The OWLeS IOP2b Lake-Effect Snowstorm: Dynamics of the Secondary Circulation

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ABSTRACT

Intense lake-effect snowfall results from a long-lake-axis-parallel (LLAP) precipitation band that often forms when the flow is parallel to the long axis of an elongated body of water, such as Lake Ontario. The intensity and persistence of the localized precipitation along the downwind shore and farther inland suggests the presence of a secondary circulation that helps organize such a band, and maintain it for some time as the circulation is advected inland. Unique airborne vertical-plane dual-Doppler radar data are used here to document this secondary circulation in a deep, well-organized LLAP band observed during intensive observing period (IOP) 2b of the Ontario Winter Lake-effect Systems (OWLeS) field campaign. The circulation, centered on a convective updraft, intensified toward the downwind shore and only gradually weakened inland. The question arises as to what sustains such a circulation in the vertical plane across the LLAP band. WRF Model simulations indicate that the primary LLAP band and other convergence zones observed over Lake Ontario during this IOP were initiated by relatively shallow airmass boundaries, resulting from a thermal contrast (i.e., land-breeze front) and differential surface roughness across the southern shoreline. Airborne radar data near the downwind shore of the lake indicate that the secondary circulation was much deeper than these shallow boundaries and was sustained primarily by rather symmetric solenoidal forcing, enhanced by latent heat release within the updraft region.

1. Introduction

Cold-air outbreaks over the North American Great Lakes in late fall and early winter often lead to lake-effect (LE) snowfall, a phenomenon that occurs when relatively cool air moves across and is modified by a much warmer large body of water. Warming and moistening of the near-surface air produces a well-mixed boundary layer driven by moist convection and deepening with fetch from the upwind shore. Such convection tends to organize linearly into bands, parallel to the low-level wind or low-level wind shear (Niziol et al. 1995; Young et al. 2002). Bands that form when the low-level fetch over the lake is small, and the lake is wide in the crosswind direction, are typically wide-spread and shallow, often becoming regularly spaced in response to a helical roll circulation (Kelly 1982, 1984; Kristovich 1993) that modulates the spatial distribution of moist convection (Yang and Geerts 2006). Over relatively narrow, elongated lakes, the larger fetch resulting from prevailing winds that are parallel to the long axis of the lake often favors the development of a single narrow band aligned with the wind (McVehil and Peace 1965; Peace and Sykes 1966; Byrd et al. 1991; Niziol et al. 1995; Steiger et al. 2013). Of all LE morphologies, these tend to be the strongest and most prolific snow producers. Such bands occasionally occur over Lake Michigan under northerly flow, but are more commonly observed over the two eastern Great Lakes, Lake Erie (oriented southwest–northeast) and Lake Ontario (oriented west–east), on account of their dimensions and the common strong westerly wind in the wake of a cold frontal passage (Kristovich and Steve 1995).

This long-fetch convection may evolve into a larger (meso-β scale) LE system that includes not only highly organized and intense singular bands as mentioned above, but also weakly organized convective cells.
(e.g., Kristovich et al. 2003). The latter partially fits the description of “broad coverage” used by Veals and Steenburgh (2015) and Campbell et al. (2016). The former generally develop within areas of low-level convergence over the lake (Peace and Sykes 1966; Holroyd 1971; Niziol et al. 1995). Steenburgh and Campbell (2017) discuss two mechanisms controlling such convergence: differential surface roughness (i.e., surface friction) across the shoreline (e.g., Holroyd 1971; Lavoie 1972; Alcott and Steenburgh 2013), and a thermal contrast, which produces a land-breeze front (Peace and Sykes 1966; Passarelli and Braham 1981; Hjelmfelt and Braham 1983; Hjelmfelt 1990; Steenburgh and Onton 2001; Onton and Steenburgh 2001; Laird et al. 2003). The latter is driven by differential surface heat fluxes over water versus land. The thermal contrast may become quite strong when there is a significant prevailing wind aligned along the shore, since surface heat fluxes over water are proportional to wind speed, resulting in larger fetch that increases the amount of heat (and moisture) that is accumulated in the boundary layer. Depending on the strength and orientation of the prevailing wind, land-breeze fronts of this type from opposing shorelines (e.g., the north and south shorelines of Lake Ontario) may converge near the middle of the lake, leading to the development of a single midlake band (Passarelli and Braham 1981; Braham 1983). Alternatively, a land-breeze front may remain close to the shoreline where the prevailing wind has an onshore component, resulting in the formation of a shallower “shoreline” band (Braham 1983; Hjelmfelt and Braham 1983; Kelly 1986). Both bands tend to occur under relatively weak onshore prevailing winds, since stronger onshore winds may inhibit land-breeze development (Biggs and Graves 1962; Walsh 1974; Hjelmfelt 1990).

Recently, Steiger et al. (2013) coined the term long-lake-axis-parallel (LLAP) band, a term that has since seen more widespread use (e.g., Minder et al. 2015; Veals and Steenburgh 2015; Campbell et al. 2016; Kristovich et al. 2017; Welsh et al. 2016), to better describe what Niziol et al. (1995) refer to as a “type-I” band. These are deep, intense bands that produce the heaviest LE snowfall and develop within broad mesoscale convergence zones under relatively strong flow. Such LE snowbands often wreak havoc on local commerce and travel (Niziol et al. 1995; Leffler et al. 1997). In long-fetch situations featuring more cellular (i.e., broad coverage) rather than banded convection, or some combination of the two, the term “LLAP system” is perhaps more appropriate. Veals and Steenburgh (2015) show that LLAP bands account for only about 14% of all LE snow hours over eastern Lake Ontario, according to 13 years of data from the Montague, New York (KTYX), Weather Surveillance Radar–1988 Doppler (WSR-88D). They essentially use the term “LLAP” in place of midlake bands and define them as “wind-parallel bands generated by land-breeze convergence when the prevailing flow is oriented along the long axis of an elongated body of water” (p. 3592). Steiger et al. (2013) state that converging land breezes may be important for LLAP band development, but they also note that these land breezes are “usually superimposed on a significant synoptic-scale westerly low-level flow” (p. 2822) (i.e., over Lake Ontario). For the purposes of this study, we do not differentiate between LLAP and midlake bands and will use the term “LLAP” throughout the rest of the paper.

In addition to the land breeze, the typical depth and intensity of well-organized LLAP bands imply that another mechanism, latent heating, is at work that helps to strengthen and deepen the shallow thermally driven land-breeze circulation. The presence of thermally driven vertical circulations within LE snowbands has been noted for some time (McVehil and Peace 1966; Lavoie 1972; Hjelmfelt and Braham 1983). By increasing the buoyancy within the band, the release of latent heat leads to a strengthening of the updraft, enhancement of the lake-scale low-level convergence, deepening of the band, and ultimately heavier snowfall over and downwind of the lake. Significant latent heating can be expected in the case examined here, owing to high surface latent heat fluxes driven by strong surface winds (10–15 m s$$^{-1}$$) and rather high lake surface temperatures ($$6^\circ$$–$$7^\circ$$C). The resulting secondary circulation is similar to the convection-induced circulation implied by Steiger et al. (2013), in which low-level moisture convergence is sustained by solenoidal forcing. Numerical simulations have shown a lake-scale solenoidal circulation in LLAP bands (Ballentine et al. 1998). The present study is the first, to our knowledge, to document this circulation with detailed observations in the vertical plane across a LLAP band.

In this paper, we examine the secondary solenoidal circulation within a LLAP band observed during intensive observing period 2b (IOP2b) of the Ontario Winter Lake-effect Systems (OWLLeS) campaign. Steenburgh and Campbell (2017) show high-resolution model data indicating that the broader LLAP system during IOP2b formed along a shallow convergence zone

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1 According to the bulk aerodynamic formula (e.g., Stull 1988), the surface latent heat flux is proportional to the mean wind speed and the difference between the saturation vapor pressure at the temperature of the lake surface and the actual vapor pressure of the air.
associated with a land-breeze front that developed along a bulge in the southern shoreline and extended downstream over eastern Lake Ontario. This study focuses on <1 h of airborne radar observations from when the LLAP system had organized into a well-defined, narrow, steady, and deep LLAP band that extended over the eastern end of the lake and inland over the Tug Hill Plateau (hereinafter “Tug Hill”). We assume that the LLAP band was in steady state during these observations, which were obtained along five cross-band legs flown in a Lagrangian sequence (following the mean flow from west to east) by the University of Wyoming King Air (UWKA) research aircraft. The intent of this study is to document the secondary circulation and its spatial evolution (or transition) from west to east, from open water to land. Thus, we examine structural differences within the LLAP band as the circulation, clouds, and precipitation are advected inland and respond to the changing lower boundary conditions.

We hypothesize that the development of a strong, solenoidally driven secondary circulation, intensified by latent heat release, is essential for such a well-organized LLAP band. To support this hypothesis, high-resolution airborne dual-Doppler (DD) radar observations are presented that exhibit the 2D kinematic structure in a vertical plane across this band. Furthermore, these observations are combined with high-resolution model output in an effort to investigate the dynamic mechanisms driving the circulation.

The following section describes the datasets, analysis methods, and model configuration employed for this study. Section 3 provides background and a brief overview of the IOP2b event while section 4 presents a comparison of the airborne radar observations and model output with regard to LLAP band vertical structure. In section 5, an in-depth analysis of the dynamic mechanisms driving the secondary circulation is given. Finally, the results are discussed in a broader context in section 6 and summarized in section 7.

2. Data and methods

a. Airborne in situ and remote sensing measurements

The UWKA, outfitted with in situ and remote sensing instrumentation, is designed for investigation of atmospheric phenomena in the lower troposphere. In situ data collected at flight level by the UWKA during IOP2b include standard measurements of temperature, pressure, humidity, and the 3D wind, as well as measurements of cloud particle size distributions from optical array probes such as a cloud imaging probe (CIP) and a cloud droplet probe (CDP). Welsh et al. (2016) and Bergmaier and Geerts (2016) discuss the in situ instrumentation on board the UWKA during OWLeS in more detail. Additional information regarding the full suite of UWKA instrumentation and measurement capabilities can be found in Rodi (2011) and Wang et al. (2012).

This study makes ample use of finescale airborne radar data from the Wyoming Cloud Radar (WCR), mounted on board the UWKA during OWLeS. The WCR has a wavelength of about 3 mm (95 GHz) and includes multiple fixed antennas oriented in various directions. For OWLeS, there were three antennas: one pointing in the zenith (up beam), nadir (down beam), and 30° forward of nadir (down-forward beam). The beam widths of these antennas ranged from 0.5° to 0.7° and the range resolution of each was about 30 m. Data from the WCR were sampled at about 20 Hz, which corresponds to an along-track resolution of 4–5 m at typical UWKA airspeeds of 80–100 m s−1.

Equivalent radar reflectivity (hereafter “reflectivity”) and radial velocity were obtained from each beam. A relatively high reflectivity noise threshold has been implemented for this study, since the majority of radar echoes within the LLAP band were quite strong, more than three standard deviations above the mean noise. This yields a minimum detectable reflectivity for the most (least) sensitive beam of about −33 (−28) dBZ at a range of 1 km. The maximum unambiguous radial velocity is ±15.8 m s−1.

In the absence of aircraft pitch and roll, the radial velocity obtained from the up and down beams represents the reflectivity-weighted vertical velocity of hydrometeors. However, banking turns and slight aircraft wobble during flight causes these radial velocities to become contaminated by both the aircraft motion and the horizontal winds. The effects of aircraft motion are automatically removed during postprocessing, but horizontal wind contamination in the event of aircraft roll must be corrected for separately by assuming a horizontal wind profile. For this study, such a profile has been obtained from a sounding taken during the flight at 2018 UTC from Oswego, New York,2 ~5 km south of the LLAP band (Fig. 1). The use of other soundings taken at the same time did not produce any significant differences in the velocity field, since typical roll angles were relatively small.

Following these corrections, an estimated hydrometeor fall speed can be removed to obtain an

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2 Another sounding was taken at the same time within the band from Sandy Creek, NY. However, wind data were missing from that sounding in the lowest ~1000 m AGL.
estimate of the air vertical velocity. A comparison of the WCR hydrometeor vertical velocities from the nearest range gates above and below flight level with the gust probe air vertical velocity shows that typical fall speeds of hydrometeors, at various flight levels, ranged from about 0.5 to 1 m s$^{-1}$ in the LLAP band (not shown). Furthermore, according to the CDP and the CIP, the LLAP band was composed mostly of unrimed or lightly rimed aggregates with mean concentrations peaking at diameters of about 0.5–1.0 mm (Welsh et al. 2016). On occasion, heavily rimed aggregates and graupel particles were observed, with diameters much larger than 1 mm, within or near strong convective updrafts (Welsh et al. 2016). While unrimed aggregates with diameters of ~1 mm or less tend to have terminal velocities close to 0.5 m s$^{-1}$ (e.g., Locatelli and Hobbs 1974), much larger graupel particles (diameters >2 mm) may fall at velocities in excess of 2 m s$^{-1}$ [see Fig. 7 in Mitchell (1996)]. Fall speed cannot be assumed to be a function of WCR reflectivity, because fall speed variations relate mainly to particle density, not size, and because reflectivity is rather insensitive to particle size in the Mie regime. As a compromise, we have chosen to assume a mean fall speed of 1 m s$^{-1}$. This value was added to the WCR hydrometeor vertical velocities to obtain what we call WCR air vertical velocity $w$. This assumption carries an uncertainty of about ±0.5 m s$^{-1}$, which is small compared to typical updrafts observed within the LLAP band (see section 4).

b. Dual-Doppler synthesis

Radial velocities from the two downward-pointing WCR beams, oriented ~30° apart, can be utilized to obtain the 2D wind field in a quasi-vertical plane below the aircraft via DD synthesis (Leon et al. 2006). The Leon et al. (2006) DD synthesis technique was further refined by Damiani and Haimov (2006), whose software has been used in this study and in many others (e.g., Geerts et al. 2006, 2011, 2015; Yang and Geerts 2006; Miao and Geerts 2007; Sipprell and Geerts 2007; Damiani et al. 2008; French et al. 2015; Bergmaier and Geerts 2016). In particular, Bergmaier and Geerts (2016) recently used this technique to present evidence of secondary circulations within two shallow lake-effect snowbands over the New York Finger Lakes during OWLeS.

Details of the vertical-plane DD synthesis technique can be found in the above-mentioned literature. In short, a volume of scatterers is initially sampled during flight by the down-forward beam some distance ahead of the UWKA. A short time later (~1 s close to the aircraft and ~19 s at a range of 3 km) the nadir beam samples that same volume of scatterers as the aircraft passes overhead. The reflectivity-weighted mean radial velocities obtained by the beams are decomposed later during the DD processing phase to determine how much of the down-forward radial velocity was due to the horizontal wind along the beam. The vertical component is mostly obtained from the nadir beam. The along-track and vertical components compose the two resolved components of the 3D wind, while the third “unresolved” horizontal component, oriented normal to the down-forward beam, remains unknown. A guess for this unresolved component can be obtained from the wind profile of a nearby sounding, if one is available. The wind profile can also be used to correct for horizontal wind contamination of the DD radial velocities due solely to aircraft attitude fluctuations, especially the roll angle. The DD synthesis combines the sounding wind with the two resolved components to find an estimate of
the 3D wind vector for the sampled volume of scatterers. For a series of measurements along some flight transect, the 3D vectors are interpolated onto a vertical grid aligned with the aircraft ground track for display.

The description above assumes that the aircraft is flying parallel to the mean flow at flight level. In such a case, the aircraft heading will be aligned with the ground track, allowing the two beams to sample the same volume of scatterers. The resolved DD horizontal wind is effectively the component of the horizontal wind parallel to the aircraft ground track. This is the ideal case, since only the resolved winds, directly inferred by the WCR beams, are interpolated onto the grid. To account for the time lag between passage of each beam over a volume of scatterers, an assumption can be made regarding the advection velocity of the scatterers, which is then built into the construction of the grid (Damiani and Haimov 2006).

The calculation and interpretation of the resolved DD wind becomes more complicated when there is a significant crosswind at flight level. In this case, the aircraft must angle, or “crab,” into the wind to maintain the desired ground-relative flight track. By crabbing, the aircraft heading and ground track are no longer aligned (yielding a “drift angle”) and the two beams do not sample the same spatial volume. Instead, the down-forward beam points to the left or right of the flight track. More importantly, the resolved component of the horizontal wind no longer lies along the ground track, but instead along the aircraft heading, since that is the direction in which down-forward beam is pointed. As such, the resolved wind will contain a significant component of the cross-track wind that, if strong enough, could overwhelm the along-track wind. In this situation, the unresolved component of the wind lies normal to the aircraft heading, not the ground track. Thus, projection of the 3D wind vector onto a grid oriented parallel to the ground track (to obtain the along-track wind component) will inherently include some portion of the unresolved wind, increasing the uncertainty in the DD analysis.

For this study, we are interested in the along-track wind since the flight tracks were designed to fly across the LLAP band, parallel to the cross-band flow, with the hope of capturing the secondary circulation. The DD measurements were therefore projected onto a straight 2D Cartesian grid between the start and end points of each flight leg. The DD analysis package also allows for the gridding of data on a 3D “curtain” grid that follows the flight track, which is ideal when the flight track deviates significantly from straight and level (Damiani and Haimov 2006). The choice of a 2D grid is preferred here, in order to isolate the cross-band wind, and is also appropriate, since the legs were quasi straight and flown at a near-constant altitude.

The approximate horizontal and vertical resolution of the grids was 90 and 60 m, respectively, indicating that some smoothing was applied to the velocities. During IOP2b, strong crosswinds of $\sim 20$–$25$ m s$^{-1}$ at a flight level (3.0 km MSL) yielded mean (maximum) drift angles of $\sim 14^\circ$ ($17^\circ$) during each leg. The drift angle varied little because the flight-level crosswinds were rather uniform. Given these angles, the mean (maximum) horizontal offset of the ground-relative down-forward beam track from the nadir beam track was about 400 m (500 m) at the maximum relevant vertical range of 3 km (i.e., near the lake surface). In a Lagrangian (parcel-relative) sense, the distance between the scatterer volumes sampled by the two beams is in fact smaller than this, since air parcels first sampled by the down-forward beam are advected by the crosswind toward the volume of air sampled by the nadir beam. Nonetheless, we assume that over this short distance the 3D wind was homogeneous in the plane normal to the flight track. In other words, we assume that the LLAP band did not vary in the third dimension (along band), at least over a short distance ($\sim 500$ m). Thus, while the volume of scatterers sampled by the two beams may not have been the same, this assumption implies that the characteristics of the two volumes, with respect to reflectivity and velocity, were similar. Given that aircraft roll angles were typically within $\pm 2^\circ$, most of the DD velocity samples were obtained within 500 m of the ground track.

The same 2018 UTC sounding launched from Oswego (section 2a) was used to obtain an estimate of the unresolved wind component. This profile was also used to correct for horizontal wind contamination in the DD data due to aircraft attitude angle fluctuations, and is assumed to be representative of the environmental flow across all flight legs. The 3D DD wind vectors were projected onto the grid by means of a simple coordinate transformation matrix based on the mean drift angle, with the $x$ axis aligned with the UWKA ground track and the $z$ axis pointing up. The resulting along-track wind (hereafter $u_{\text{track}}$) is therefore positive (negative) away from (toward) the UWKA. To avoid confusion, and since all the legs in the study are plotted with north on the right, $u_{\text{track}}$ will always be displayed with positive values pointing north irrespective of the flight direction. (The aircraft flew back and forth across the band.)

It is important to stress that $u_{\text{track}}$ carries some uncertainty due to the numerous assumptions made during the DD synthesis, the main one being that the unresolved portion of the horizontal wind is represented by the wind from a proximiy sounding. This wind varied with height but was uniform in the along-track direction.
The use of different sounding wind profiles from around the lake (from Oswego, Sodus Point, New York, and Darlington, Ontario)\(^3\) at about the same time yields slightly different values of \(u_{\text{track}}\). The average (90th percentile) root-mean-square difference between these \(u_{\text{track}}\) estimates for all the flight legs analyzed in this study is 0.7 (1.2) m s\(^{-1}\). This measure of uncertainty is small compared to the typical magnitude of \(u_{\text{track}}\), which approached \(O(10)\) m s\(^{-1}\), as will be shown below.

c. WRF Model

As mentioned in the introduction, this study aims to not only describe the secondary circulation across a LLAP band, but also to dynamically interpret it. The UWKA data do not adequately describe the thermodynamic structure of the LLAP band, because the cross-band flight legs were too high (mostly at 3.0 and 1.7 km MSL, with two legs at \(\sim 1.1\) km MSL) and too short. Sounding data on opposite sides of this LLAP band were also incomplete, since no radiosondes were released to the north. Therefore, in order to obtain a complete dynamically and microphysically consistent depiction of this LLAP band, we resort to the Weather Research and Forecasting (WRF) Model simulation presented in Campbell and Steenburgh (2017) and described further in Steenburgh and Campbell (2017). The three model domains, with 12-, 4-, and 1.33-km grid spacing, are shown in Fig. 1a. The WRF output used in this study is from the 1.33-km domain, which is centered over Lake Ontario and the surrounding region of interest. Parameterization schemes utilized in the simulation that are of interest to this study include the Thompson cloud microphysics (Thompson et al. 2008) and Yonsei University planetary boundary layer (Hong et al. 2006) schemes. For a full description of the model and its configuration, we refer the reader to Campbell and Steenburgh (2017).

Cross sections with lengths of \(\sim 145\) km were obtained along the approximate UWKA flight tracks, extending about 50 km beyond each of the flight tracks’ end points to allow for better examination of the cross-band environmental characteristics. An arbitrary number of points were chosen to lie along the cross section, at which weighted means of model output variables were calculated using the four closest horizontal grid points in the model. The distance \(\Delta x\) between each of the points along the cross sections is \(\sim 1.2\) km. All data from the model half-\(\eta\) levels within the cross sections have been linearly interpolated to constant height levels with a \(\Delta z\) of 50 m. This was done to allow for the calculation of buoyancy, which is a function of height and is given by

\[
B = g \left( \frac{\theta_v - \bar{u}_v}{\theta_v} - r_h \right),
\]

where \(g\) is the gravitational acceleration at the surface of the earth, \(\theta_v\) is virtual potential temperature, \(\bar{u}_v\) is the mean \(u_v\) at a constant height level, and \(r_h\) is the mixing ratio of precipitation (rain, ice, snow, and graupel). The General Meteorological Package (GEMPAK) software\(^4\) was utilized to calculate kinematic frontogenesis \(F\). GEMPAK uses the form of the 2D horizontal frontogenesis equation that is given by

\[
F = -\frac{1}{2} |\nabla_p \theta|(D \cos 2b - \delta),
\]

where \(\nabla_p \theta\) is the horizontal potential temperature gradient along a constant pressure surface, \(D\) is the total deformation due to stretching and shearing, \(b\) is the angle between the axis of dilation and the isotherms, and \(\delta\) is the horizontal divergence.

d. Other data sources

Level-II 0.5° base reflectivity data from the KTYX WSR-88D S-band (10.7-cm wavelength) radar, located east of Lake Ontario on Tug Hill, was downloaded from the National Climatic Data Center (NCDC) Next Generation Weather Radar (NEXRAD) archive in level-II format (Crum et al. 1993). Three GPS-based radiosondes, two from Vaisala and one from GRAW, were launched by three OWLeS teams at \(\sim 2015\) UTC to obtain sounding profiles. Following the field campaign, the sounding data were quality controlled and archived by the National Center for Atmospheric Research (NCAR) Earth Observing Laboratory (EOL).

3. The IOP2b event: Background, horizontal structure, and environmental conditions

IOP2b took place from 2300 UTC 10 December to 0200 UTC 12 December 2013 (Kristovich et al. 2017), during which a strong LLAP system was present over eastern Lake Ontario and areas downwind of the lake, including Tug Hill (Fig. 1b). This system featured

\(^3\)The soundings from Sandy Creek, NY, and North Redfield, NY, were not used due to missing low-level wind data.

\(^4\)GEMPAK is available through the University Corporation for Atmospheric Research Unidata Program (http://www.unidata.ucar.edu/software/gempak/).
periods of both banded and cellular convection (Campbell et al. 2016). The 24-h snowfall totals ending at 0000 UTC 12 December ranged from 47.8 cm at Sandy Creek, New York (SC), to 101.5 cm at North Redfield, New York (NR), located on Tug Hill (Minder et al. 2015; Campbell et al. 2016). According to Campbell et al. (2016), the greatest snowfall rates of the event were observed when the LLAP system featured a highly organized, intense, LLAP-band structure from 1745 to 2101 UTC. The UWKA flight was conducted during this well-organized phase (1801–2138 UTC). Recent papers (i.e., Minder et al. 2015; Campbell et al. 2016; Welsh et al. 2016) describe the synoptic environment during IOP2b and around the time of the UWKA flight. In short, a deep upper-level trough was present throughout the event, allowing very cold arctic air to move over the relatively warm waters of Lake Ontario. Additionally, strong low-level unidirectional westerly flow and the presence of a deep unstable boundary layer over the lake led to conditions that were conducive for LLAP band development (Niziol et al. 1995). The well-organized band sampled by the UWKA occurred in the presence of an approaching upper-level shortwave trough. Cooling of the midtroposphere ahead of the upper-level trough coincided with deepening of the boundary layer, according to a sequence of soundings at NR (Campbell et al. 2016).

The UWKA flew 12 legs oriented in a north–south fashion across the band, the first five of which (Figs. 2a,c) were flown at 3.0 km MSL, near cloud top, from 1841 to 1936 UTC. Flying near cloud top made it possible, through vertical DD synthesis of the WCR radial velocity data, to resolve the 2D cross-band wind field over nearly the entire depth of the band. Horizontally, the band was well organized during these legs, appearing in KTYX radar imagery as one long, narrow linear feature extending from the middle of Lake Ontario eastward over Tug Hill and beyond (Figs. 2a,c). A closer look reveals that the distribution of radar echoes from west to east was not entirely continuous at 1849 UTC (Fig. 2a), as individual convective cells appear to have been embedded within the band over the lake. Also, a much weaker, less organized band was present to the north of the main band at this time. Between about 1900 and 2100 UTC [see Fig. 4b in Minder et al. (2015)] the band took on a nearly continuous appearance on radar from lake to land as it became more organized, as shown in Fig. 2c. Maximum reflectivity, observed over
Tug Hill, was >30 dBZ. The linear (as opposed to cellular) appearance of convection is important as it improves the two-dimensionality assumption made for the derivation of $u_{\text{track}}$ (section 2b).

These radar observations can be compared with WRF-derived reflectivity (Figs. 2b,d). The model reproduces the band and captures its overall horizontal size and linear structure fairly well, but is unable to match the observed band’s position; the model places the band farther south than observed, closer to the southern shoreline of eastern Lake Ontario. Interestingly, the model produces a second weaker band to the north of the main band, as observed early on by KTYX (Fig. 2a). Yet, the model sustains this band, whereas in reality it decayed (Fig. 2c). This secondary band appears to be associated with a line of surface convergence emanating from Point Petre on the north side of the lake (see Steenburgh and Campbell 2017).

Three soundings were obtained near the band around 2015 UTC (Fig. 3; for launch locations see Fig. 1b). One of these soundings, from Oswego, was launched south of the band, while another was launched from within the band at SC. The last sounding was launched from SC, which was located along the southern edge of the band during the time of the sounding. No sounding was obtained north of the band during IOP2b. In the lowest ~1 km, the Oswego and NR soundings were cooler than SC (in terms of $\theta_v$) by about 0.5–1.0 K, with NR being the coolest and most stable (Fig. 3a). In the lowest ~250 m, the Oswego sounding was actually warmer than SC, and unstable, probably since the launch location was only several hundred meters from the shoreline of the warm lake (as opposed to SC and NR, which were about 10 and 25 km from the shoreline, respectively). The $\theta_v$ profiles for Oswego and SC were nearly moist neutral in the lowest ~1 km (Fig. 3b), with SC being slightly warmer and closer to saturation [see Fig. 4 in Welsh et al. (2016)]. Above this, the SC sounding maintained a nearly moist-neutral profile up to ~3 km MSL, about the depth of the LLAP band. As with $\theta_v$, the $\theta_e$ profile at NR was cooler than at both SC and Oswego below 1 km MSL but warmer above 1.5 km.

At NR, just 16 km east-southeast of SC, this shallow (~0.5 km deep) cool layer, below a well-mixed saturated layer up to 3.2 km MSL, appears to be due to the northward advection of cooler overland air that had not been modified by the lake (Welsh et al. 2016; Steenburgh and Campbell 2017). Modeling work by Steenburgh and Campbell (2017) and Campbell and Steenburgh (2017) show that a land-breeze front had developed early in IOP2b along the southeastern shoreline of the lake and was still present at 1800 UTC, stretching northeastward across Tug Hill and undercutting the band. They also show that the southwest flow south of this land breeze (referred to in those papers and here as LBF2) experienced almost no modification by either Lake Ontario or Lake Erie, and was thus cooler than the air north of the land breeze (i.e., at SC), which had traveled some distance over Lake Ontario. Observed low-level winds were out of the west-southwest at both Oswego and NR, but unfortunately were missing at SC (Fig. 3). The observed ~0.5-km cold pool depth at NR corresponds well with
the modeled land-breeze depth over Tug Hill (section 4b). The LBF2 boundary will be revisited later.

4. Observations and modeling of LLAP band vertical structure

4a. WCR observations

WCR vertical transects from the five 3.0 km MSL flight legs are presented here. Legs 1–3 were situated over the east end of Lake Ontario while legs 4 and 5 were completed over land downwind of the lake (Fig. 2). Welsh et al. (2016) used similar transects of WCR reflectivity and vertical velocity to examine the west–east transition of this LLAP band. This study focuses on the WCR-derived 2D along-track (i.e., cross-band) wind field below the UWKA. To provide context for the DD analysis, and to allow direct comparison with WRF model output, WCR reflectivity and vertical velocity from the nadir and zenith beams are reexamined here. [The WCR observations are presented within Figs. 4–8, with reflectivity, estimated air vertical velocity \( w \), and the DD-derived \( u_{\text{track}} \) given in panels (a), (b), and (c), respectively.] The aspect ratio (vertical exaggeration) of these panels ranges between 2:1 and 3:1.

The LLAP band was already quite deep along leg 1, the westernmost of the five, with cloud tops nearly 3 km above the lake surface, near flight level (Fig. 4a). The band cloud tops remained nearly fixed at this altitude in all five legs (cf. Fig. 7 in Welsh et al. 2016). Several rather shallow (typically <1 km deep) updrafts of 3–5 m s\(^{-1}\) were intercepted offshore, mainly along legs 1 and 2 (e.g., at 43.55° and 43.70°N in Fig. 4b). We describe these updrafts as convective, as they were strong enough to loft hydrometeors (Houze 2014). This lofting is evident in Fig. 4a by the coincident depressions in WCR reflectivity (e.g., at 43.55° and 43.70°N). The KTYX radar detected some of these shallow convective cells to the north of the emerging LLAP band (Fig. 2a), although most cells simply were immersed in the deeper LLAP band, and too shallow for KTYX to detect. The band was accompanied by a well-defined secondary circulation along leg 1, which consisted of two opposing vortices on the either side of the band (Fig. 4c). While the low-level flow never changed sign across the length of the leg, the strong 9–10 m s\(^{-1}\) southerly inflow on the south side of the band weakened to \( \sim 5-6 \) m s\(^{-1}\) on the north side, with the main zone of convergence located near \( \sim 43.67\)°N, just south of one of the stronger updrafts. This inflow layer was <1.5 km deep, with the strongest winds occurring in the lowest 500 m. A clear divergence pattern is seen aloft from 1.5–2.0 km MSL. Presumably, the low-level converged flow associated with the circulation helped draw the shallow convective cells toward the main LLAP band updraft. The strength of the secondary circulation will be discussed in further detail at the end of section 4.

Farther east along leg 2, the band began to strengthen as a rather erect dominant updraft developed within the band core, peaking at \( >5 \) m s\(^{-1}\) (Figs. 5a,b). The secondary circulation exhibited better structure along this leg, with a clear convergence/divergence signal in the DD data (centered near \( \sim 43.64\)°N) and a horizontal vorticity couplet, depicted well by the wind vectors, that straddled the main updraft (Fig. 5c). The magnitude of the stronger inflow/outflow to the north of the updraft was about \( \pm 5-6 \) m s\(^{-1}\), which is not insignificant when compared to the prevailing \( \sim 15-25 \) m s\(^{-1}\) westerly flow [as seen in surface observations near the shoreline (not shown); also cf. Fig. 3] that drove the LLAP band onshore. A second area of weaker low-level convergence was present at \( \sim 43.58\)°N, south of the primary updraft and coincident with another shallower updraft. This convergence was probably associated with a shallow land-breeze front [hereafter LBF1, consistent with Steenburgh and Campbell (2017) and Campbell and Steenburgh (2017)] that is seen in the model data (see section 4b). The curved appearance of reflectivity here (between 43.5° and 43.55°N in Fig. 5a) supports the idea that this was a land-breeze front, with the hydrometeors falling along the slanted leading edge of the boundary.

The two convergence zones observed along leg 2 (located just off the lake’s eastern shore) were also present along leg 3, near 43.65° and 43.7°N (Fig. 6c). Here, the secondary circulation was characterized most prominently by strong southerly inflow approaching 10 m s\(^{-1}\) (Fig. 6c) and an intense 9 m s\(^{-1}\) updraft (near 43.7°N in Fig. 6b). Much weaker southerly low-level flow extended north of the updraft to about 43.75°N. It is likely that leg 3 did not extend far enough toward the north side of the lake for the northerly inflow to be sampled by the WCR. The updraft was located directly above the northernmost surface confluence zone and produced a strong bounded weak-echo region (BWER) almost reaching flight level (Fig. 6a). At flight level, in situ vertical velocity exceeded 10 m s\(^{-1}\). This updraft signifies the upward branch of the dual-vortex circulation. As will be shown in section 5, the shallow convergence zone near 43.65°N and the strong southerly inflow were likely associated with one, or both, of the land-breeze fronts, since the model data suggest that LBF1 and LBF2 may have merged along this leg.

Along leg 4—flown just east of the lake—the LLAP band assumed a less convective, broader, and more stratified appearance since it was removed from its source of heat and moisture (Figs. 7a–c and 8a–c). Widespread ascent occurred across the band (Fig. 7b), due to a combination of frictional convergence, isentropic
FIG. 4. A six-panel comparison of WCR observations along leg 1 (1841–1848:30 UTC) and WRF output along model cross section A–A’ (at 1840 UTC). Shown from the WCR are (a) reflectivity, (b) air vertical velocity $w$, and (c) $u_{\text{track}}$ and wind vectors. Shown from WRF are (d) model-derived reflectivity (shaded) and $\theta_e$ (contours every 1 K), (e) $w$ (shaded) and $\theta_e$ (contours every 1 K), and (f) $u_{\text{track}}$ and 2D wind vectors. The components of the wind vectors are $u_{\text{track}}$ and $w$. The vertical component of the vectors has been slightly exaggerated. UWKA flight level is given by the white dashed line in (a) while in situ measurements of $w$ and $u_{\text{track}}$ are color coded at flight level in (b) and (c), respectively. In (d)–(f), the UWKA flight track is plotted as a black or red dashed line and the ground (lake) is shaded in gray (blue). Mean horizontal vorticity values from within the light blue boxes in (c) and (f) are given in Table 1.
lift, and purely orographic ascent (Welsh et al. 2016; Campbell and Steenburgh 2017). While the main updraft along leg 4 was weaker than along leg 3, the secondary circulation pattern was still evident in the DD observations and quite strong (Fig. 7c). The LLAP band reflectivity exhibited somewhat of an anvil structure (Fig. 7a), with hydrometeor streaks tilting toward the core at low levels and deflecting outward aloft. The

Fig. 5. As in Fig. 4, but for leg 2 (1853–1900:45 UTC) and model cross section B–B′ (1900 UTC).
shallow convergence zone to the south in previous legs, presumably associated with one or both of the land-breeze fronts, was no longer evident as a separate entity along leg 4. Presumably, it had merged with the broader LLAP band convergence at 43.74°N. Further evidence for this interpretation comes from the tilting of this convergence zone with height between 43.7° and 43.75°N in Fig. 7c, toward the colder side of the land-breeze front.
(i.e., the left side). The shallow \( \sim 500 \)-m-deep leading edge of the southerly flow (i.e., at 43.74\(^\circ\)N) resembles a density current head, capped by front-to-rear northerly flow (shaded in blue) that rose over this leading edge within the primary updraft. This suggests a significant buoyancy gradient across the boundary, which WCR data cannot confirm, but is consistent with the lower \( \theta_e \) observed at NR (section 3). NR is
located ~10 km east of leg 4 at a latitude of 43.625°N, south of the boundary.

Over Tug Hill along leg 5—about 30 km inland from the Lake Ontario shoreline—the reflectivity and vertical velocity fields indicate continued broadening of the LLAP band and weakening of the main updraft (Figs. 8a,b). The convergence zone was less clearly tilted toward the left, possibly indicating weaker baroclinicity.
(Fig. 8e). However, the convergence/divergence pattern associated with the secondary circulation was remarkably coherent, perhaps more so than along the other legs. This may have been aided by the suspected land-breeze front seen in the previous two legs, which would have helped strengthen the low-level southerly flow and enhance low-level convergence over Tug Hill. The persistence of the circulation undoubtedly contributed to the deep inland penetration of the LLAP band and associated lake-effect snowfall.

To estimate the strength of the circulation along each leg, we have calculated the mean horizontal vorticity \( \omega_{\text{horiz}} \) from the measurements of WCR \( u_{\text{track}} \) and \( w \) within a pair of rectangular areas, shown as light blue boxes in panel (c) in Figs. 4–8. This was repeated with WRF \( u_{\text{track}} \) and \( w \) (obtained along the model cross sections) using the boxes shown in panel (f) in the same figures. The model results will be discussed in section 4b. Here, the horizontal vorticity refers to the component normal to the flight track, given by

\[
\omega_{\text{horiz}} = \frac{\partial w}{\partial x} - \frac{\partial u_{\text{track}}}{\partial z},
\]

where \( x \) is positive toward the right (i.e., north) in Figs. 4–8. The mean vorticity (Table 1) is equal to the circulation multiplied by the area of the box (Stokes’s theorem). In other words, it is the circulation strengthened normalized by the box size. In each panel, we have attempted to position the pair of boxes so that they adequately sample the two vortices composing the secondary circulation. Each WCR-based box (panel c) has a width of 0.15° latitude, except along legs that did not extend far enough to the north for the WCR to fully sample the northern vortex (i.e., legs 1 and 3). The WRF-based boxes [panel (f)] all have widths of 0.3° latitude. All boxes extend from just above the surface (e.g., 0.2 km MSL for legs 1–3, 0.4 km MSL for leg 4, and 0.6 km MSL for leg 5) up to 2.5 km MSL.

The observed mean \( \omega_{\text{horiz}} \) within both the southern (dual-Doppler S box) and the northern (dual-Doppler N box) circulation vortices strengthened from west to east over the water, between legs 1 and 3, before weakening slightly over land (Table 1). Each reached its maximum over the lake, the southern vortex along leg 3 and the northern vortex along leg 2. This is consistent with the stronger vertical shear observed in the DD data along those legs (cf. Figs. 5c and 6c). We must be clear that these results are highly dependent on the north–south positioning of the boxes used in this analysis. However, they do indicate that the observed secondary circulation strengthened over the water toward the downwind shore, and was only slightly weaker inland.

### Table 1. Dual-Doppler and WRF mean horizontal vorticity (10^{-3} s^{-1}) from within the areas bounded by the light blue boxes shown in panels (c) and (f) in Figs. 4–8 (flight legs 1–5).

<table>
<thead>
<tr>
<th>Leg S box N box</th>
<th>Cross section S box N box</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 1.6 –2.2 1.9</td>
<td>A–A’ 3.2 –1.8 2.5</td>
<td></td>
</tr>
<tr>
<td>2 0.8 –4.1 2.5</td>
<td>B–B’ 4.1 –1.9 3.0</td>
<td></td>
</tr>
<tr>
<td>3 4.0 –2.8 3.4</td>
<td>C–C’ 3.6 –2.5 3.1</td>
<td></td>
</tr>
<tr>
<td>4 2.9 –2.9 2.9</td>
<td>D–D’ 4.1 –3.1 3.6</td>
<td></td>
</tr>
<tr>
<td>5 2.7 –2.9 2.8</td>
<td>E–E’ 2.9 –3.2 3.1</td>
<td></td>
</tr>
</tbody>
</table>

#### b. WRF simulations

The finescale radar observations from legs 1–5 are compared with corresponding output variables from the WRF simulation along the cross sections identified in Figs. 2b,d in order to assess how well the model captures the overall structure and evolution of the band. Model composite reflectivity, \( w \), and \( u_{\text{track}} \) along each cross section are shown in panels (d)–(f) of Figs. 4–8. These can be directly compared to the WCR observations in panels (a)–(c). Since the cross sections are more than 3 times as long as each of the flight tracks, the aspect ratio of these panels is about 7:1. Furthermore, the proximity of the model cross sections to LBF1 and LBF2 is shown in Fig. 9.

To a first degree, the model results and observations appear to correspond fairly well. As demonstrated earlier in Figs. 2b,d, the model was successful at reproducing both the main band as well as a smaller, secondary band to the north. In cross sections A–A’, B–B’, and C–C’ (corresponding to legs 1–3), the northern band is distinct from the primary band (Figs. 4d, 5d, and 6d), but merges with it along cross sections D–D’ and E–E’ (i.e., legs 4 and 5) over land (Figs. 7d and 8d). A band similar to this modeled one, although weaker, can be seen around this time in the KTYX reflectivity imagery, just out of reach of the WCR transects (cf. Fig. 2a). The model also places the main LLAP band closer to the southern shore of Lake Ontario, near the location of LBF1. This is about 10 km south of the primary convergence zone identified in the DD analysis during the first three legs (cf. Figs. 4c–6c).

The maximum reflectivity within the band is larger in the model (~25–30 dBZ) than in the WCR data (~20 dBZ), since the reflectivity calculations using model output assume a 10-cm wavelength beam. Therefore, only reflectivity patterns should be compared, not values. Model vertical motions are expected to be weaker than those observed by the WCR, mainly because of the large difference in resolution. The shortest WRF-resolved wavelength is about 6 times the horizontal grid spacing.
showing a distinct, erect updraft of convective strength permitted in this WRF simulation, and WRF does the offshore transects (e.g., Figs. 4b and 5b). Convection is reproducing the secondary circulation and its basic broad, relatively weak updraft along leg 5 (cf. Figs. 8a,b). The updraft and much weaker northern band originating from Point Petre, which moved north of the main band. LBF2 was also beginning ascent along LBF2, which had completely undercut and lowered the LLAP band, while the northernmost one is associated with heavy LE snowfall during IOP2b. Two such boundaries, which they describe as land-breeze fronts (i.e., thermally driven, shallow boundaries in the generation of heavy LE snowfall during IOP2b. Two such boundaries, which they describe as land-breeze fronts (i.e., LBF1 and LBF2), are evident in the model data along the southern shore of Lake Ontario at 1910 UTC (Fig. 9).

Although the model is not perfect in its simulation of the band, it captures the general characteristics quite well. The good correspondence between the model and observations, especially with regard to the vertical and horizontal flow, suggests that the model can further be used to investigate the dynamics of the secondary circulation and, in particular, the mechanisms driving the flow around the band. This analysis is carried out in section 5.

5. Dynamics of the secondary circulation

Steenburgh and Campbell (2017) highlight the role of thermally driven, shallow boundaries in the generation of heavy LE snowfall during IOP2b. Two such boundaries, which they describe as land-breeze fronts (i.e., LBF1 and LBF2), are evident in the model data along the southern shore of Lake Ontario at 1910 UTC (Fig. 9). LBF1 emerges from an inflection point along the shoreline and serves as the locus for development of the broader LLAP system (Steenburgh and Campbell 2017),

or ~8 km. Thus, the model cannot be expected to resolve the WCR-observed shallow convective updrafts seen in the offshore transects (e.g., Figs. 4b and 5b). Convection is permitted in this WRF simulation, and WRF does produce a distinct, erect updraft of convective strength (~3 m s⁻¹ along C–C’) above the main confluence zone, nearly doubling in strength from A–A’ to C–C’ (Figs. 4e and 6e). This updraft is linear, rather than cellular, which agrees with observations given the lack of cellularity in the LLAP band as seen in KTYX (Figs. 2a,c) and Doppler on Wheels (not shown) base reflectivity imagery. As in the observations, the main LLAP band updraft in the model is rather narrow over the lake. However, the model updraft does not broaden and weaken as much over land along D–D’ (Fig. 7e), and even appears to split into two separate weak updrafts over Tug Hill along E–E’ (Fig. 8e). Campbell and Steenburgh (2017) suggest that the southernmost updraft along E–E’ might be part of the original LLAP band, while the northernmost one is associated with ascent along LBF2, which had completely undercut and moved north of the main band. LBF2 was also beginning to interact with, and perhaps enhance, the updraft of the weaker northern band originating from Point Petre, which was still separate from the main LLAP band at this time. In any case, it is not clear whether this agrees with the WCR observations, which show only one band with one rather broad, relatively weak updraft along leg 5 (cf. Figs. 8a,b).

Perhaps most importantly, the model does very well at reproducing the secondary circulation and its basic structure (Figs. 4f–8f). The depth of the low-level convergent flow is slightly shallower than was observed, only about 1 km deep. However, the magnitude of the modeled circulation was similar to the observations for both the southern and northern vortices (Table 1). The two vortices were more matched in strength (i.e., the circulation was more symmetric) than observed, although observations were hampered by the limited flight leg lengths. In addition, while the mean $\omega_{\text{horiz}}$ within the southern vortex of the circulation varies between about $3 \times 10^{-3}$ and $4 \times 10^{-3}$ s⁻¹, the northern vortex gradually strengthens from west to east (A–A’ to E–E’). The circulation hardly weakened over Tug Hill (Table 1). Remarkably, the model also captures the observed convergence zone tilt along D–D’ (cf. Figs. 7f and 7c). The model $\theta_v$ contours (Fig. 7d) indicate that this tilt is due to a colder (by ~1–2 K) shallow air mass behind LBF2, which is wedging underneath warmer air along the convergence boundary and lifting the northerly flow within the updraft. Since the modeled LLAP band is located slightly south of the observed location (Fig. 2), the modeled land breeze and the main LLAP band are collocated over the lake. The updraft and much weaker convergence zone within the smaller northern band emanating from Point Petre also coincide with a horizontal $\theta_v$ gradient in the model (e.g., near 43.7°N in Figs. 4e–6e), suggesting the presence of a second shallow airmass boundary to the north, out of range of WCR observations. Steenburgh and Campbell (2017) do not classify this northern boundary as a land-breeze front, however, since the baroclinicity is fairly weak compared to LBF1 and LBF2 (also see Fig. 9).

![Figure 9. Map of WRF output at 1910 UTC from the lowest half-$\eta$ level (~25 m AGL) centered over southeastern Lake Ontario showing $\theta_v$ (shaded), the 20-dBZ reflectivity contour (thick white lines), and horizontal wind vectors. Terrain contours every 250 m MSL are given by the thin black lines. The locations of the cross sections for legs 1–5 are given by the thick black lines. The two land-breeze fronts mentioned in the text, LBF1 and LBF2, are labeled and marked by the dashed blue lines.](image-url)
which ultimately concentrates into the intense LLAP band described here. LBF2 develops along the southeast corner of the lake and undercuts the LLAP system at an angle over Tug Hill. The formation of both boundaries is attributed to differential surface heat fluxes affecting respective parcel trajectories some distance upstream, as well as differential surface roughness (Steenburgh and Campbell 2017). The thermal contrast across LBF1 gradually weakens downstream of the shoreline inflection point, since the colder air mass is heated more by the underlying open water (Fig. 9). Meanwhile, LBF2 is strongest just inland. Its presence is validated by the thermal contrast observed between the SC (warm side) and NR (cold side) soundings (Fig. 3), and implied by the DD data along legs 3–5 (Figs. 6c, 7c, and 8c). Figure 9 also suggests that the two land-breeze fronts merge near C–C′ (i.e., near leg 3), consistent with the strengthening observed in the southerly inflow along leg 3. What is clear from Fig. 9 is that the strongest θυ gradient along C–C′ is associated with LBF2 (cf. Fig. 6e).

These land-breeze fronts result in significant 2D (horizontal) frontogenesis, and thus, over time, a thermally direct circulation with ascent on the warm side. Steenburgh and Campbell (2017) show that frontogenesis associated with the downstream extension of LBF1 extends across the lake to the eastern shoreline at 1800 UTC, decreasing slightly with the weakening temperature gradient over the water (see their Figs. 5c and 11). This frontogenesis is captured along A–A′ (Fig. 10a) and B–B′ (Fig. 10b), while stronger frontogenesis associated with LBF2 is found along C–C′, near the southern shore of Lake Ontario (Fig. 10c). Frontogenesis is also found along the weaker temperature gradient north of the LLAP band associated with the convergence line emanating from Point Petre (e.g., near leg 3), consistent with the strengthening observed in the southerly inflow along leg 3. What is clear from Fig. 9 is that the strongest θυ gradient along C–C′ is associated with LBF2 (cf. Fig. 6e).

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However, this frontogenetic circulation alone does not explain the depth of the LLAP band or the intensity of the resulting snowfall. For a LE band that is driven predominantly by such a shallow circulation (i.e., a classic shoreline band), in the absence of additional sources of buoyancy, the updraft—and the snowband—would presumably be fairly shallow, limited in depth by the stable air aloft. Yet, in the model the snowband is clearly 2 or 3 times deeper than the land-breeze front (depth of approximately 3 km or less) in all five legs, with its main updraft penetrating well into the stratified layer above (e.g., Figs. 6d,e). Embedded within the band is the secondary circulation, with two opposing horizontal vortices, which extends deeper than the frontogenetic zone to a height of approximately 3 km MSL (Figs. 4f–8f). Even though the circulation initiates along LBF1 earlier in IOP2b, it is sustained and enhanced solenoidally by the horizontal buoyancy gradient across the band, where Δθυ is 1–2 K (Figs. 4e, 5e, and 6e; see section 2c for buoyancy formula). This buoyancy gradient can also be seen in Fig. 11 and is strongest along the first three cross sections over the lake. Such a circulation is reminiscent of the convective circulation near the top of thunderstorms (e.g., Damiani et al. 2006; Morrison 2016). Over the lake, θυ is larger near the water surface than a few hundred meters higher (Figs. 4d, 5d, and 6d), implying surface-based CAPE. But this CAPE is quite shallow, at most 1.5 km deep. Sustained convective ascent along the LLAP band overshoots and lifts the equilibrium level from west to east.

As is the case in thunderstorms, buoyancy is generated and maintained by latent heat release in the updraft. Past modeling studies have shown that the role of latent heat release (by both condensation and freezing) in LE snowbands is to strengthen the updrafts and convective circulations (Ballentine 1982; Hjelmfelt and Braham 1983; Hjelmfelt 1990; Onton and Steenburgh 2001). Latent heating in this study occurs mainly in the narrow updraft column in all five cross sections, increasing from A–A′ to C–C′ and remaining strong in D–D′ (Figs. 11a–d). The divergent flow at upper levels disperses this heat (and also the hydrometeors) sideways. (Note that the buoyancy distribution at higher levels in Fig. 11 simply reflects the large-scale westerly thermal wind.) This latent heating, reaching up to approximately 3 km along C–C′ (Fig. 11c), is essential to explain the depth, intensity, and persistence of the LLAP band (Hjelmfelt and Braham 1983; Hjelmfelt 1990). Farther downstream along E–E′, latent heating is weaker and shallower yet nonetheless still contributing to the buoyancy within the band (Fig. 11c). The amount of latent heating relates to the adiabatic liquid water content, which in turn relates to the cloud base (i.e., LCL) temperature. This temperature is at least partially determined by the sensible and latent heat fluxes from the lake surface, which are a function of the surface wind speed and lake surface temperature. All these were relatively high in IOP2b.

6. Discussion

The observations and modeling results shown here provide a unique look at the kinematic and dynamic structure of an intense, well-organized LLAP band over Lake Ontario during the OWLeS campaign. As is shown by Steenburgh and Campbell (2017), shoreline
geometry, and its effect on differential surface roughness and the development of land-breeze fronts, was a significant factor in generating the primary convergence zones associated with this LLAP system. The relationship between these land-breeze fronts and the distribution of heavy snowfall over Tug Hill is examined by Campbell and Steenburgh (2017). The present study focuses on the west–east transition of the vertical

Fig. 10. WRF output showing total frontogenesis (shaded) and $\theta_e$ (contours every 1 K) along cross sections (a) A–A’ at 1840 UTC, (b) B–B’ at 1900 UTC, (c) C–C’ at 1910 UTC, (d) D–D’ at 1920 UTC, and (e) E–E’ at 1930 UTC.
secondary circulation within the band. In particular, we examine the mechanisms responsible for deepening and intensifying the circulation as the band moves across the lake, and for maintaining it as the band moves over land.

Vertical-plane DD radar data indicate a strong secondary circulation across the LLAP band, sustaining a strong updraft in the band’s core. While this circulation started as a shallow baroclinic circulation within LBF1...
along Lake Ontario’s southern shore, it was considerably deeper near the eastern (downwind) end of the lake. The depth and circulation strength of this well-organized LLAP band is attributed to solenoidal forcing, specifically the horizontal buoyancy gradient enhanced by latent heat release in the updraft. The continued presence of the land-breeze fronts acted to enhance the circulation by strengthening the low-level convergence. Ultimately, the picture painted here, and to a lesser extent in Steenburgh and Campbell (2017), is that the IOP2b LLAP band was driven simultaneously by both baroclinic (frontogenetic) and solenoidal (convective) forcings throughout its life cycle. The findings of Steenburgh and Campbell (2017) imply that near-surface frontogenetic forcing was likely responsible for the upstream development of both the band and the secondary circulation along LBF1 near the south shore of the lake. This frontogenetic forcing rapidly weakened toward the eastern end of the lake. The work presented here observationally confirms the existence of both land-breeze fronts and documents an impressive solenoidal circulation centered on a convective updraft. Using coincident model output, we find that the secondary circulation and LLAP band intensified toward the eastern end of the lake, due primarily to the contribution of latent heating. This latent heating both strengthened and deepened the solenoidal forcing and, along with the enhancement of low-level convergence from LBF2, helped to maintain the circulation within the band well inland over Tug Hill. The interplay between these mechanisms may differ in other events, or even during different phases of the IOP2b LLAP system. This raises a number of general questions. How do these two mechanisms interact to determine whether a LLAP system is weakly organized, with more scattered convection (as was the case earlier in IOP2b and in other LLAP systems observed during OWLeS), or well-organized with a narrow, continuous band of convection? Does the strength of the secondary circulation control the transition from weakly banded organization, in which convective cells experience a life cycle (i.e., cellular convection), to a well-defined band, in which the low-level convergence, a mostly undilute updraft, and upper-level divergence are persistent (i.e., linear convection)? Is the strength and depth of the secondary circulation tied to the strength of land-breeze fronts, which were stronger during IOP2b at 1200 UTC (when the system featured weaker bands or nonbanded modes) than they were at 1800 UTC (when the band was strengthening), or rather to the strength of the solenoidal forcing (buoyancy), as is suggested by model output examined here? Finally, how does this forcing relate to CAPE, and thus the upstream temperature profile, lake surface temperature, and surface wind speed? These questions become even more complex when considering that other environmental factors may have also contributed to the evolution and intensification of the LLAP band in this study. For example, what role did the passage of several upper-level shortwave features between 1200 and 1800 UTC (Campbell et al. 2016) play in the band’s intensification, and in the evolution and deepening of the circulation?

Campbell et al. (2016) note that this LLAP band produced copious snowfall east of Lake Ontario during its well-organized phase, both over the lowlands and over Tug Hill farther downwind. The strong updraft in such a band may deepen the convective boundary layer, which is critical for the development of heavy LE snow (e.g., Byrd et al. 1991). In a previous study of the IOP2b LLAP band, Welsh et al. (2016) attribute the heavy snowfall downwind of the lake to a combination of isentropic lift (related to frictional convergence, surface heat fluxes, and terrain) and the collapse of convection which had suspended hydrometeors while over open water. In addition, Campbell and Steenburgh (2017) discuss the impacts of LBF2 on the distribution of precipitation over and around Tug Hill. We propose that the presence of a strong secondary circulation, which clearly had some inertia as the LLAP band moved off the lake, was also important for allowing heavy snowfall within the band to penetrate well inland.

7. Conclusions

This study documents the kinematic structure of a well-organized LLAP band observed in IOP2b during OWLeS. The main findings, based mainly on airborne vertical-plane dual-Doppler radar data from the Wyoming Cloud Radar and WRF Model output, are as follows:

- Near the eastern (downwind) end of Lake Ontario, the LLAP band exhibited a strong secondary (cross band) circulation, with \( \approx 10 \text{ m s}^{-1} \) low-level inflow, a persistent convective updraft in the core (peak strength of \( \approx 7–10 \text{ m s}^{-1} \)), and upper-level divergence just below cloud top, about 3 km high. The circulation intensified toward the downwind shore and only gradually weakened inland.
- While shallow frontogenetic (land breeze) forcing was present in the model along a land-breeze front (LBF1) near the southern shoreline of the lake, the deepening and intensification of the secondary circulation from west to east as revealed by the WCR and seen in the WRF output was driven primarily by solenoidal (buoyancy) forcing, especially near the eastern end of the lake where the near-surface horizontal temperature gradient was weaker. This forcing was enhanced by latent heat release within the band’s persistent
updraft, over a depth exceeding the depth of the land-breeze front. The strong secondary circulation concentrated and linearly aligned the convective updrafts and latent heat release, explaining the organization of convection into a deep, narrow, and intense band.

- This secondary circulation was maintained downwind of the lake, partially due to the enhancement of surface convergence by a second land-breeze front (LBF2), and thus was likely a factor in the heavy LE snowfall observed over land during this event.

These findings may not generally apply to LLAP systems. The WCR observations presented here are from a period of <1 h during which the band was most intense and well organized. They may not be representative of when the band was less organized and weaker, or when the LE system featured nonbanded modes. The role of the secondary circulation and latent heating within deep, intense LLAP bands like the one examined here remains not fully understood. Thus, further work into the nature of LLAP systems and their dynamics is warranted, given the rich observational dataset provided by the OWLeS campaign.

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