Airborne Radar Observations of Lake-Effect Snowbands over the New York Finger Lakes

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(Manuscript received 14 March 2016, in final form 12 July 2016)

ABSTRACT
The vast majority of lake-effect snow research throughout the years has focused on the North American Great Lakes since they are often associated with strong lake-effect events that produce heavy downstream snowfall. This study investigates a lake-effect snow event that instead occurred over two smaller lakes, the New York Finger Lakes, which are just $O(5)$ km wide and $O(50)$ km long. A pair of well-defined snowbands that formed over Seneca and Cayuga Lakes, the two largest of the Finger Lakes, were sampled from above by a vertically pointing Doppler radar and lidar on board the University of Wyoming King Air (UWKA). With typical widths matching the widths of the lakes, and depths of less than 1000 m, the long-lake-axis-parallel bands were actually quite intense for their size. For example, updrafts of 2–3 m s$^{-1}$ or greater within the band cores were common, and reflectivity occasionally exceeded 5 dBZ. Airborne dual-Doppler data show that both bands were sometimes accompanied by a well-defined thermally driven secondary circulation. Lidar data reveal that the Cayuga Lake band contained significantly more liquid water than the band over Seneca Lake, which was composed mainly of ice. Dissipating lake-effect ice clouds, carried downstream from Lake Ontario toward Seneca Lake, likely seeded the emerging convection over Seneca Lake, resulting in an accelerated depletion of liquid in the Seneca Lake band via more efficient snow growth.

1. Introduction
Much attention has been devoted to understanding how large lakes (e.g., the North American Great Lakes) influence the weather and climate of the surrounding region. Lake-effect (LE) convection, which typically ensues when cold air masses migrate over relatively warmer open water during the fall and winter, often leads to heavy snowfall downwind of the lakes impacting local travel and commerce. More than a half century of research related to LE snowstorms has yielded significant insight into the processes controlling their development and evolution (e.g., Wiggin 1950; Peace and Sykes 1966; Holroyd 1971; Lavoie 1972; Kelly 1982; Byrd et al. 1991; Niziol et al. 1995; Laird et al. 2003; Veals and Steenburgh 2015; Welsh et al. 2016). For instance, it is well understood that heat and moisture fluxes between a warmer body of water, such as a lake, and colder ambient air above will gradually lead to warming and moistening, and therefore destabilization, of the lower boundary layer over the water. Over time (or with ‘‘fetch’’), the depth of the boundary layer increases, either through continued modification by the lake or by other processes such as thermally driven surface convergence or even synoptic forcing, to the point that clouds and precipitation form. Over the North American Great Lakes, a lake-to-850-hPa temperature difference of at least 13 K (i.e., the dry adiabatic lapse rate) is typically required for the development of LE precipitation (Holroyd 1971). The fetch over the lake and the alignment with the lakes’ long axis helps determine the morphology and intensity of snowbands that develop (Wiggin 1950; Lavoie 1972; Hjelmfelt 1990; Niziol et al. 1995). Under larger fetch conditions where the mean boundary layer flow is closely aligned with the long axis of a lake (in particular, the rather long and narrow eastern Great Lakes, Erie and Ontario), LE convection will typically organize into one long and often intense band oriented parallel to the wind. In such a situation, the boundary layer over the center of

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DOI: 10.1175/MWR-D-16-0103.1
the lake is afforded more time to destabilize and deepen (Byrd et al. 1991; Kristovich et al. 2016), allowing updrafts to become more buoyant and the thermal contrast between the air over the lake and surrounding land to increase. These factors inevitably yield a land-to-lake pressure gradient over the lake, inducing the development of a low-level convergence zone that acts to enhance the updrafts and often helps focus the LE precipitation into an organized band.

Under a weaker wind regime, the flow resulting from a lower land-to-lake temperature difference (i.e., a "land breeze") may drive the formation of a LE band on its own (Passarelli and Braham 1981). In both cases, a secondary circulation may emerge (e.g., Steiger et al. 2013), with divergent flow aloft and descending air on the flanks of the band, further strengthening the main updraft and helping to maintain the band as it moves onshore.

While much of the LE snow research over the years has focused on the Great Lakes, LE snowfall originating from smaller lakes has also been studied. Included in these lakes is the Great Salt Lake (Carpenter 1993; Steenburgh et al. 2000; Alcott and Steenburgh 2013; Yeager et al. 2013), Lake Champlain (Payer et al. 2007; Laird et al. 2009a), and Lake Tahoe (Cairns et al. 2001; Laird et al. 2016). The New York Finger Lakes, although much smaller in size than any of the Great Lakes, occasionally produce LE snowfall as well. The 11 elongated lakes that compose the Finger Lakes have surface areas of 2.6–175 km² and lie about 50 km south of the much larger Lake Ontario (surface area \( \approx 19000 \text{ km}^2 \)) (Fig. 1a). Their general orientation is from north to south, although several lakes veer off to the southeast or southwest. The two largest lakes, Seneca (SL) and Cayuga (CL) Lakes, are about 60 km long, less than 5 km wide, and have mean (maximum) depths of 89 (188) and 55 (133) m, respectively. Laird et al. (2009b, 2010) examined climatological aspects of Finger Lakes LE precipitation events over 11 winters. They found that, during those events, winds were typically quite weak (<5 m/s\(^{-1}\)) and out of the north or northwest. These wind directions are oriented parallel to the long axes of most of the Finger Lakes. Laird et al. (2009b) also found that, while heavier snowfall rates during Finger Lakes events were not altogether uncommon, the average event duration was relatively short (<10 h), yielding downstream snowfall totals that were typically quite meager compared to the significant accumulations often seen with the more intense, long-lived LE storms over Lake Ontario and Lake Erie.

According to their study, the highest frequency of LE events was associated with SL and CL. Part of this is due to the fact that these two lakes usually remain relatively warm and free of extensive ice coverage throughout the winter since they are quite deep for their size.\(^1\)

There has been a renewed interest in LE snow in the last few years, in part due to the Ontario Winter Lake-effect Systems (OWLs) field campaign that was conducted during the winter of 2013–14 (Kristovich

\(^{1}\) Seneca Lake (and perhaps Cayuga Lake) is also apparently fed by underground springs, which help replenish the supply of warmer water throughout the winter. For more information see New York State Department of State (2015).
et al. 2016). Although OWLeS was particularly focused on LE snow events associated with Lake Ontario, one intensive observation period (IOP) was conducted over the Finger Lakes, during the morning of 22 January 2014. That IOP is the focus of this paper. Previous Finger Lakes LE snow studies (i.e., Laird et al. 2009b, 2010) were generally limited to observations from surface instrumentation, soundings, and output from model data. The OWLeS project provides a much richer set of observations, most notably finescale airborne in situ and remote sensing measurements over the Finger Lakes from the University of Wyoming King Air (UWKA) research aircraft.

The primary objective of this study is to document the 2D cross-band structure and downwind evolution of LE snowbands over SL and CL during this IOP. We primarily focus on snowband structure as it relates to precipitation and airflow (both in the vertical and across the band). This paper is organized in the following manner. Section 2 discusses the data and instrumentation used in the analysis. Section 3 provides more information on the event of interest and presents the UWKA observations. These findings are discussed in section 4. Finally, conclusions are drawn in section 5.

2. Data and instrumentation

a. Airborne in situ observations

The UWKA, designed for the study of lower-tropospheric phenomena, was outfitted with an ensemble of in situ sensors and several remote sensing instruments during OWLeS. The 3D wind at flight level is derived in real time through a method described by Brown et al. (1983) using dynamic and differential pressure measurements obtained by a five-hole gust probe on the nose boom ahead of the aircraft. This technique has been improved using GPS aircraft attitude and velocity measurements (Haimov and Rodi 2013). Other standard meteorological variables (i.e., temperature, air pressure, and humidity) are measured as well. Several optical array probes that measure particle size distributions and concentrations were mounted on board the UWKA, including a cloud imaging probe (CIP; http://www.dropletmeasurement.com/cloud-imaging-probe-cip) from Droplet Measurement Technologies, Inc. (DMT). The CIP used here measures particles with diameters of 0.01–2.51 mm and sorts them into 101 bins with equal widths of 25 μm. In situ liquid water content (LWC) is estimated by several probes, including the DMT cloud droplet probe (CDP; Lance et al. 2010) which also estimates the size distribution of drops with diameters of 2–50 μm.

b. Airborne remote sensing instrumentation

Two remote sensing platforms were mounted on the UWKA during OWLeS: the Wyoming Cloud Radar (WCR) and Wyoming Cloud Lidar (WCL). A description of how these platforms have been integrated on board the aircraft is provided by Wang et al. (2012). The WCR is a W-band (3-mm wavelength) pulsed Doppler radar well suited for finescale examination of clouds but also capable of detecting various types of precipitation particles, including snow crystals. Three fixed antennas, with beams directed upward (zenith), downward (nadir), and down forward (about 30° forward of nadir), were utilized during OWLeS. Data were sampled from each beam quasi simultaneously, allowing for vertical cross sections of reflectivity and hydrometeor vertical velocity to be obtained. Sampling occurred about every 5 m along the flight track and at 15-m range increments from the antennas. The maximum unambiguous Doppler velocity was ±15.8 m s⁻¹. To determine hydrometeor vertical velocity, Doppler velocity measurements from the upward- and downward-pointing beams were first corrected for aircraft motion and further corrected for horizontal wind contamination that results from deviations in aircraft pitch and roll angles. For the latter correction, a vertical profile of the horizontal wind was obtained from a nearby surface sounding (see section 2d).

Furthermore, an estimate of the air vertical velocity was calculated by assuming a mean hydrometeor fall speed of 0.5 m s⁻¹ and removing it from the WCR hydrometeor vertical velocity measurements. This assumes that the radar scatterers (predominantly the larger ice particles) were sufficiently large to attain this fall speed, and not significantly rimed so they did not exceed this fall speed. Images from the optical array probes (not shown) show no evidence of riming at flight level (i.e., near cloud top). The most common particles appear to have been unrimed plates, irregular crystals, and small aggregates with a maximum long axis of ~0.8–1.0 mm, but more commonly up to 0.2 mm in size (see section 3b). Such crystals are unlikely to substantially aggregate at the observed very low cloud temperatures (~20° to ~28°C). Locatelli and Hobbs (1974) show that small (<3 mm) unrimed ice crystals of various habits generally fall at speeds closer to 0.5 than 1.0 m s⁻¹, while Mitchell (1996) also shows that 0.5 m s⁻¹ is an appropriate fall speed estimate for similar particles with diameters around 1 mm. In situations where larger slightly rimed or aggregated particles were actually present, or in very weak echoes (< ~20 dBZ) where all the particles were small (<0.2 mm), the error in the air vertical velocity estimate may be on the order of 0.3–0.5 m s⁻¹.
The WCL is a polarization backscatter lidar system that contains two separate upward- and downward-pointing lasers with wavelengths of 355 and 351 nm and range resolutions of approximately 3.75 and 1.5 m, respectively. This setup allows for various cloud and aerosol properties to be measured above and below the aircraft and displayed as cross sections in the same manner as the WCR data. The sensitive signal can quickly be attenuated when flying in or near clouds containing numerous droplets. Thus, analysis of WCL backscattered power cross sections are helpful in qualitatively distinguishing between liquid, ice, and mixed-phase clouds.

c. Dual-Doppler synthesis

An airborne dual-Doppler (DD) technique described by Damiani and Haimov (2006) was adopted for examination of the 2D wind field in the vertical plane below the UWKA. This technique has been used in many other studies in the past decade (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). To execute the technique, the along-beam Doppler velocities from the two downward-pointing WCR beams (separated by 30°) are interpolated either onto a straight 2D Cartesian grid, or a curtainlike 3D grid (this choice depends on how straight the flight track is) and then decomposed to find the contribution of the wind that lies within the grid plane. Given that the flight legs presented here were quasi straight and flown at a constant altitude for the majority of their durations, a straight 2D grid was chosen. The grid resolution chosen for our DD analysis was $30 \times 30$ m$^2$. This implies relatively little smoothing, given the WCR range resolution and beam spacing.

The down-forward beam first obtains a mean radial velocity from a volume of scatterers ahead of and below the aircraft. A short time later (~6 s per kilometer of range) the nadir beam passes overhead and ideally samples the same volume of scatterers. To account for the movement of the scatterers, an assumption is made about their advection speed, which is incorporated into the construction of the grid. For this study, it is assumed that the scatterers moved with the mean wind velocity measured at flight level by the gust probe. Such an assumption is appropriate since wind velocities within the boundary layer up to flight level (~1.1 km MSL) were nearly constant throughout the flight (see section 3a).

The implementation of alternative advection assumptions [discussed in Damiani and Haimov (2006)] did not significantly alter the DD results shown in this study. During the DD processing, the radial velocities from both beams are corrected for cross-track wind component to be more easily seen.

d. Additional data sources

Three Vaisala radiosondes were released by one of the OWLeS teams stationed in Stanley, New York (about 10 km west of the northern tip of SL; Fig. 1b), between 1441 and 1730 UTC. The sounding measurements were collected at 1 Hz and quality controlled by the National Center for Atmospheric Research (NCAR) Earth Observing Laboratory (EOL). We also...
used standard surface observations of temperature, dewpoint, and wind across western New York (available online at http://mesowest.utah.edu/).

Supplementary resources further include 0.5° base reflectivity data from the Binghamton, New York (KBGM), S-band (10.7-cm wavelength) Weather Surveillance Radar-1988 Doppler (WSR-88D) obtained from the National Climatic Data Center (NCDC) online data repository (www.ncdc.noaa.gov/nexradinv/), as well as high-resolution (250 m) false-color imagery from the Moderate Resolution Imaging Spectroradiometer (MODIS) on board the Terra polar-orbiting satellite (ge.ssec.wisc.edu/modis-today/index.php). A false-color image was chosen as it allows low-level clouds to be visually distinguished from underlying snow cover or lake ice. In addition, synthetic maps were generated with 12-km North American Mesoscale Forecast System (NAM) analysis data archived on the NCDC National Operational Model Archive and Distribution System (NOMADS) website (nomads.ncdc.noaa.gov/data.php). Topographic maps were created with 3-arc-second (~90 m) land elevation data from the Shuttle Radar Topography Mission (STRM; obtained online at dds.cr.usgs.gov/srtm/version2_1/SRTM3/).

3. Event description and airborne observations

a. 22 January 2014 synopsis

A two-day LE snow event occurred over the New York Finger Lakes region during 21–22 January 2014 when persistent boundary layer winds out of the north or northwest allowed for periods of organized banded LE convection to develop over both SL and CL. Inspection of the 1200 UTC synoptic conditions from the NAM (about two hours prior to takeoff of the UWKA) shows a negatively tilted shortwave trough at 500 hPa with an axis extending from southeastern Ontario down through New York and Pennsylvania. The trough was accompanied by a surface low pressure system off the mid-Atlantic coast (Fig. 2a), helping to usher in very cold low-level air from the north (Fig. 2b). Temperatures around −20°C at 850 hPa (Fig. 2b) and 1000–500 hPa thicknesses near 5040 m across western New York (Fig. 2c) highlight the frigidity of the air mass. This is also seen in the surface (2 m) temperatures, although a pocket of relatively warmer air is clearly seen immediately downwind of Lake Ontario (Fig. 2d) where the air was 4–8 K warmer than surrounding areas not downwind of the lake. This is indicative of boundary layer airmass modification associated with the warmer lake waters. Winds at the surface and 850 hPa were aligned out of the north or north-northwest (Figs. 2b,d). The conditions seen here are generally consistent with the typical synoptic setup of Finger Lakes LE snow events as described by Laird et al. (2010).

The UWKA flew during 1401–1751 UTC (0901–1251 EST) with the primary area of focus being over SL and CL. Figure 1b shows the topography of the Finger Lakes region along with the UWKA flight track and the flight legs examined in this study. The flight legs, oriented from southwest to northeast across and approximately normal to the underlying lakes, were completed in two sets of five. Legs 1 and 1.5 were each only flown once, in the first and second set of legs, respectively. The first set (i.e., legs 1–5) was flown from 1414 to 1508:30 UTC while the second set (i.e., legs 1.5–5) was flown from 1640 to 1737 UTC. All legs were 35–45 km in length (~7–8 min of flight time) and separated laterally by about 20 km, with the exception of legs 1.5 and 2, which were separated by about 10 km. Flight altitudes varied between 700 and 1100 m MSL. An effort was made to keep each leg straight and level, although intentional changes in flight level were made in the first set of legs during legs 1 and 2.

High-resolution false-color MODIS imagery from around 1530 UTC (Fig. 3a) reveals two long and narrow LE cloud bands over or near the open waters of SL and CL. The MODIS image shows that the northern tip of CL was covered by ice while SL was completely ice free. Smaller cloud bands were situated over several of the other Finger Lakes to the east as well. LE convection organized in wind-parallel bands was emanating southward from Lake Ontario, with remnants of these bands reaching as far south as SL. Despite the visual presence of cloud bands over the Finger Lakes, KBGM WSR-88D base reflectivity at 1436 UTC (during the first set of flight legs), 1534 UTC (near the time of the MODIS image), and 1701 UTC (during the second set of flight legs) only captured portions of the bands at the far southern ends of SL and CL (Figs. 3b–d). Farther north over the lakes, it appears that the shallow tops of the bands were being overshot even by the lowest elevation angle (0.5°) radar beam. In fact, the SL band does not appear to be visible at all in these images. A similar radar limitation prevented Laird et al. (2009b) from including the far western Finger Lakes in their climatology. The KBGM 0.5° beam heights at the southern ends of SL and CL were about 1370 and 1010 m AGL, respectively, and increased to about 2040 and 1910 m AGL at the northern ends. Photographs taken with the UWKA forward-pointing camera (Fig. 4) indicate that cumulus cloud tops within the bands over the northern portions of the lakes were typically about as high as flight level (i.e., ~1100 m MSL). A trio of soundings from 1441, 1554, and 1730 UTC, taken upwind of the Finger Lakes at Stanley, New York (cf. Fig. 1b), all
show a strongly capped shallow mixed layer about 1 km (100 hPa) deep (Fig. 5). This layer, which warmed by −2°–4°C and deepened slightly between the first and last soundings (presumably due to daytime heating), contained the LE bands emanating from Lake Ontario. Winds throughout the layer were out of the northwest at about 5 m s⁻¹ (10 kt) in each sounding, veering within the capping inversion, but gradually shifted from about 350° at 1441 UTC to 333° at 1554 UTC and then to 325° at 1730 UTC. This shift in winds has implications for the development of the LE bands over SL (Fig. 1b).

Vertical profiles of WCR reflectivity, vertical velocity, and WCL backscatter power from both sets of flight legs are presented here. These profiles, utilizing the two vertically pointing beams from both the WCR and WCL, allow for careful examination of cloud band structure and the spatial and temporal evolution of the bands during the flight. The 1554 UTC sounding shown in Fig. 5 was used for the correction of vertical velocities in both the vertical profiles presented in this section and in the DD data presented in section 3c. These corrections were briefly explained in sections 2b and 2c. We will examine each set of legs separately.

**b. Cross-band WCR and WCL vertical profiles**

Vertical profiles of WCR reflectivity, vertical velocity, and WCL backscatter power from both sets of flight legs are presented here. These profiles, utilizing the two vertically pointing beams from both the WCR and WCL, allow for careful examination of cloud band structure and the spatial and temporal evolution of the bands during the flight. The 1554 UTC sounding shown in Fig. 5 was used for the correction of vertical velocities in both the vertical profiles presented in this section and in the DD data presented in section 3c. These corrections were briefly explained in sections 2b and 2c. We will examine each set of legs separately.
1) 1414–1508 UTC FLIGHT LEGS

WCR reflectivity from the first set of legs is shown in Fig. 6. Low-level winds across the Finger Lakes were more or less northerly at this time (cf. Figs. 3b–d and 5), nearly parallel to the long axis of SL and CL. During leg 1 (Fig. 6a), the LE band over SL (at $x = 6–8$ km) was in its early stages of formation and was, therefore, very weak and shallow, only about 200 m deep. A photograph looking northeast from on board the UWKA at ~1414 UTC (Fig. 4a) shows this developing convection (the UWKA crossed the band on the left side of the image, where cloud tops were lower). The weaker radar echoes just east of SL may have been associated with dissipating ice clouds that were part of the residual LE convection emanating from Lake Ontario. These particular clouds seem to have been somewhat transient and perhaps had thermal and reflective characteristics that were too similar to the underlying snow cover, similar to blowing snow (e.g., Palm et al. 2011), and are thus not easily seen in MODIS imagery (cf. Fig. 3a, about an hour later). Nonetheless, the residual ice particles from these clouds or others like them were likely key to the rapid natural seeding of the developing convection over SL, as will be shown later. No precipitation was observed over the northern end of CL at this time, where there was substantial ice cover (Fig. 3a).

The SL band had grown rapidly by leg 2, with maximum reflectivity approaching 5 dBZ and echo tops extending 500–600 m above the ground below (Fig. 6b). At this point the band was only about 4 km in width. The fact that the band was not situated directly over the lake here was merely due to the fact that the lake features a slight bend to the east where this leg was flown (cf. Fig. 1b), allowing the snowband to briefly move over the lake’s western shore. It is along this leg that CL-based convection is first evident in WCR data (i.e., at $x = 21$ km). This convection extended over most of the depth of the mixed layer, but the reflectivity was...
quite low, lower than along leg 1 over SL, indicating few/small ice crystals only. The entire width of CL is visible in the UWKA photograph taken from the east (looking west), indicating that precipitation over the lake was very light (Fig. 4b). Also note the presence of numerous steam devils above the water surface, indicative of rapid vertical transport of heat and moisture within the unstable air over the water. By leg 3, the SL
band had grown to about 800 m in depth and reflectivity
nearly 10 dBZ within a well-defined core (Fig. 6c). The
band width, however, remained at about 4–5 km. It had
also shifted to the eastern shore primarily due to the
slightly curved shape of the lake beneath (Fig. 1b). The
CL band was much more apparent along this leg, albeit
still quite a bit weaker than the SL band. Irregularities
in the shape of CL were also responsible for the ap-
parent westward displacement of the band from the
lake surface.

Both snowbands persisted onshore south of the lakes
while maintaining some of their structure, as shown
along leg 4 (Fig. 6d). The SL band was briefly inter-
cepted at flight level. By leg 5, only the band coming off
SL remained intact, near x = 8 km (Fig. 6e). The weak
(i.e., less than −10 dBZ) radar echoes farther east rep-
resent light orographic snow showers over the higher
terrain. It is worth noting that the orientations of the
drainage valleys leading into the two lakes from the
south are quite different. The fact that the valley leading
into SL is longer and oriented from north to south while the one leading into CL has several sharp bends (cf. Fig. 1b) may help explain why the SL circulation (and thus snowband) persisted as far south as it did while remaining relatively unobstructed by the terrain below.

The same type of five-panel plot is also shown for WCR vertical velocity in Fig. 7. As mentioned earlier, a hydrometeor fall speed estimate of $0.5 \text{ m s}^{-1}$ has been subtracted from the vertical velocity measurements to retrieve an indirect estimate of the air vertical motion. In these plots and throughout the rest of the vertical velocity plots in this paper, warmer colors represent upward motion while cooler colors represent sinking motion. The vertical motions over the lakes were characteristic of convection driven by surface fluxes, with multiple thin updrafts and downdrafts situated adjacent to each other (Figs. 7a–c). The strongest updrafts (of about 2–3 m s$^{-1}$) were seen in the developing CL band along leg 2 ($x = 22$ km; Fig. 7b) and the SL band along leg 3 ($x = 5$ km; Fig. 7c) and leg 4 ($x = 5$ km; Fig. 7d). These updrafts were all very narrow, $<1$ km wide. Some downward compensation to the sides of the updrafts was also observed, especially in the SL band along leg 3. The vertical velocity profiles quickly became dominated by terrain-driven

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**Fig. 7.** As in Fig. 6, but for estimated air vertical velocity $w$. Warm colors signify positive vertical velocity and thus upward air motion. The air vertical velocity measurements from the UWKA gust probe are plotted at flight level for comparison with the WCR data below the aircraft.
motions downwind of the lakes along leg 4 and especially leg 5 (Figs. 7d,e), making it difficult to identify coherent vertical circulation structure within the bands.

A qualitative analysis of the WCL backscatter power profiles along each of these legs indirectly suggests that liquid water droplets were in abundance within the CL band along legs 2–4 owing to the almost immediate extinction of the lidar beam near cloud top (Figs. 8b,c). Beam attenuation is characterized by a loss of returned signal at increasing range from flight level, with total extinction typically indicating the presence of a significant amount of supercooled liquid water. Conversely, the scattering of the lidar beam within the SL band was largely due to ice particles (which are far less numerous than droplets), despite the band being deeper than the one over CL. This can be inferred from the fact that the lidar beam suffered no rapid extinction within the SL cloud, but rather gradual attenuation. Furthermore, the lidar depolarization ratio (not shown) within the attenuating layer over SL was high, which is an indication of scattering by nonspherical particles. The stark difference in cloud liquid water between the two bands suggests that glaciation was occurring much more efficiently in the SL band. Even the most recently developed SL clouds (leg 1, Fig. 8a) were largely glaciated. This appears to be due to natural seeding by ice cloud remnants.

**Fig. 8.** As in Fig. 6, but for WCL backscatter power. A 100-m-thick lidar blind zone surrounds the aircraft flight level.
from the Lake Ontario convection, which would have allowed for more efficient depositional growth and faster consumption of liquid water. At the observed low temperatures (and thus large saturation vapor pressure difference between a water surface and an ice surface), droplets should be consumed rapidly by nearby ice crystals, through the Bergeron process. Why these ice clouds reached SL, but not CL (cf. Fig. 3a), is not clear.

Given the orientation of the flight leg pattern with respect to the lakes (cf. Fig. 1b), it is evident from these vertical profiles that both bands tended to follow the geometry of the lakes below, but their strength was limited by the fetch over the lakes. Early during the flight, the SL band developed quickly and was stronger and well aligned with the northerly low-level flow, which happened to have a large fetch over SL. The CL band, on the other hand, was slower to develop and was quite a bit weaker as the fetch over CL at this time was smaller. As the wind gradually backed during the flight, its orientation became more favorable for LE snow development over CL.

2) 1640–1737 UTC FLIGHT LEGS

The second set of flight legs began about 90 min after the conclusion of the first set and, with the exception of the northernmost leg, followed the same flight track. Once again, reflectivity profiles for each of the legs are given in Fig. 9. During this period the SL band was deeper and stronger over the northern portions of the lake (legs 1.5 and 2) while remaining well defined farther south. A photograph taken earlier (at 1617:30 UTC) from over the northern shore of SL (looking toward the southeast) shows how quickly the band had developed and deepened to the south (Fig. 4c). An eastward shift in band position relative to the lake had also occurred, likely in response to the northwesterly shift in low-level wind direction across the region (section 3a). The wind shift not only caused the band to migrate toward the eastern shore of SL, but was also responsible for the increase in radar echoes over the higher terrain between SL and CL owing to the eastward advection of moistened air and snow particles. Furthermore, the wind shift increased the fetch over the southern portion of CL, allowing the CL band to exhibit better structure along leg 2 and to deepen and strengthen significantly by leg 3 (Figs. 9b,c). However, precipitation over CL along leg 2 appears to have remained very light, as indicated by the good visibility below cloud base as seen from the UWKA (Fig. 4d).

Vertical velocities were also more impressive during this flight period (Fig. 10). Most apparent is the large updraft (1.0–1.5 km wide) along leg 2 in the SL band, with a maximum velocity of nearly 4 m s⁻¹ (x = 5–6 km; Fig. 10b). Strong (3–4 m s⁻¹) updrafts were also seen within the bands along leg 1.5 (x = 9 and 25 km; Fig. 10a), elsewhere along leg 2 (x = 21.5 and 23 km; Fig. 10b), and leg 3 (x = 31 km; Fig. 10c). These updrafts weakened as the bands moved onshore downwind of the lakes, where smaller terrain-driven updrafts again began to dominate (Figs. 10d,e).

Backscatter power profiles from the WCL indicate continued natural seeding of developing clouds over SL, and the continued lack thereof over CL. However, the developing convective clouds over SL (along legs 1.5 and 2) appeared to contain more liquid water than during the first set of flight legs (Figs. 11a,b). This is evident from the rapid extinction of the WCL power in the first few gates, and the LWC (as large as 0.2 g m⁻³, not shown) at flight level in these clouds. The reason why leg 1.5 was flown in the second sequence, rather than leg 1 (Fig. 1b), is because leg 1.5 was just far enough south over CL to intercept the first convective cloud. Clearly the tops of this cloud are liquid (Fig. 11a).

The advection of ice particles to the east of SL from near cloud top is apparent in Fig. 11 (x = 12–20 km along leg 1.5; x = 7–14 km along leg 2), as revealed by the weak attenuation. Along legs 3–5, the SL band had moved off of the lake—away from its source of liquid water—and was composed mostly of ice (Figs. 11c–e). The band over CL, which was more aligned with the wind at this time, continued to be composed primarily of liquid water along the first four legs.

Flight level particle size distributions from the CIP and mean LWC from the CDP are shown in Fig. 12 for several sections in which the UWKA flew through the top of either the SL (four times) and CL (once) band during this set of legs. Only individual LWC measurements of greater than 0.005 g m⁻³ were included in the calculation of the mean, which naturally ignores portions of cloud where only ice particles were present. The actual penetrations were very short, about a minute or less in duration, although the size distributions are from slightly longer periods of time (1–3 min) surrounding the penetrations. While the largest ice particles were on the order of about 1 mm in diameter along each of the SL legs, the total concentration of particles (especially small particles) and mean LWC gradually decreased to the south. This supports the notion that additional ice from LE convection over Lake Ontario was seeding the SL band during its early stages of development. This limited the amount of liquid water even in new cumulus clouds over northern SL (<0.1 g m⁻³) and rapidly depleted liquid water to the south. Upstream measurements from the CIP (not shown), collected during the north-south flight legs between Lake Ontario and SL, indicate that very small ice was indeed present in the clear air between the upstream clouds and the band over SL during.
one of the legs (i.e., when the photo in Fig. 4c was taken), but not the other. There were very few ice particles (on the order of $1 \text{ L}^{-1}$ for the smallest particles as shown by the blue line in Fig. 12) and greater mean LWC in the CL band along leg 3, consistent with the WCL data suggesting that this band was composed mostly of liquid. However, this penetration was exceptionally short, lasted only for $\sim 15 \text{s}$, and the lack of CL band penetrations along other legs prevents further assessment of the downwind microphysical evolution of the band. The brevity of the band penetrations and the fact that the UWKA only sampled the tops of the bands does indeed limit the interpretation of band evolution to cloud top only. It is possible that the size distributions and LWC values shown here are not fully representative of the lower portions of the LE bands.

c. WCR vertical dual-Doppler analysis

The presence of lake-scale surface convergence zones within single LE snowbands aligned with the long axis of larger lakes has been well established by modeling and observational work. These convergence zones are hypothesized to be associated with thermally driven secondary circulations (Passarelli and Braham 1981;
Hjelmfelt and Braham 1983; Laird et al. 2003), although few observations of these circulations can be found in the literature (e.g., Steiger et al. 2013). We are curious if similar secondary circulations occur within LE bands over smaller lakes such as the Finger Lakes. Information regarding the along-track 2D wind field below the UWKA around and within the snowbands presented in this study can be gleaned through WCR DD synthesis. Here, we take a close look at a well-defined section of each band from the second set of legs in which a strong updraft was present.

We first examine a short section of leg 2 across the SL band, shown by the box in Figs. 9b and 10b. Figure 13 presents a zoomed-in view of WCR reflectivity and air vertical velocity from within this box, and the corresponding along-track perturbations $u'$ within the 2D DD wind field below the UWKA. Here $u'$ is relative to the mean along-track wind throughout the entire depth of the band ($\bar{u}$). Within this section, $\bar{u} = 1.2 \text{ m s}^{-1}$ (positive if from the left), and it has been removed only to highlight the circulation cells. Cool and warm colors indicate the strength of the wind anomaly. In these plots, positive perturbations always point to the right (i.e., northeast). Thus, a left–right transition from orange to blue (blue to orange) would be indicative of convergent (divergent) flow. As with Fig. 7, a $0.5 \text{ m s}^{-1}$ hydrometeor fall speed estimate has been removed from both the vertical velocity and DD data to obtain estimates for the vertical velocity.

**Fig. 10.** As in Fig. 9, but for estimated air vertical velocity $w$. 
air motion $w$. Perturbation vectors ($u', w$) are also plotted to show the full 2D flow.

In this first section, the DD data suggest that a secondary circulation does indeed appear to have been present within the SL band (Fig. 13c). A shallow layer, 200–300 m deep, near the surface included a convergence zone (at $x = 4$ km) collocated with the surface position of the band’s updraft. Near the top of the updraft, divergent flow was present in both the DD and flight level data. The secondary circulation—which actually comprised two separate circulations on either side of the updraft—is perhaps illustrated even more intuitively by the vector field, which shows the air converging into the strong updraft at the surface and diverging aloft. Some sinking motion is also seen on either side of the updraft in the vertical velocity data (Fig. 13b). The full circulation, with the updraft in the center, appears to have only been 3–4 km wide and up to 1000 m deep. This circulation was also seen within the SL band along leg 1 but not along legs 3–5 (not shown). In other words, it vanished quickly as the band was advected away from the lake surface. It was also observed in the earlier set of flight legs (over SL along legs 3 and 4) when the band was shallower.

A similar circulation was seen in the CL band, despite the fact that the band was not nearly as strong, nor as deep, as the band over SL. One of the strongest sections of the band (shown by the red box in Figs. 9c and 10c),

![Figure 11](image-url)
sampled by the WCR along leg 3, is shown in Fig. 14. In this section, a $u'$ of $-3.1 \text{ m s}^{-1}$ was removed to obtain $u_0$.

Here, the band structure was quite symmetric with the top of the band exhibiting an anvil-like appearance, implying divergent flow. There were elevated areas of relatively higher reflectivity (>0 dBZ) within the band core and a prominent central updraft of 3–4 m s$^{-1}$ (at $x = 5 \text{ km}$) flanked by compensating downward motion on either side (Figs. 14a,b). The same general along-track perturbation pattern seen in the SL band shows up here as well, with a shallow layer of surface convergence once again collocated with the updraft, near the western shoreline of the lake, and divergence near the top of the band (Fig. 14c). By leg 4, the circulation had vanished (not shown) as the band moved downwind of the lake over higher terrain where orographic effects began to dominate the flow.

Although lake-induced buoyancy presumably drives these circulations, shallow drainage flow forced by the higher surrounding terrain may also contribute to the development of the surface convergence zone within these small bands (Laird et al. 2009b). There may have been some weak drainage flow contributing to the low-level easterly inflow descending from the slightly higher terrain to the east of the SL band in Fig. 13c. However, the dynamics of the secondary circulation, including cloud buoyancy and any contribution from drainage flow, cannot be investigated given the lack of adequate in situ observations, especially at low levels within the convergence layer.

4. Discussion

This paper presents a unique set of airborne radar and lidar observations of a New York Finger Lakes LE snow event. The observations are unique in the sense that airborne measurements of any kind within LE cloud bands have generally been obtained from bands over the much larger Great Lakes, primarily Lake Michigan (e.g., Passarelli and Braham 1981; Kelly 1982,1984; Braham 1990; Chang and Braham 1991; Braham et al. 1992; Braham and Dungey 1995; Kristovich et al. 2003; Schroeder et al. 2006; Yang and Geerts 2006; Barthold and Kristovich 2011). Airborne remote sensing measurements from within LE bands have been far less common (e.g., Kristovich et al. 2003; Schroeder et al. 2006; Yang and Geerts 2006; Welsh et al. 2016).

Our results provide further insight into the types of banded LE convection that lakes at these smaller scales are capable of producing. Specifically, they demonstrate that, even over very narrow lakes like the Finger Lakes (width $\approx 5 \text{ km}$), LE convection can organize into wind-parallel bands about as wide as the lake, with coherent vertical structures, secondary circulations (surface convergence and upper-level divergence), and updrafts of sufficient depth and strength to produce light snowfall. Although much smaller in overall size and depth, the bands observed here were similar to the “type-I” bands defined by Niziol et al. (1995). Alternatively referred to as long-lake-axis-parallel (LLAP) bands (Steiger et al. 2013; Veals and Steenburgh 2015; Welsh et al. 2016), type-I bands are most commonly seen over larger lakes (e.g., Lake Ontario, whose width is $\approx 70 \text{ km}$) when the prevailing low-level winds are aligned with the long axis of the lake, maximizing the fetch and, therefore, the heat and moisture flux into the boundary layer. A thermally induced convergence zone develops near the surface and the convection becomes organized linearly into a single, intense snowband with a dominant central updraft.

The fact that bands of similar structure and behavior are able to form over the far narrower Finger Lakes, with significantly less surface area and shorter fetches (and consequently less available heat and moisture), supports our suspicion that upstream airmass modification from Lake Ontario is often an important factor in helping to destabilize and moisten the boundary layer prior to LE development over the Finger Lakes. However, the lack of soundings from upstream of Lake Ontario during this event makes it difficult to know the extent to which the air mass was modified.

This study also provides evidence suggesting that Lake Ontario influenced the Finger Lakes bands in another way. Namely, the addition of ice particles into the
upstream boundary layer from residual transient LE snowbands coming off Lake Ontario likely seeded developing convective clouds over SL. This led to rapid glaciation and snowfall, within a few kilometers of the first convective towers emerging over SL.

5. Summary

A pair of LE snowbands that formed over the two largest New York Finger Lakes were examined with an airborne Doppler radar and lidar on board the UWKA. These observations provide a unique perspective on LE snowbands that form when arctic air is advected along the length of small, elongated lakes (width ~5 km). Two sets of UWKA legs were flown across the lakes from southwest to northeast, perpendicular to the snowbands. The bands were easily identified in high-resolution satellite imagery during the UWKA flight but were too shallow over the lakes to be detected by WSR-88D low-elevation scans. Vertical profiles of reflectivity and Doppler velocity across these bands indicate that they were about as wide as the lake, and generally no more than 1 km deep. At times, their primary updrafts approached or exceeded 4 m s\(^{-1}\) but tended to weaken when the bands moved over land away from their source of buoyancy. The SL band was the stronger of the two throughout the flight as the prevailing low-level flow (northerly) tended to have a greater fetch over SL than over CL, although a slight shift in the flow direction to northwesterly later in the flight allowed the CL band to strengthen during the second set of flight legs.

Secondary circulations were observed within both bands at different times by the WCR DD data. These circulations were presumably thermally driven and were best defined when the band was over the lake and contained a strong updraft. The general appearance of the bands suggests that they may have been dynamically similar to type-I (i.e., “LLAP”) LE bands that typically form over the Great Lakes, although the source lakes were at least an order of magnitude narrower. Lake Ontario likely influenced the development of these Finger Lakes bands by modifying the upstream boundary layer through the addition of heat and moisture. This study provides evidence that residual, fully glaciated LE snowbands advected from Lake Ontario seeded incipient convective clouds over at least one of the Finger Lakes.
A future high-resolution modeling study would complement this analysis well, given the number of interesting observations encountered in this study. Among other things, modeling work would be able to investigate the possibility of natural seeding by upwind LE convection and the connection between boundary layer modification over Lake Ontario and LE snowband formation over the Finger Lakes. Additionally, a closer look at the dynamics of the observed secondary circulations within these small bands and the contribution of drainage flow from the surrounding hillsides to the low-level convergence would be of interest.

Acknowledgments. This study would not have been possible without the enthusiastic participation of all those involved with the OWLeS campaign. Special thanks go out to the UWKA flight crew, who put in long hours to make these flights possible, and to the folks from Millersville University for launching the soundings used in this study. This paper benefitted from discussions with and input from Neil Laird, Nicholas Metz, and Jeff French, as well as from the suggestions and careful critique offered by Daniel Kirshbaum and three anonymous reviewers. We also thank Binod Pokharel for his assistance with data analysis, and Rick Damiani and Sam Haimov for allowing us to utilize their dual-Doppler code. We acknowledge the use of MODIS imagery from the Land, Atmosphere Near real-time Capability for EOS (LANCE) system operated by the NASA GSFC/Earth Science Data and Information System (ESDIS) with funding provided by NASA headquarters. Funding for this work came from NSF Grant AGS-1258856 and NASA Grant NNX15AI08H.

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