Boundary Layer and Microphysical Influences of Natural Cloud Seeding on a Lake-Effect Snowstorm

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ABSTRACT

The first detailed observations of the interaction of a synoptic cyclone with a lake-effect convective boundary layer (CBL) were obtained on 5 December 1997 during the Lake-Induced Convection Experiment. Lake-effect precipitation and CBL growth rates were enhanced by natural seeding by snow from higher-level clouds and the modified thermodynamic structure of the air over Lake Michigan due to the cyclone. In situ aircraft observations, project and operational rawinsondes, airborne radar, and operational Weather Surveillance Radar-1988 Doppler data were utilized to document the CBL and precipitation structure for comparison with past nonenhanced lake-effect events. Despite modest surface heat fluxes of 100–200 W m$^{-2}$, cross-lake CBL growth was greatly accelerated as the convection merged with an overlying reduced-stability layer. Over midlake areas, CBL growth rates averaged more than twice those previously reported for lake-effect and oceanic cold-air outbreak situations. Regions of the lake-effect CBL cloud deck were seeded by precipitation from higher-level clouds over the upwind (western) portions of Lake Michigan before the CBL merged with the overlying reduced-stability layer. In situ aircraft observations suggest that in seeded regions, the CBL was deeper than in nonseeded regions. In addition, average water-equivalent precipitation rates for all of the passes with seeded regions were more than an order of magnitude greater in seeded regions than nonseeded regions because of higher concentration of snow particles of all sizes. A maximum snowfall rate of 4.28 mm day$^{-1}$ was calculated using aircraft particle observations in seeded regions, comparable to snowfall rates previously reported for lake-effect events, often with much larger surface heat fluxes, but not interacting with synoptic cyclones.

1. Introduction

Lake-effect snowstorms have important societal and economic consequences for those living near the Laurentian Great Lakes, especially within about 50 km of their eastern and southern shores (e.g., Schmidlin 1993; Schmidlin and Kosarik 1999; Kunkel et al. 2002). In some of these regions, annual winter precipitation is nearly double that of areas not substantially affected by lake-effect precipitation (Scott and Huff 1996). Predictions of how these snowstorms will affect specific communities are notoriously difficult because of the large spatial and temporal variations they exhibit (Niziol 1987). While our understanding of the mesoscale characteristics of lake-effect storms has steadily improved, much less has been reported on the common enhancements of snow fields due to interactions between lake-effect snowstorms and synoptic cyclones (e.g., Niziol et al. 1995). For this study, we will define these as synop-
tically enhanced lake-effect snowstorms (as opposed to “classic” lake-effect storms not interacting with such cyclones).

Environmental conditions common to synoptic cyclones may be favorable for lake enhancements. For example, Niziol et al. (1995) described cases where lake-enhanced snowfall was generated by midtropospheric cyclonic vorticity advection and surface convergence along a trough. In addition to influences of mesoscale upward motions, it would be anticipated that environmental conditions often associated with cyclones could also lead to increased snowfall. Factors that might be anticipated to enhance lake-effect processes, particularly in the cold air north and west of cyclones, include increased wind speeds (leading to increased surface fluxes), decreased low-level stability upwind of the lake, and increased upwind relative humidity (Kristovich and Laird 1998). In addition, precipitation from higher-level clouds could fall into lower lake-effect clouds (natural cloud seeding), a process commonly observed in situations with multiple cloud layers. This study investigates the development of a lake-effect convective boundary layer (CBL) during a synoptically enhanced lake-effect snowstorm. Comparisons are made with previous lake-effect systems to gain insight into the enhancement processes.

A number of studies have detailed the cross-lake growth of the lake-effect CBL. For example, Lenschow (1973) was among the first to use aircraft measurements to describe boundary layer airmass modification over the Great Lakes. More recently, Chang and Braham (1991) documented the rapid growth of a CBL using airborne in situ observations, and Kristovich et al. (2003) used field observations to detail the coevolution of the CBL and mesoscale convective structure in a lake-effect event over Lake Michigan. Several authors have shown large increases in boundary layer growth rate when a low-stability layer was present above the lake-effect CBL (e.g., Lenschow 1973, 13 November 1970 case; Agee and Gilbert 1989). Chang and Braham (1991) found that CBL growth rate increased rapidly at one location over Lake Michigan as a result of the lake-effect convection penetrating a layer of nearly neutral static stability or latent heat release in the cloud and snow that formed over the lake.

Another potentially important factor in the development of heavy snowfall in lake-enhanced events is through modification of the lake-effect clouds and snow due to seeding from higher-level cloud layers. Studies of several non-lake-effect cases showed precipitation rates may be greatly increased by the “seeder–feeder” process (e.g., Cunningham 1951; Herzegh and Hobbs 1981; Locatelli et al. 1983; Waldstreicher 2002), but this process has not been quantified in a lake-enhanced event. It is difficult to compare the microphysical characteristics of the present case with nonsynoptically enhanced lake-effect properties since little detailed research on microphysical processes in lake-effect systems has been published. Braham (1990) conducted perhaps the most comprehensive study of snow spectra from lake-effect events using in situ particle probe measurements over Lake Michigan. Snow-particle-size spectra were found to be well approximated by a negative exponential function,

\[ N(D) = N_0 e^{-\lambda D}, \]  

where \( N(D) \) is number of crystals per unit volume with diameters between \( D \) and \( D + dD \) (Rogers and Yau 1989). Braham (1990) found ice-water contents in lake-effect storms from 0.003 to 0.321 g m\(^{-3}\). Braham and Dugney (1995), Braham et al. (1992), and Kristovich and Braham (1998) used similar methods to investigate spatial and vertical distributions of precipitation intensity in lake-effect events. A method similar to Braham (1990) will be employed in the current study to determine differences in spectra and snow intensity between seeded and nonseeded regions in a synoptically enhanced lake-effect snow event.

The present study seeks to detail the lake-enhanced CBL thermal and microphysical structure and evolution over Lake Michigan on 5 December 1997, using observations taken during the Lake-Induced Convection Experiment (Lake-ICE; Kristovich et al. 2000). Section 2 describes observation platforms available during Lake-ICE and methodologies employed in this study. Section 3 details evidence of natural cloud seeding of the CBL from above and documents the cross-lake CBL growth. The local microphysical and CBL changes in seeded regions of the CBL are given in section 4. Discussion and conclusions are given in sections 5 and 6.

2. Data and methodology

The goals of the current study are to investigate 1) the cross-lake CBL growth, 2) local variations in CBL depth associated with natural cloud seeding, and 3) local microphysical changes associated with seeding that occurred over Lake Michigan on 5 December 1997. Datasets were analyzed using methods described below.

a. Lake-ICE and operational data

Figure 1 gives aircraft flight locations on 5 December 1997 and sites of other facilities during Lake-ICE (Kris-
tovich et al. 2000). The University of Wyoming King Air (hereafter KA) flights were conducted in a series of crosswind stacks over central Lake Michigan. These stacks were performed to obtain observations of the upper portion of the lake-effect CBL. King Air data from approximately straight, level flight legs in each flight stack were used for several aspects of this investigation. Flight legs were generally 25–50 km in length and flown at three to five altitudes (Table 1). The National Center for Atmospheric Research (NCAR) Electra repeatedly traversed the short axis of Lake Michigan, nearly perpendicular to the KA’s stacks. These traverses were performed to collect observations by the Electra dual-Doppler Radar (ELDORA) system and examine the evolution of the CBL across the lake. Atmospheric soundings were collected by both aircraft near the upwind shore of Lake Michigan. Both aircraft flew from approximately 1630 to 1930 UTC.

A wide range of data was available to document atmospheric processes over and near Lake Michigan on 5 December 1997. On the KA, temperature, dewpoint temperature, and pressure measurements were obtained by the Minco Element Reverse Flow thermometer, the Cambridge Model 137C3 dewpoint probe, and the Rosemount 1501 pressure sensor, respectively (http://flights.uwyo.edu/). Water vapor mixing ratio, potential temperature, and equivalent potential temperature, important for analyses of CBL evolution, were computed from these measurements. Flight audio and video recordings from the KA provided qualitative information. The NCAR Electra measured similar parameters to those obtained by the KA (NCAR 2002) and also carried the ELDORA system (Wakimoto et al. 1996; Hildebrand et al. 1996).

Since natural cloud seeding by snow falling from aloft was hypothesized to have been a major component of the interaction between the lake-induced CBL and a synoptic cyclone on 5 December 1997, analysis of microphysical observations was critical. The KA’s complement of instrumentation for measuring cloud and precipitation characteristics included three Particle Measuring Systems (PMS) probes: a Forward-Scattering Spectrometer Probe (FSSP-100), a two-dimensional optical array cloud probe (2D-C, not operational on 5 December 1997), and a two-dimensional precipitation probe (2D-P). The FSSP measured particle sizes and concentrations in the range of 2.0–47.0 μm at a resolution of 3.0 μm. Coincidence and dead-time corrections to the raw data, described by Bren- guer (1989), were applied by the University of Wyoming. FSSP measurements were thought to be dominated by cloud droplets in the environment under

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**Table 1.** Beginning and ending times, approximate flight distances, and altitudes of the KA flight legs on 5 Dec 1997.

<table>
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<th>Flight leg</th>
<th>Begin time (UTC)</th>
<th>End time (UTC)</th>
<th>Flight distance (km)</th>
<th>Mean alt (radar altimeter; m)</th>
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**Fig. 1.** Locations of Lake-ICE facilities on 5 Dec 1997. Solid lines over Lake Michigan, labeled with two-letter identifiers, indicate approximate locations of flight stacks flown by the KA. The dashed line over Lake Michigan indicates the location of the NCAR Electra cross-lake flight legs. The ISS sites are labeled ISS1–3. The location of the Green Bay, WI, WSR-88D is labeled KGRB.
consideration, where ambient temperatures were generally 0° to –10°C. The 2D-P primarily detected 200–8000-μm (diameter) snow particles and were processed at NCAR to remove spurious particles (Baumgardner 1989).

Rawinsonde data obtained at three Integrated Sounding System (ISS) sites (Fig. 1) were useful in examining cross-lake CBL growth. During the times of aircraft operations, rawinsonde launches were made every 3 h from each site, but launches were less frequent outside this period on 5 December 1997. No automated corrections were performed on the data, but they were visually quality controlled. Additional sounding data from the National Weather Service (NWS) rawinsonde site at Green Bay, Wisconsin (GRB), were also used.

To gain information on precipitation fields over and near Lake Michigan, level II data from the KGRB Weather Surveillance Radar-1988 Doppler (WSR-88D) were examined using WSR-88D Algorithm Testing and Display System (WATADS) software (NSSL 2000). During the times of interest to this study, this radar was operated in volume coverage pattern 32, as recorded in the level II dataset. At the locations of the westernmost flight stacks conducted by the KA, the center of the radar beam at 0.5°-elevation angle (the lowest angle utilized) was approximately 1030 to 1040 m above the surface. Note that in situ observations taken by the KA (discussed in section 3b) indicate that the boundary layer depth was 850 to 900 m above the surface, suggesting that the radar beam was primarily above the lake-effect clouds and precipitation. WATADS software was also used to examine data collected by ELDORA. Lake Michigan surface temperature images, derived from infrared satellite measurements, were obtained from the Great Lakes Environmental Research Laboratory’s Internet site (GLERL 2000).

b. Microphysical characteristics

Ice-particle size spectra were observed using the 2D-P probe on the KA. The 2D-P-observed snow spectra were fit by a log-linear size distribution using the least squares method, giving values of $N_0$ and $\lambda$ in Eq. (1). There are a number of methods described in the literature for calculating ice-water content (IWC) and precipitation rates from particle spectra, which can yield different values than reported here. However, in order to allow for a direct comparison with previous studies of lake-effect precipitation (i.e., Braham 1990; Braham et al. 1992; Braham and Dungey 1995), we followed methods employed by these studies. IWC was calculated assuming spherical snow particles with a bulk density of 0.025 g cm$^{-3}$:

$$\text{IWC} = \pi N_0/40\lambda.$$  

For completeness, IWC was also calculated by

$$\text{IWC} = \sum N(D) M(D),$$

where $N(D)$ is the number concentration of particles with mean diameter $D$ and $M(D)$ is the mass of such particles. Water-equivalent precipitation rates were derived similarly to Braham et al. (1992) by calculating snow flux (SF),

$$\text{SF} = \sum N(D) M(D) V(D),$$

where $V(D)$ is the fall speed of particles of mean diameter $D$. Dividing SF by the density of liquid water yields water-equivalent precipitation rate. Relationships for $M(D)$ and $V(D)$ from Locatelli and Hobbs (1974) for “aggregates of densely rimed radiating assemblages of dendrites or dendrites” were used:

$$M(D) = 0.037D^{1.9}$$

and

$$V(D) = 0.79D^{0.27},$$

with units of $V$ in m s$^{-1}$, $M$ in mg, and $D$ in mm.

3. Synoptic conditions and boundary layer growth

a. Synoptic environment

The synoptic environment for this event was outlined by Laird et al. (2001). In summary, as a surface low pressure system moved from northern Lower Michigan to southern Ontario, Canada, during 4–5 December 1997, weak cold advection overspread the Lake Superior–Lake Michigan region (Fig. 2a). Temperature differences between the lake surface (GLERL 2000) and 850 hPa were in the range of 14°–19°C, only slightly in excess of the 13°C criterion typically thought to be required for initiation of significant lake-effect snowstorms in the Great Lakes region (e.g., Niziol et al. 1995). Laird et al. (2001) estimated, using bulk transfer methods, that total lake-surface heat fluxes (latent and sensible combined) were only 100–200 W m$^{-2}$ on this day, on the low end of similar observations taken during previously studied lake-effect snowstorms (e.g., Chang and Braham 1991; Kristovich and Laird 1998; Kristovich et al. 1999; Sousounis and Fritsch 1994). Despite only marginal conditions for lake-effect snows, interaction between the lake-induced CBL and the syn-
optic cyclone produced some of the heaviest lake-effect snowfall observed during Lake-ICE. Figure 3 shows that from 4 to 6 December 1997, 10–15 cm of snow fell along the downwind shore of Lake Michigan, with 5+ cm amounts over nearly half of Lower Michigan.

Above the surface, several shortwave troughs and vorticity lobes circled around the western edge of the departing low pressure system, over the Lake Superior–Lake Michigan region (Laird et al. 2001). Widespread high-level clouds were observed throughout the region, obscuring the lower lake-effect boundary layer clouds from observation by satellite (Fig. 2b). Surface and radar observations indicate intermittent light snowfall across the upper Midwest and extreme south-central Canada on 5 December 1997, even in areas well removed from lake-effect processes.

b. Cross-lake boundary layer growth

While the synoptic environment gives rise to conditions appropriate for lake-effect snowstorms, processes within the CBL are directly responsible for lake-effect snow development (e.g., Chang and Braham 1991; Kristovich et al. 2003). Sounding data from project aircraft and rawinsondes were used to detail the cross-lake growth of the lake-effect CBL. The vertical cross section of equivalent potential temperature \( \theta_e \) derived from these datasets, is summarized in Fig. 4. For purposes of this study, the base of the lowest \( \theta_e \) inversion layer was taken as the approximate top of the CBL, consistent with Chang and Braham (1991) and Kristovich et al. (1999, 2003).

Near the upwind (Wisconsin) shore, the CBL was deeper than has often been reported upwind of Lake Michigan in classic lake-effect events (e.g., Chang and...
Braham 1991; Kristovich et al. 1999). The mean CBL top height \(z_i\) was near 940 hPa, or about 300 m above the lake surface (ALS) at this location. This may have been the result of reduced stability associated with the nearby cyclone. Interestingly, a local maximum in \(\theta_e\) appears near the Wisconsin shoreline at all levels in Fig. 4. While the data available preclude a detailed documentation of this local maximum in \(\theta_e\), a similar feature has been observed in other cases of lake-induced convection (e.g., Kristovich et al. 1999, 2003). Kristovich et al. (1999) suggested that this feature could be associated with subsidence near the upwind shore, where frictional differences between land and water could create surface divergence and possibly a standing wave in the lower troposphere.

A layer of nearly neutral atmospheric stability between the boundary layer capping inversion and another strongly stable layer at around 750 hPa is evident upwind of Lake Michigan (left side) in Fig. 4. NWS soundings taken at 1200 UTC 5 December 1997 at International Falls, Minnesota (INL); Minneapolis, Minnesota (MPX); GRB; Alpena, Michigan (APX); and Detroit, Michigan (DTX) were examined to determine if the origin of this elevated low-stability layer was due to the influences of Lake Superior. Each sounding, including those upwind of Lake Superior, showed evidence of a layer of near-neutral stability (i.e., a nearly dry adiabatic temperature lapse rate) in the midtroposphere. This suggests that this layer is not associated with airmass modification by the Great Lakes, but is due to larger-scale processes associated with the nearby cyclone.

This low-stability layer had an important influence on the growth of the CBL over Lake Michigan. Upwind of the location where the CBL reached the low-stability layer, the CBL grew in a commonly observed manner, with initially rapid growth (from about \(-88.0^\circ\) to \(-87.5^\circ\) longitude) then slowing growth (to about \(-87.1^\circ\) longitude). Only a small change in CBL depth was seen between KA flight stacks AB and CD. Over this region, the CBL deepened by approximately 490 m across a fetch of 48 km, for a mean rate of depth change of 10.2 m km\(^{-1}\). There are no in situ data available to determine CBL depth between about \(-87.1^\circ\) and \(-86.5^\circ\) longitude. However, by \(-86.5^\circ\) longitude, ISS1 sounding data indicated a well-mixed layer up to approximately 790 hPa. Assuming a linear increase in CBL depth between \(-87.0^\circ\) and \(-86.5^\circ\) longitude, this suggests an average CBL growth rate of about 25 m km\(^{-1}\). Over this region of the lake, the CBL penetrated and progressed rapidly to the top of the upwind low-stability layer.

Averaged across the lake, the CBL growth rate was about 11.5 m km\(^{-1}\). The rapid deepening of the CBL would be expected to contribute to heavier snow production over the eastern half of the lake.

c. Evidence for natural cloud seeding

As midtropospheric short waves circled the central low, they produced southward-moving mesoscale pre-
Precipitation bands in the vicinity of Lake Michigan. Observations from surface, satellite, sounding, and project aircraft indicate that over western regions of Lake Michigan, precipitation from these bands seeded the lake-effect CBL clouds that developed at lower levels over the lake. The most direct evidence for natural cloud seeding over Lake Michigan on 5 December 1997 came from in situ cloud and snow measurements obtained by the KA’s PMS probes. Figure 5 shows percentages of KA flight passes with observed snow (2D-P concentrations over 0.1 L⁻¹) and supercooled cloud (FSSP concentrations over 10 cm⁻³). Over the western half of Lake Michigan (flight stacks AB and CD, left-hand side of Fig. 5), snow frequencies exceeded cloud frequencies at heights near or above the boundary layer top. For example, during pass AB1, conducted well above the boundary layer clouds, snow was detected nearly half of the time. In pass CD3, conducted at a lower altitude but still above z_i, the cloud and snow percentages were 5% and 62%, respectively. The snow observed during passes AB1 and CD3 apparently developed in a higher cloud deck, which fell into the lower supercooled cloud layer at the top of the lake-effect CBL.

NCAR ISS soundings also give supporting evidence that higher cloud decks were present above the lake-effect CBL on this date. Figure 6 shows examples of ISS observed soundings taken upwind and downwind of Lake Michigan. While specific cloud layer heights and depths varied with location and time, project soundings, KA data, and visual observations generally indicate a deep layer of near-saturated conditions, as expected with a nearby cyclone, often with two distinct cloud layers. Observations by the NCAR ELDORA radar frequently indicated a weak precipitation layer above the lake-effect CBL (Fig. 7).

KGRB WSR-88D radar observations and regional satellite observations on 5 December 1997 show the presence of generally southward moving mesoscale precipitation bands over western Lake Michigan. Figure 8 shows the reflectivity field at two times in the vicinity of KA flight stack AB. In general, regions of radar-observed reflectivity (maximum values 5–10 dBZ) had horizontal dimensions of 10–30 km. An overall decrease in reflectivity regions was seen during flight stacks AB and CD. At a range of approximately 90–95 km, near KA flight stacks AB and CD, the center of the lowest elevation radar beam (0.5°-elevation angle) from KGRB would be more than 400 m higher than the observed CBL depths (850–900 m).

To estimate the potential contribution from snow in the lake-effect CBL to reflectivity values observed by KGRB, Schroeder (2002) calculated the portion of the radar beam potentially filled by CBL targets (snow), weighted by the Gaussian gain function of the beam. Values of beam-center height and half-power beam-width for a range of 93 km were obtained from Sirvatka (2001). Using radar reflectivities calculated from 2D-P observations within the top portions of the lake-effect
cloud deck (flight legs AB2 and AB3, approximately −5 to 1 dBZ), the contribution to the KGRB radar observation would be minimal: up to approximately −9 dBZ. If it is assumed that the CBL has a constant reflectivity from the surface to the boundary layer top of 5 to 10 dBZ, values occasionally observed in the CBL by the ELDORA, then the contribution could possibly be as much as 1 to 2 dBZ. This seems somewhat unrealistic, however, since CBL reflectivities have generally been observed to decrease upward through the cloud layer (Kristovich and Braham 1998).

Additional information on the source of the KGRB WSR-88D-observed reflectivity features over western Lake Michigan is gained by examining their evolution and movement. Wind profiles measured by ISS2 at 1500 and 1800 UTC (Fig. 9) indicated near-surface WNW boundary layer winds of 8 to 15 m s⁻¹ backing to the NNW−NNE above the CBL top. This is in general agreement with aircraft observations. The average movement of the reflectivity features observed by KGRB implies an advecting northerly wind (350°–360°, 6 to 10 m s⁻¹). Such winds are clearly not found in the boundary layer, but are present over a deep layer above the CBL. This suggests that the precipitation detected by the KGRB WSR-88D near the western KA flight stacks (AB and CD) must have been generated above the boundary layer. WSR-88D reflectivity observations will be used in the aircraft data analyses to determine locations of natural cloud seeding.

Both operational and Lake-ICE datasets provide evidence that the CBL clouds over Lake Michigan were naturally seeded in some locations on 5 December 1997. These data include surface observations, satellite imagery, radar observations, upwind and downwind ISS vertical profiles, and in situ cloud and snow microphysical data collected by the KA.

4. Local effects of natural cloud seeding

a. Determination of seeded portions of aircraft tracks

To use aircraft observations to examine the influence of natural cloud seeding on the lake-effect boundary layer, radar reflectivity observations by the KGRB WSR-88D are used as an independent dataset to identify portions of the aircraft track that are in seeded regions. A comparison between aircraft observations of snow above the CBL with the locations of radar-observed precipitation were carried out to establish their spatial correlation. Equivalent reflectivity factors were calculated using methods given in Sekhon and Srivastava (1970) and Smith et al. (1975), adjusted for particle density (provided by P. Smith 2005, personal communication):

\[ Z_e = \frac{720N_0}{\lambda^2} \rho^{7/3}. \]  

Figure 10 shows results of this analysis for one such pass above the lake-effect CBL over western Lake Michigan (AB1). The reflectivity field corresponding with this flight leg is shown in Fig. 8a. While reflectivity values calculated from the KA 2D-P probe observations were often different, locations of precipitation observed by the KA above the lake-effect boundary layer corresponded well with regions of higher reflectivity values measured by the KGRB WSR-88D radar. This analysis suggests that the WSR-88D observations can be used as an independent data source to identify approximate regions of natural cloud seeding.

Criteria chosen to identify portions of the aircraft flight legs in seeded regions are based on the findings described in section 3c. Since the patches of reflectivity observed by the KGRB WSR-88D were generally 10–30 km across, it was required that seeded portions of
Fig. 8. WSR-88D radar-observed reflectivity over central Lake Michigan region on 5 Dec 1997: (a) 1639 and (b) 1659 UTC. Data shown are from the 0.5°-elevation scan. White arrows denote approximate locations of the KA during Flight Stack AB. Anomalous reflectivities are seen radiating to the SE of the radar site. The radar image was prepared using GRLevel3 software, version 1.0.0.2.
the flight legs exceed a minimum distance in these patches. Specifically, seeded portions were defined as those along which the aircraft’s path intersected at least five consecutive radar gates with reflectivities above a threshold value. Only one below-threshold gate embedded within the seeded portion of the flight leg was allowed. While calculations in section 3c suggest that the CBL would contribute minimally to the radar-observed reflectivity values, we employed a 0-dBZ threshold for seeded portions of the flight legs AB2, AB3, and AB4. Lower thresholds gave qualitatively similar results for these passes. It should also be noted that if lower thresholds were used, then passes AB5 and CD1 would also have seeded portions with similar results to those reported below.

b. Microphysical characteristics

Since several of the KA passes in the westernmost flight stack (labeled AB in Fig. 1) were within both seeded and nonseeded regions, comparison of in situ observations from these regions enabled us to determine the effects of natural cloud seeding on the snow-particle spectra. Ice-particle-size spectra observed by the 2D-P probe were averaged for all seeded and nonseeded portions of each KA flight leg. The duration of each sample was on the order of several minutes (tens of kilometers). These sample durations are comparable to, or larger than, those in Braham (1990).

Figure 11 gives average spectra for seeded and nonseeded regions of each KA flight leg in the westernmost flight stack (AB). The spectra were well represented by least squares–fit exponential functions, with $R^2$ (correlation coefficient squared) values greater than 0.96 for all spectra. Sampling durations, spectral parameters $N_0$ and $\lambda$, and derived quantities from the distributions appear in Table 2. In all passes, snow tended to be considerably more intense in seeded areas, with larger intercept ($N_0$) values by factors of 2 to 3.5, and slopes ($\lambda$’s) 1.3 to 1.8 times smaller than their nonseeded counterparts. Thus, the integrals beneath the seeded exponentials, which are proportional to IWC, are appreciably larger in seeded regions. For example, the average of the downward SF values for seeded portions of the flight tracks was 2.71 mm day$^{-1}$ versus 0.22 mm day$^{-1}$ for nonseeded segments, more than an order of magnitude difference. The maximum observed average water-equivalent snowfall rate was 4.28 mm day$^{-1}$. For comparison, aircraft sample-mean SF values reported by Braham and Dungey (1995) over Lake Michigan were 1.40 mm day$^{-1}$ for wind-parallel band cases and 2.23 mm day$^{-1}$ for midlake/shoreline band events.

WSR-88D observations indicate that the amount and areal coverage of seeding decreased throughout flight stacks AB and CD. Since aircraft observations were not continuously available above the CBL during this time...
period, it is not possible to quantify the amount of seeding of snow into the lake-effect clouds. However, a comparison of two flight legs taken within a 16-min time period (AB1 and AB2) can help illustrate the effects of seeding. The average SF during flight leg AB1, conducted entirely above the lake-effect CBL, was 5.4 mm day\(^{-1}\). This is more than an order of magnitude greater than the SF near the CBL top (flight leg AB2) in nonseeded areas, 0.25 mm day\(^{-1}\). However, SF in areas of seeding in flight leg AB2 were much greater than those in nonseeded portions of the same pass and comparable to the above-CBL SF rates, averaging 4.28 mm day\(^{-1}\). The significant downward flux of ice near CBL top is a unique consequence of seeding into the lake-effect cloud layer.

It should be noted that there is a trend in Fig. 11 and Table 2 for decreasing snow intensity with decreasing height (from CBL top toward the lake surface) in stack AB. This result is inconsistent with previous investigations of lake-effect events (e.g., Kristovich and Braham 1998). As previously noted, KGRB WSR-88D reflectivity images indicate that seeding precipitation in the KA’s operations area was decreasing in both intensity and areal coverage during the time and location of the westernmost flight stack (1638–1727 UTC). Aircraft observations also show similar changes. For example, a similar snowfall rate may have been expected in passes AB2 and AB5, which were conducted at the same altitude about 30 min apart. However, average SF values in AB2 were much greater than those in AB5 (1.74 versus 0.39 mm day\(^{-1}\)). Since lower passes in flight stack AB occurred at increasingly later times, it is concluded that the observed decrease in snow intensity with decreasing height was likely due to the local decrease with time in the intensity of snow entering the CBL from above.

c. Local variations in boundary layer depth

Previous studies of lake-effect CBLs (e.g., Chang and Braham 1991; Cooper et al. 2000), and similar types of marine CBLs (e.g., Boers and Melfi 1987), suggest that additional latent heat release may result in increased convective intensity and deepening of the CBL. To investigate whether this process occurred in seeded regions on 5 December 1997, time series of potential temperature and supercooled cloud droplet total concentration were examined for each of the KA flight passes that had seeded segments in the western (flight stack AB) regions of Lake Michigan, before the CBL merged with an elevated low-stability layer. Figure 12 gives time series for several passes in flight stack AB. Note that since the KA passes were oriented crosswind, the impacts of cross-lake CBL growth on along-pass variations were minimized. The passes show features implying that the CBL was locally deeper in areas of cloud seeding. For example, in the seeded portion of the pass AB2 at 820-m height (Fig. 12a), the cloud droplet concentrations and potential temperatures were inversely related, with potential temperatures generally more than 1 K colder in clouds than in cloud-free regions. This feature is commonly observed near or just above the mean boundary layer top height, \(z_t\) (Stull 1988; Agee and Hart 1990). In the nonseeded portion of the pass, however, clouds were much less frequent, and environmental potential temperatures were \(\sim 0.5\) K warmer, suggesting that the KA was sampling further above \(z_t\). Since the KA was in level flight, this implies that the CBL was locally deeper.
in the seeded portion of the pass. Similar features were observed in the pass AB5 at 821-m height (not shown).

In addition, local deepening of the CBL in seeded regions was indicated by observations lower in the lake-effect CBL. Potential temperature and cloud droplet concentrations were generally out of phase along the two lower-level passes (AB3 and AB4; Figs. 12b and 12c), suggesting the pass was still close to \( z_i \). Again, clouds were more frequent in the seeded area than in the northern nonseeded area and the mean potential temperature increased by about 1 K northward from the seeded area to the nonseeded area. These observations strongly suggest CBL deepening occurred in the seeded areas.

### Table 2

<table>
<thead>
<tr>
<th>KA pass</th>
<th>Sample duration [s (km)]</th>
<th>( N_0 ) (mm(^{-3}) L(^{-1}))</th>
<th>( \lambda ) (mm(^{-1}))</th>
<th>IWC ((10^{-3} \text{ g m}^{-3}))</th>
<th>Water-equivalent SF (mm day(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seeded</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AB2</td>
<td>180 (14.4)</td>
<td>3.16</td>
<td>1.60</td>
<td>38 (56)</td>
<td>4.28</td>
</tr>
<tr>
<td>AB3</td>
<td>117 (9.4)</td>
<td>4.04</td>
<td>1.89</td>
<td>25 (47)</td>
<td>3.37</td>
</tr>
<tr>
<td>AB4</td>
<td>148 (11.8)</td>
<td>1.15</td>
<td>2.31</td>
<td>3.2 (7.0)</td>
<td>0.47</td>
</tr>
<tr>
<td>Nonseeded</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AB2</td>
<td>308 (24.6)</td>
<td>0.97</td>
<td>2.83</td>
<td>1.2 (3.9)</td>
<td>0.25</td>
</tr>
<tr>
<td>AB3</td>
<td>514 (41.1)</td>
<td>1.14</td>
<td>2.49</td>
<td>2.3 (5.9)</td>
<td>0.40</td>
</tr>
<tr>
<td>AB4</td>
<td>359 (28.7)</td>
<td>0.59</td>
<td>3.26</td>
<td>0.4 (0.1)</td>
<td>0.003</td>
</tr>
</tbody>
</table>

In the seeded portion of the pass. Similar features were observed in the pass AB5 at 821-m height (not shown).

In addition, local deepening of the CBL in seeded regions was indicated by observations lower in the lake-effect CBL. Potential temperature and cloud droplet concentrations were generally out of phase along the two lower-level passes (AB3 and AB4; Figs. 12b and 12c), suggesting the pass was still close to \( z_i \). Again, clouds were more frequent in the seeded area than in the northern nonseeded area and the mean potential temperature increased by about 1 K northward from the seeded area to the nonseeded area. These observations strongly suggest CBL deepening occurred in the seeded areas.

**Fig. 12.** Time series of FSSP total cloud drop concentration (black; scale at left in cm\(^{-3}\)) and potential temperature (gray; scale at right in K) arranged by altitude for KA passes in the westernmost flight stack. Gray bars indicate divisions between seeded and nonseeded segments of the flight legs, based on WSR-88D observations of 0 dBZ. Vertical dashed lines indicate the edges of WSR-88D-measured reflectivity patches (~3.5 to ~4.5 dBZ). Mean altitudes for each pass are indicated.
5. Discussion

The primary findings of this investigation of the 5 December 1997 synoptically enhanced lake-effect snow case are 1) rapid cross-lake CBL growth was observed, 2) the CBL was locally deeper in regions of natural cloud seeding, and 3) snow was locally intensified in regions of natural cloud seeding. This section discusses the scientific significance of these findings.

a. Cross-lake CBL growth

Rapid cross-lake lake-effect CBL growth was observed over Lake Michigan on 5 December 1997, despite marginal environmental conditions for lake-effect snows (e.g., Niziol et al. 1995). For comparison, Table 3 lists total surface fluxes and rates of cross-lake CBL growth for cases of widespread lake-effect convection from several investigations (adapted from Kristovich et al. 2003). For the current case, WNW surface winds of 5 to 8 m s\(^{-1}\), lake-surface temperatures of 5\(^\circ\) to 7\(^\circ\)C (GLERL 2000), and surface air temperatures of \(-1\)\(^\circ\) to \(-5\)\(^\circ\)C combined to produce total (sensible plus latent) estimated heat fluxes of 100–200 W m\(^{-2}\) (Laird et al. 2001). Surface heat fluxes observed on 5 December 1997 are in the lower range of values of 100–700 W m\(^{-2}\) from these previous studies. Despite comparatively low heat fluxes, interaction with the synoptic weather system on 5 December 1997 enabled CBL growth rates to be higher than those reported in these previous studies. Over a midlake portion of the lake, the average growth rate was estimated to be more than twice those previously reported. This, in turn, contributed to the development of heavy snow over western Lower Michigan.

Stability in the lowest \(-60\) hPa (up to about 500 m ALS) of the atmosphere upwind of Lake Michigan (Fig. 4), which limits CBL growth forced by surface fluxes, was weaker than in most previous lake-effect studies (i.e., Chang and Braham 1991; Kristovich 1993; Kristovich et al. 2003). Above the boundary layer capping inversion, a layer of near-neutral static stability (\(-200\) hPa thick) was present, in association with a midtropospheric cyclone centered over the Great Lakes. As this less-stable air mass passed over Lake Michigan, the resulting CBL reached about 1780 m near the downwind shore. The resulting CBL growth rate was greater than previous studies with low-level winds oriented perpendicular to the long axis of Lake Michigan (Table 3), in part due to the merging of the lake-effect CBL and the higher-level low-stability layer.

Similar processes were observed by Chang and Braham (1991), who reported a nearly discontinuous change in CBL growth rate at a fetch of \(-35\) km across Lake Michigan on 20 January 1984. As the CBL penetrated a midlevel near-neutral layer, the CBL rapidly grew. Chang and Braham (1991) argued that the increase in growth rate could be due to either latent heat release during snow formation or the penetration of the CBL with the higher-level near-neutral layer. Results from the present case suggest that latent heat release played a secondary role on this day.

b. Local CBL depth changes in seeded areas

Aircraft in situ and KGRB WSR-88D radar reflectivity features indicate the presence of regions of natural seeding of the lake-effect CBL by snow from higher-altitude cloud decks. For this study, WSR-88D data were used to determine locations along KA flight legs over western Lake Michigan where seeding was occurring. The use of operational radar data to determine regions of natural cloud seeding represents a new approach; for this case, it provided a valuable technique.
for comparison of CBL characteristics between seeded and nonseeded areas.

Use of WSR-88D data and aircraft-observed time series of potential temperature and total cloud drop concentration from KA crosswind passes imply that the CBL over Lake Michigan on 5 December 1997 was locally deeper in areas seeded from above. Local deepening was likely not influenced by cross-lake CBL growth because 1) aircraft passes were conducted nearly perpendicular to the CBL wind, and thus at nearly constant fetch from the upwind shore, and 2) the locations of seeding (as determined from WSR-88D observations) changed from pass to pass, so a geographic influence on CBL depth is unlikely.

There are three potential causes of the local deepening. First, the local deepening could have been caused by increased latent heat release in the lake-effect boundary layer clouds in seeded regions. Chang and Braham (1991) argued that one possible cause of (in their case, cross-lake) CBL growth was latent heat release in cloud and snow development, in agreement with past CBL simulations (e.g., Boers and Melfi 1987). Additional latent heat release through increased ice formation in seeded regions could amplify the local mesoscale circulations (Hjelmfelt 1990), leading to local increases in CBL depth. Temporal changes in seeder precipitation intensity during the current study made it difficult to quantify the amount of latent heating that may have been added.

Another possible mechanism for local CBL deepening in seeded areas is the evaporation of ice particles in the subsaturated layer just above the CBL. Cooling associated with the evaporation might have decreased the stability of this layer, allowing CBL convection to locally raise $z_c$. Since only one KA pass was conducted entirely above the CBL in areas with seeding, and since there were no passes conducted higher in the subsaturated layer, it is difficult to determine spatial variations in lapse rate using the Lake-ICE data from 5 December 1997. However, it should be noted that relative humidities with respect to ice were nearly always above 80% along the entire flight leg, which was in the cloud-free layer, and no local cooling of the air was noted. This implies that evaporation would have been minimized, therefore diminishing the likely importance of this effect.

A third possible reason for the local CBL deepening is that seeded regions were collocated with regions of mesoscale updraft (associated with the synoptic cyclone-induced cloud bands). Such regions of updraft could act to produce precipitation in higher-level cloud decks as well as contribute to CBL deepening. If this were the case, then natural seeding itself would be of little direct consequence to the local deepening. Operational Eta Model output from 5 December 1997 was examined in an attempt to verify this scenario, but the resolution was not sufficient to distinguish vertical velocities on the scale of KA pass lengths (several tens of kilometers). King Air data from stack AB did not provide enough spatial coverage in the cross-lake direction to adequately sample such mesoscale circulations. A numerical mesoscale modeling study of this or a similar case should be used to investigate the possibility of the collocation of seeding areas and mesoscale updrafts more closely.

c. Local microphysical and snow intensity differences in seeded areas

This study documents local changes in snow-particle-size spectra and increases of snow intensity in naturally seeded regions of the lake-effect CBL on 5 December 1997 (Table 2). In-cloud IWCs were calculated in the same manner as Braham [1990; Eq. (2) in the current paper] and, for comparison, using bin-by-bin particle concentrations [Eq. (3)]. IWC calculated using Eq. (2) ranged from $3.2 \times 10^{-3}$ to $3.8 \times 10^{-2}$ g m$^{-3}$ in seeded areas and $4 \times 10^{-4}$ to $2.3 \times 10^{-3}$ g m$^{-3}$ in nonseeded areas. A similar difference was seen in IWC calculated by Eq. (3), but the individual IWC values were higher [due primarily to slightly underestimated concentrations in small particle size bins when estimating the spectra with an equation of form Eq. (1)]. The difference between seeded and nonseeded IWC values suggests that seeding was a key process in locally intensifying snowfall.

Overall, IWC values are on the lower end of those determined by Braham [1990, using Eq. (2)] for five classic lake-effect events (0.003 to 0.321 g m$^{-3}$). Two differences between the current study and Braham (1990) may account for this difference. First, the current samples were mainly within upper portions of the lake-effect cloud layer, while Braham’s measurements were made below the lake-effect clouds. Ice contents would be expected to be small in upper portions of the CBL (Kristovich and Braham 1998). In addition, snow intensity generally increases across the lake from upwind to downwind shores. The current observations focused on areas over the upwind half of Lake Michigan, while Braham (1990) examined snow characteristics near the downwind shore.

Other previous studies calculated downward SFs using methods similar to the present study. Braham and Dungey (1995) reported aircraft SF values of 0 to 4.2 mm day$^{-1}$ below 500-m altitude for widespread wind parallel band cases and up to 5.5 mm day$^{-1}$ for midlake/shoreline band cases. Chang and Braham (1991) found...
the average surface SF in the 20 January 1984 case to be 0.8 mm day\(^{-1}\). In the current study, SF values in seeded areas were generally on the high end of observations in these previous studies (up to 4.28 mm day\(^{-1}\)). Nonseeded SF values (up to 0.40 mm day\(^{-1}\)) were well within SF values reported by Braham and Dungey (1995) and Chang and Braham (1991), further illustrating the importance of seeding on snowfall rates on 5 December 1997.

Studies of seeding effects in other meteorological situations, while not directly comparable, had some interesting differences from the current study. The SF increases in the present study are much larger than the range of 0.24 to 1.68 mm day\(^{-1}\) increases reported by Locatelli et al. (1983) for a synoptic case in which a 1-km-thick stratocumulus layer was seeded by dendritic crystals originating in altocumulus. They found an increase of 2–8 times from above the stratocumulus layer to the bottom. The reason for the greater degree of precipitation enhancement on 5 December 1997 may have been in part due to the higher SFs entering the feeder layer from above [near 0.24 mm day\(^{-1}\) in Locatelli et al. (1983), compared with 4.28 mm day\(^{-1}\) in pass AB2]. However, other environmental or microphysical characteristics of the two cases likely played significant roles as well. Lack of observations throughout the cloud layer in the present case, and temporal changes in the seeding layers, prevent a detailed comparison. However, these topics may lend themselves to numerical modeling studies.

6. Conclusions and final remarks

Lake-effect snowstorms can have great practical and economic impacts upon those living near the downwind shores of the Laurentian Great Lakes. To improve the forecasting of these events, which has proven difficult (Niziol 1987), the scientific community must develop a better understanding of processes involved in complex lake-effect situations. The current study explores the effects of a synoptic cyclone on the lake-induced CBL that developed over Lake Michigan on 5 December 1997 (a so-called lake-enhanced event), the first operational day of Lake-ICE (Kristovich et al. 2000). To our knowledge, a detailed observational study of this sort of case has not been previously carried out.

It was postulated that the cyclone had the following three effects on the CBL: 1) cross-lake CBL growth was accelerated, compared with classic cases, 2) the CBL was locally deepened in regions of natural cloud seeding, apart from the cross-lake growth, and 3) microphysical snow-growth processes were locally enhanced, within the CBL clouds, in seeded regions. Results of the current study strongly suggest that cross-lake CBL growth observed was much faster at midlake regions than observed in a classic case with similar surface fluxes, the CBL was locally deeper in seeded areas, and snowfall rates were greatly intensified in seeded areas.

Evidence from surface observations, satellite data, Lake-ICE project soundings, and in situ aircraft microphysical measurements strongly suggests that natural cloud seeding was occurring at certain locations over Lake Michigan on 5 December 1997. WSR-88D data from KGRB were found to be useful, independent indicators of seeding locations for this case. Based on these reflectivity data, crosswind KA passes near the western shore were parsed into seeded and nonseeded segments and compared to determine the effects of seeding.

In situ aircraft observations indicate that the lake-effect CBL was deeper in seeded regions. The deepening is thought to have been associated with either enhanced latent heat release produced by seeding or with regions of mesoscale updraft, which may have acted to lift the entire CBL. Evaporative cooling of seeding particles in the subsaturated air above the CBL, and associated destabilization, is another possible explanation.

Ice-particle-size spectra were analyzed to determine microphysical differences between seeded and nonseeded areas. Seeded spectra, in all cases, implied more intense snowfall than their nonseeded counterparts. Derived parameters, such as IWC and SF, also reflected this difference.

The present study poses new questions for future work. Questions arising directly from results presented here include 1) What was the cause (e.g., latent heat release, mesoscale updraft, or evaporative destabilization) of the local CBL deepening observed in seeded regions over Lake Michigan on 5 December 1997? 2) Is snowfall enhancement in regions of natural cloud seeding in a lake-enhanced event typical and are the mechanisms inferred for this case typical? 3) How did liquid water contents in this case compare with conditions typically encountered in classic lake-effect cases and synoptic cyclones originating over land and water? Were liquid water contents sufficient to suggest that accretion/riming was the primary snow growth mechanism for ice particles entering the CBL clouds from above? 4) What is the proper interpretation, in terms of physical processes, of the vertical changes in snow spectra observed in the 5 December 1997 case? 5) How would a water budget, such as that studied by Chang and Braham (1991) for a more classic lake-effect case, be modified in a lake-enhanced case such as 5 December 1997, where significant downward fluxes of ice were entering the CBL from above?
Answers to some of these questions perhaps lie in the Lake-ICE data; others could be addressed by detailed numerical experiments or new field investigations of lake-effect/enhanced processes. More work is needed to confirm the results of the current study, since several features have been documented for the first time.

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