z \rightarrow \infty \), which demonstrates the need for an increasingly long vertical wavelength to achieve a viable gravity wave as \( l^2 \) approaches \( k^2 \). The point where \( l^2 = k^2 \) is termed the “short-wave cutoff,” and it is the smallest horizontal wavelength for which there is a gravity wave solution. As \( l^2 \) varies with height gravity solutions may or may not exist for a given horizontal wavenumber \( k \). If a gravity wave solution exists in one layer and not in layers above or below, then the gravity wave will become trapped within a layer and that layer can be termed a wave duct. As Lindzen and Tung (1978) point out, the duct must be at least 1/4 of a vertical wavelength deep in order for a gravity wave to be maintained in the duct.

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