Lake-Effect Thunderstorms in the Lower Great Lakes

SCOTT M. STEIGER
Department of Earth Sciences, State University of New York at Oswego, Oswego, New York

ROBERT HAMILTON
National Weather Service Forecast Office, Buffalo, New York

JASON KEELER*
Department of Earth Sciences, State University of New York at Oswego, Oswego, New York

RICHARD E. ORVILLE
Department of Atmospheric Sciences, Texas A&M University, College Station, Texas

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ABSTRACT

Cloud-to-ground (CG) lightning, radar, and radiosonde data were examined to determine how frequently lake-effect storms (rain/snow) with lightning occurred over and near the lower Great Lakes region (Lakes Erie and Ontario) from September 1995 through March 2007. On average, lake-effect lightning occurred on 7.9 days and with 5.8 storm events during a particular cool season (September–March). The CG lightning with these storms had little inland extent and was usually limited to a few flashes per storm. Some storms had considerably more, with the most intense storm (based on National Lightning Detection Network observations) producing 1551 CG flashes over a 4-day period. Thundersnow events were examined in more detail because of the rarity of this phenomenon across the United States. Most lake-effect thundersnow events (75%) occurred in November and December. An analysis of model sounding data using the Buffalo Toolkit for Lake Effect Snow (BUFKIT) software package in which lower boundary conditions can be modified by lake surfaces showed that thundersnow events had an 82% increase in the mean height of the −10°C level when compared with nonelectrified lake-effect snowstorms (1.2 vs 0.7 km AGL), had higher lake-induced equilibrium levels (EL; above 3.6 km AGL) and convective available potential energy (CAPE; >500 J kg⁻¹), had low wind shear environments, and were intense, single-band storms. A nomogram of the altitude of the −10°C isotherm and EL proved to be useful in predicting lake-effect thundersnowstorms.

1. Introduction

Lake-effect thunderstorms are of special interest because they occur during cold-air outbreaks when the air over the land is stable and citizens do not expect electrical storms. Moore and Orville (1990) observed that lake-effect thunderstorms (rain/snow) tend to be single-banded storms, which occur when the mean low-level flow in the atmosphere is parallel to the long axis of the lake (e.g., west-southwesterly winds over Lake Erie and west-northwesterly over Lake Ontario after a cold-frontal passage). Five lake-effect snowstorm structures have been identified by Niziol et al. (1995). Type-1 snow-bands form parallel to the longest axis of a lake and are usually associated with the most intense snowfalls and likely lightning because the distance (fetch) the air travels over the water is the greatest. Kristovich and Steve (1995) show a greater percentage of lake-effect storms are type-1 bands across Lakes Erie and Ontario when compared with their upwind counterparts (Lakes
Superior, Huron, and Michigan) because the prevailing flow (westerly) during the winter is more parallel to the longest axes of the lower Great Lakes. Hence, we have chosen to focus our analysis on Lakes Erie and Ontario because lake-effect lightning is more likely to occur in this region.

Lake-effect snowstorms are impressive (snowfall rates > 5 cm h⁻¹) mesoscale events that are occasionally accompanied by lightning. Although less than 0.1% of snow reports in the United States are accompanied by thunder (Curran and Pearson 1971), roughly 20% of the significant lake-effect snows (>15.2 cm (6 in.) in 12 h) across western New York are lightning-bearing events (Buffalo, New York, National Weather Service Forecast Office 2007, personal communications).

These narrowbanded storms produce lightning mainly in the early part of the lake-effect season [September–December; Moore and Orville (1990)]. The entire lake-effect season is defined as the months during which much-colder air can cross the warmer Great Lakes and produce these storms: typically September–March. It is thought that the potential for lake-effect storms with cloud-to-ground (CG) lightning is greater during these earlier months because of the warmer lake surface temperatures and the greater convective cloud depths relative to the rest of the lake-effect season. A decrease in heat and moisture fluxes from the lake surface and an increase in overall convective stability by late December are causes for the early-season bias of these storms (Moore and Orville 1990). In addition, the air’s moisture content is typically greater during the September–December period than it is later because of the warmer ambient temperatures.

Because most lake-effect thunderstorms occur early in the cool season (Moore and Orville 1990), a warmer lake surface and convective boundary layer are critical in the development of these storms. In-cloud temperature profiles closer to 0°C are more favorable for graupel formation. Graupel plays a crucial role in separating electrical charge through frictional collisions with smaller ice crystals and supercooled water droplets (Reynolds et al. 1957, Takahashi 1978, 1984). MacGorman and Rust (1998, 61–75) give a thorough description of the current hypotheses for charge-separation mechanisms in thunderstorms. One of the most accepted hypotheses involves noninductive (no background electric field) charge separation with ice crystals becoming positively charged as they ascend and heavier graupel particles becoming negatively charged as they descend after collision in a typical thundercloud. Current electrification models strongly support the theory that graupel production in the layer from −10°C to −20°C is essential for the initial stages of lightning development (Zajac and Weaver 2002).

The charge structure of typical summertime thunderstorms can be described as a tripole [weak positive charge center near cloud base, main negative charge region above, and main positive charge center near the top of the storm; MacGorman and Rust (1998, 51–52)]. However, very little is known about the electric field aloft in wintertime storms such as lake-effect storms (MacGorman and Rust 1998, p. 288). In addition, any association between snowfall rate and lightning occurrence is inconclusive (MacGorman and Rust 1998, p. 290).

Forecasting lake-effect thunderstorms has been difficult because environmental parameters that adequately discriminate between lightning and nonlightning events have not been found. For example, observed proximity soundings (near lake) associated with lake-effect thunderstorms versus nonelectrified storms show small or no differences (Schultz 1999). Based on these proximity soundings, Schultz concluded that there is little or no convective available potential energy (CAPE) present in lake-effect storms with or without lightning. These observations may not be representative of the conditions over the lake where these storms develop. Hence, the lake temperature and its effect on the lower troposphere are accounted for in this study through the use of the Buffalo Toolkit for Lake Effect Snow (BUFKIT) software package (Mahoney and Niziol 1997) and its ability to modify numerical model output to better incorporate effects of local surface characteristics (i.e., lakewater). A major goal of this study is to reevaluate some of Schultz’s analysis (e.g., of CAPE) in discriminating environments favorable for lake-effect lightning from those that do not produce lightning using the “lake induced” parameters calculated in BUFKIT in several case studies [e.g., lake-induced (L-I) CAPE, L-I equilibrium level].

Jiusto and Holroyd (1970) noted that the passage of a midtropospheric short-wave trough can enhance a lake-effect band by increasing the depth of the convective boundary layer. As a result, the trough passage can increase the depth of the lake-effect cloud and the layer in which charge separation occurs. An analysis of lightning associated with convective snowfall in the Hokuriku District of Japan suggests that cold environments are less favorable for lightning occurrence, because no lightning flashes were observed when the altitude of the −10°C isotherm was below 1.4 km (Michimoto 1993).

A climatological analysis of lake-effect lightning events (all precipitation types and only snow) for Lakes Erie and Ontario based on 12 yr (1995–2007) of lightning-flash data for the cool (September–March) season is provided. This study represents the first comprehensive climatological description of lake-effect lightning events. The CG lightning-flash data were obtained from
the National Lightning Detection Network (NLDN; Cummins et al. 2006).

Results of this study on thundersnow events in the lower Great Lakes suggest that if the depth of the layer from the surface to \(-10^\circ C\) is too shallow, then sufficient amounts of graupel are likely not produced to actively separate charge given a convective element of lake-effect snow and lightning does not occur. Various other convective parameters such as CAPE and equilibrium level (EL) are also examined to determine their possible contributions to lightning production in lake-effect snowstorms. The effect of the underlying lakewater temperature and its modification of the lower boundary layer will be shown to be very important in diagnosing lake-effect thunderstorm environments.

2. Data and methodology

a. Climatological study

Dates (defined as a day’s time period from 0000 to 2359 UTC) with CG lightning flashes in lake-effect storms were determined through a multiple-step elimination process. NLDN (Cummins et al. 2006) data were analyzed to determine days with CG lightning between 1995 and 2007. These data were obtained from Vaisala, Inc. The current NLDN only measures CG lightning flashes whose median location accuracy over the Great Lakes region is better than 500 m, and the flash detection efficiency is greater than 90% (Cummins et al.). Even though the NLDN has been in operation since 1989, limitations in archived Weather Surveillance Radar-1988 Doppler (WSR-88) data prevented analysis before 1995 [all of the Next-Generation Doppler Radar (NEXRAD) network radars were not online until the mid-1990s]. Days with CG flashes that occurred from September through March, which we defined as the lake-effect season, were retained. The domain included the lower Great Lakes (Lakes Erie and Ontario, eastern Lake Huron, and the Georgian Bay) and surrounding land areas (from 40.3° to 45.5°N and from 83.75° to 75.0°W). Once dates with CG lightning from September through March were determined, radar base reflectivity imagery (NCDC 2007) from each of these days was used to filter out dates without lake-effect storms. Days associated with obvious non-lake-effect features (e.g., thunderstorm squall lines) were eliminated from the dataset.

Lake-effect snowstorms (the previous methods applied to lake-effect lightning that occurred with any precipitation type) that occurred on Lakes Erie and Ontario were examined for the presence of lightning from the winters of 1995/96 through 2006/07. The lake-effect snowstorms in this part of the study were defined as producing at least 17.8 cm (7 in.) of snow in 12 h. This snowfall accumulation is used by the National Weather Service (NWS) Eastern Region to define lake-effect winter storms for the NWS Buffalo County Warning Area (National Weather Service 2005). Nearly 200 lake-effect snow events were identified over Lakes Erie and Ontario. Because synoptic snow events were excluded from this dataset, forcing from organized midtropospheric short waves and frontal passages was excluded.

The thermal profiles used in this part of the study were supplied by simulated (forecast) proximity soundings from the operational Eta/Weather Research and Forecasting-Nonhydrostatic Mesoscale Model (WRF-NMM) binary universal form for the representation of meteorological data (BUFR) data for Buffalo, Jamestown, Rochester, Watertown, and Syracuse, New York. BUFR is a World Meteorological Organization standard for storing data, including upper-air information (Alpert 2008). BUFR data contain the native-vertical-
resolution, uninterpolated model output. Because this format can store a large amount of data and metadata in a small amount of disk space, it is ideal for using in programs such as BUFKIT (Mahoney and Niziol 1997), in which it can be read in its native format. The initial model time (0000 or 1200 UTC) for the BUFKIT analyses was the time immediately prior to the lightning observations or, in the case of no lightning, the time prior to the most unstable (e.g., environmental lapse rates approaching dry neutral) forecast period of the lake-effect event. We used mean lake surface temperatures from the Great Lakes Environmental Research Laboratory (GLERL 2007) and the Erie County Water Authority for Lake Erie to modify the Eta/WRF-NMM BUFR model input through the equations given in Phillips (1972) to better represent lake boundary layer conditions (usually warmer and more moist than the original model data). Hence, the use of BUFKIT enabled a more realistic CAPE value to be calculated over the lake. Simulated lake-induced CAPE calculations in BUFKIT incorporate heat and moisture fluxes into the boundary layer from an underlying lake surface. In this case, the waters of Lakes Erie and Ontario add potential energy to the parcel. We chose parameter values, such as the lake-induced CAPE, nearest in time to the onset of lightning for each event. For activity near Lake Erie, the Buffalo data were analyzed; near Lake Ontario, Watertown/Rochester data were used depending on band proximity to each location. When there was no lightning reported for an event (i.e., nonlightning event), the highest value of CAPE during the lake-effect storm event was used.

The lightning data were obtained from two main sources, with the majority of the information being gathered from reports from volunteer snow spotters. The second source was the NLDN. The National Weather Service in Buffalo started an extensive snow-spotter network in the mid-1990s (Fig. 1). This network was developed to better define and verify lake-effect snowfall in real time. Over 200 snow spotters were asked to report snowfall amounts and rates in given time increments and, in addition, were asked to report the occurrence of any lightning and thunder. These reports included intracloud (IC) and CG lightning flashes. Additional CG lightning-strike information was obtained from the NLDN. When possible, these lightning reports were compared with the Buffalo (KBUF) and Montague (KTYX), New York, radars for 35-dBZ echo verification (Gremillion and Orville 1999; Hodanish et al. 2004). This assisted in distinguishing between nighttime lightning and possible power-line or transformer flashes.

FIG. 1. Snow-spotter locations (provided through the courtesy of the NWSFOs in Buffalo and Binghamton) throughout western and north central NY used in this study. Country, state, and selected county outlines are shown. Locations discussed in this paper are shown, and an inset map is given to show locations relative to all of the Great Lakes.
Table 1. Maximum values of lake-induced CAPE (L-I CAPE) and EL for the case-study events with and without lightning.

<table>
<thead>
<tr>
<th>Case study</th>
<th>L-I CAPE (J kg⁻¹)</th>
<th>EL (km)</th>
<th>Lightning</th>
</tr>
</thead>
<tbody>
<tr>
<td>28–31 Jan 2004</td>
<td>845</td>
<td>3.7</td>
<td>No</td>
</tr>
<tr>
<td>12–13 Oct 2006</td>
<td>2400</td>
<td>7.7</td>
<td>Yes</td>
</tr>
<tr>
<td>3–12 Feb 2007</td>
<td>1790</td>
<td>5.3</td>
<td>Yes</td>
</tr>
</tbody>
</table>

C. Case studies

Specific case studies utilizing the BUFKIT software (Mahoney and Niziol 1997) were conducted for lake-effect storms with and without lightning as in section 2b. BUFKIT can modify model output over bodies of water (e.g., the Great Lakes) to depict more accurately the lower levels of the atmosphere. Once again simple equations were used in BUFKIT to approximate the modification of both surface temperature and dewpoint by the lake (Phillips 1972). The operational Eta (Janjić 1994)/WRF-NMM (Skamarock et al. 2005) numerical data were used for our case studies. As of 2008, the WRF-NMM is run with 12-km horizontal resolution and 60 vertical levels. This modified (through BUFKIT algorithms) model output was analyzed to calculate the lake-induced CAPE, lifted condensation level (LCL), level of free convection (LFC), and EL (see Table 1).

The lake-effect cloud layer was defined as the layer bounded by the lake-induced LCL and EL. A hypothesis, expanded from that of Michimoto (1993), is that the presence of the layer from −10°C to −25°C within the lake-effect cloud layer is a good predictor of CG lightning. This temperature range permits mixed phases of precipitation to occur (graupel, supercooled water, and ice), which is crucial for electrical charge separation. Indeed, Jiusto and Holroyd (1970) observed conical graupel and highly rimed ice crystals in lake-effect snowstorms with lightning. To test this hypothesis, BUFKIT time–height cross sections (see Fig. 8, described below) were used to determine when the layer from −10°C to −25°C was within the convective cloud layer during a lake-effect storm. Three storms were analyzed (28–31 January 2004, 12–13 October 2006, and 3–12 February 2007). The 28–31 January 2004 case was chosen because it was an intense snowstorm that did not produce any CG lightning according to the NLDN and National Weather Service storm spotters (a control). The other two storms were chosen to test our hypothesis on both early- and late-season storms that had lightning.

3. Results

a. Lake-effect lightning climatology

Cloud-to-ground lightning was reported with lake-effect storms on 95 days during the study period, for an average 7.9 days per lake-effect season. The standard deviation was 4.3 days. For simplicity, storm duration was defined as the number of consecutive days with CG lightning in a lake-effect storm. According to this definition, the 95 days were distributed among 70 discrete lake-effect thunderstorm events. This yields an average of 5.8 lake-effect storm events per lake-effect season, with a standard deviation of 2.6 storms. Figure 2 shows the distribution of CG lightning in lake-effect storms across the lower Great Lakes region. The largest flash-density values (>0.26 flashes per square kilometer or >7 flashes in a 25-km² grid box) were over the eastern and southern portions of Lakes Erie and Ontario. Most of the lightning was over the water and did not extend far inland.

Figure 3 shows the monthly distribution of lake-effect thunderstorm (rain/snow) events. Over the 12-yr period, the greatest frequency (18 events) occurred in both October and November (an average of 1.5 events per month in a given year). There were a significant number of events extending into midwinter (10 events occurred in January).

b. Lake-effect thundersnow

The vast majority of the 43 lake-effect thundersnow events examined over Lakes Erie and Ontario occurred early in the cold season. Nearly 75% of the thundersnows occurred in November and December, coinciding with the time of warmest lake surface temperatures (Fig. 4). Conversely, thundersnows were infrequent during the second half of the meteorological winter (beyond 15 January), with less than 10% of all cataloged events taking place during this period (Fig. 5).

The average lake surface temperature was 16% higher for lake-effect snow events that produced lightning than it was for nonlightning events. Greater than 70% of the lake-effect thundersnows examined occurred with lake temperatures at or above 10°C, and only 10% occurred when lake temperatures were below 4°C. Lake Erie typically freezes over by mid-February (Fig. 4) because of the shallowness of the lake. This partially explains the abundance of lightning during early-season events over Lake Erie, but similar results were also observed over Lake Ontario, which remains largely unfrozen throughout the entire winter.

Modified observed and forecast proximity soundings from Buffalo, Jamestown, Rochester, Syracuse, and Watertown, New York, showed that lake-effect thundersnow events were more associated with a warmer lower troposphere. This finding is similar to that of Schultz (1999) for lake-effect snowstorms in northern Utah and western New York. The three parameters used in identifying the warmer lower troposphere were the height of the lake-induced EL and the altitudes of the −10°C and
The lake-induced EL is the height at which a lake-modified parcel will stop accelerating upward and can serve as a proxy for storm cloud top. The \(-10^\circ\)C isotherm level was examined because it has been identified as the most effective location for graupel initiation (Fukuta and Takahashi 1999), whereas the layer from 0 to \(-10^\circ\)C is the most efficient zone for graupel growth through active riming (Staudenmaier 1999). A deeper layer from the surface to \(-10^\circ\)C is important because it contains a higher concentration of supercooled water droplets, which is very favorable for riming on preexisting ice crystals. Graupel, ice crystals, and supercooled water droplets are essential ingredients in separating electrical charge in convective clouds (Takahashi 1978; Deierling et al. 2006).

Based on the Eta/WRF-NMM BUFR lake-modified soundings, the average estimated height of the \(-10^\circ\)C isotherm for lake-effect snowstorms with lightning was 1.2 km AGL and the mean height was 0.7 km for non-lightning events. This difference is statistically significant (at the \(p = 0.05\) significance level) and represents an 82% increase. Lake-effect thundersnow events were associated with a deeper layer whose temperature was greater than \(-10^\circ\)C, which is significant in terms of the microphysics and the development of graupel as noted above. Likewise, lake-effect thundersnow soundings had a deeper layer from the surface to \(-20^\circ\)C by an average of 25% in comparison with soundings associated with events that did not produce lightning (mean \(-20^\circ\)C isotherm heights of \(>3\) vs 2.4 km, respectively).

Lake-effect thundersnow events had a deeper convective cloud layer than those events without lightning. The lake-induced EL, calculated by incorporating the lake surface temperature into the simulated thermal profile with BUFKIT, was used to estimate potential cloud depth. The average modified (lake induced) equilibrium level for lightning events was 37% greater than for nonlightning events (statistically significant at the \(p = 0.05\) level), with elevations typically higher than 3.6 km AGL.

In addition to the thermal profiles of the lake-effect thundersnow events being “warmer” than their non-lightning counterparts, they also were associated with greater instability. The effects of heat and moisture fluxes from the lake surface were incorporated into the
sounding with BUFKIT. We believe this has resulted in a more realistic analysis of stability values than those of Schultz (1999). The mean lake-induced CAPE value for the thundersnow events was 550 J kg\(^{-1}\), approximately 30% greater than the CAPE associated with the events that did not produce lightning.

A small wind shear environment allows the lake-effect convection to become more concentrated. The latent heating within the cloud and the lake-effect band’s convective and mesoscale vertical motions can be better organized, which leads to an intense band with less shear (Niziol et al. 1995). Given a warm and deep enough convective boundary layer, a flow in all of the low levels of the troposphere that is nearly parallel to the long axis of an underlying lake can produce a single plume of heavy lake-effect thundersnow (Niziol et al.).

Nearly all of the thundersnow events found over Lakes Erie and Ontario from 1995 through 2007 were associated with intense single bands of snow (type 1; Niziol et al. 1995). Over Lake Erie, bands of lake-effect thundersnow typically occurred when the prevailing low-layer (e.g., 950–850-hPa mean) wind direction was between 240° and 270°, whereas 270°–290° was most favorable for Lake Ontario (see Fig. 6). These orientations result in the greatest residence time and fetch (distance that air travels over the water) for the convective elements over each lake, respectively, which allows for maximum heat and moisture transfer from the lake surface to the atmosphere and the development of intense single lake-effect bands. According to Fig. 6, the nonlightning events [which were still intense, producing >15.2 cm (6 in.) of snow in 12 h; see section 2b] were associated with wind directions that were more veered (more westerly for Lake Erie and more northwesterly for Lake Ontario) than those associated with the lightning events. These wind directions are more likely to produce weaker single and multiple bands of snow (type 2; Niziol et al. 1995) because they result in less fetch and hence less of a chance for lightning.

The strong low-level convergence that focuses the lake convection into a single band also maximizes the instability and lake flux interactions. This was an important variable that typically discriminated between very unstable lake-effect snow profiles that produced lightning and those that did not. The clustering of NLDN and surface spotter reports of CG lightning flashes at the east ends of both Lakes Erie and Ontario supports the importance of single-banded storms forming parallel to the long lake axis in producing lake-effect thunderstorms. For example, little if any CG lightning was found along the south shore of Lake Ontario in the Rochester metropolitan area, whereas lake-effect thundersnow events were fairly common in the Buffalo metropolitan area and between the Lake Ontario shoreline and the Tug Hill Plateau (see Figs. 1 and 2). The lack of CG flashes more than 45 km inland of the shores of both lakes also points to the importance of heat and moisture...
Our reanalysis of more than 10 yr of lake-effect snowstorms, using archived radar, satellite, model, and snow-spotter data, coupled with the forecast experience of the second author (RH) in the Buffalo National Weather Service Forecast Office (NWSFO), revealed a very strong correlation between the thundersnow events and lake-effect band type 1 as categorized by Niziol et al. (1995).

Another interesting finding was the presence of lightning within 1 h of graupel being reported at the surface. Whereas graupel was not often reported during a lake-effect thundersnow event, lightning and thunder often accompanied the occurrence of graupel. This was yet another benefit provided by the NWS Buffalo snow-spotter network.

c. Case studies

Figure 7 shows the CG flash density for a historic early-season lake-effect snowstorm that affected Buffalo during 12–13 October 2006 (Buffalo NWSFO 2007). In addition to the unusual amount of snowfall for this time of year (maximum report of >60 cm), this lake-effect storm was anomalously electrified (at least 185 CG flashes were reported in Fig. 7). The forecast maximum lake-induced CAPE according to the WRF-NMM BUFKIT (from the cycle initialized at 0000 UTC 13 October using a lake temperature of 17°C) for Buffalo was 2400 J kg\(^{-1}\) at 0200 UTC 13 October; at this time the lake-induced EL was 7.3 km AGL. The layer from \(-10^\circ\) to \(-25^\circ\)C (the favorable zone for graupel and ice crystal production) was within the lake-effect cloud layer from initialization (0000 UTC) until approximately 1600 UTC 13 October (Fig. 8). By 1600 UTC, the lake-effect band had drifted north of Buffalo and weakened. The National Weather Service in Buffalo reported that there was lightning for a period of more than 12 h during this storm.

The 3–12 February 2007 lake-effect snowstorm was not as electrically active as the historic 12–13 October 2006 storm. However, several flashes were witnessed by students and faculty at the State University of New York at Oswego (SUNY Oswego). Because of limitations in archived BUFKIT data, we focused on the 0000 UTC 5 February WRF-NMM run for a central Lake Ontario location (43.65°N, 77.38°W) using a lake temperature of 3.3°C. Lake-induced CAPE reached a maximum of 1790 J kg\(^{-1}\) at 0500 UTC 5 February, with a lake-induced EL of 5.3 km AGL (Table 1); however, at that time the layer from \(-10^\circ\) to \(-25^\circ\)C was not within the convective cloud layer. From 1200 to 1500 UTC on 5 February, the layer from \(-10^\circ\) to \(-25^\circ\)C was within the convective cloud layer; it was during this time that lightning was observed on the SUNY Oswego campus (see Fig. 1 for location).
For the null case, 28–31 January 2004, we used the WRF-NMM run for the central Lake Ontario location initialized at 1200 UTC 29 January. The lake temperature was 2.8°C. Forecast lake-induced CAPE peaked at 845 J kg\(^{-1}\) at 1300 UTC 29 January, with a lake-induced EL of 3.7 km AGL (see Table 1). The layer from −10°C to −25°C was within the lake-effect cloud layer from 1200 to 1500 UTC 29 January; however, no lightning was observed during this time by the NLDN and NWS weather spotters.

4. Discussion

a. Climatology

The majority of the CG lightning in lake-effect storms, shown in Fig. 2, occurred either in midlake or along the southern and eastern shores of the lakes. Moore and Orville (1990) show similar results. The majority of CG lightning over land occurs within 20 km of the shoreline. The occurrence of fewer lightning flashes beyond 20 km could be due to the lake-effect clouds becoming glaciated as they move farther inland. We hypothesize that as a lake-effect band’s clouds move away from the lake, which is the band’s heat and moisture source, the production of supercooled water droplets decreases and a greater percentage of the hydrometeors become ice crystals as convective potential and updraft strength weakens. Once the production of supercooled water droplets decreases considerably, graupel concentrations decrease. As discussed earlier, graupel is believed to be necessary for charge separation and lightning in clouds (Reynolds et al. 1957; Takahashi 1984).

Although more than one-half of the storm events (51%; see Fig. 3) occurred in October and November, approximately 27% of all lake-effect storms with lightning occurred in the January–March period. This result is significant, considering that Jiusto and Holroyd (1970) concluded few lake-effect storms with CG lightning occur after December. Moore and Orville (1990) concluded that CG lightning only occurs in lake-effect storms during the September–December period. Storm duration (Fig. 9) was limited to 1 day for the vast majority of events, but some storms lasted 2–3 days and one storm (29 September–2 October 2003) lasted 4 days.

In a typical case, only a few CG flashes were recorded per storm, which agrees with the results of previous
studies (e.g., Moore and Orville 1990). The total flash rate (CG, cloud to cloud, and intracloud) can be significantly larger. This was observed on 2 December 2005 in Oswego (NWS Buffalo Snow Spotter 0604 2006, personal communication) when approximately 20 flashes were observed while the NLDN indicated only 5 CG flashes near the area (within 30 km of Oswego). Some storms had a significantly larger number of CG flashes. For example, according to the NLDN, the 29 September–2 October 2003 storm produced 1551 CG flashes.

b. Case studies and forecasting implications

A layer from $-10^\circ$ to $-25^\circ$C within the lake-effect convective cloud layer appears to be a necessary, but not sufficient, condition for lightning occurrence with lake-effect storms, based on the three cases shown (and others not discussed in this paper). We speculate that the lower instability (shown by lake-induced CAPE) present during the 28–31 January 2004 storm was not sufficient for enough charge separation to initiate lightning, despite the presence of a mixed-phase cloud. Although Schultz (1999) concluded that there is little or no CAPE observed in lake-effect storms, BUFKIT software allows the user to input lake surface temperature and modify model data to account for lake-induced CAPE to allow an approximation of the thermodynamic conditions near and within snowbands forming over a lake.

The lake-effect thundersnow events examined show a clear relationship with deep warmer lower-tropospheric layers. These results are similar to Schultz’s (1999) findings, but advances in thermal-profile diagnostic software (e.g., BUFKIT) enabled a more realistic examination in this study of the lower boundary layer and its interaction with the underlying lake surface. The warmer lakewater of the early winter season helps to deepen the favorable layer from $0^\circ$ to $-10^\circ$C by adding heat and moisture to the lowest levels of the boundary layer. A warmer lower troposphere can result in a greater abundance of graupel, which aids in the separation of electrical charge.

The results of this study confirm that most of the thundersnow events occur during November and December when the lake surfaces are at their warmest of the cold season. As stated by Kristovich and Laird (1998), the magnitudes of heat and moisture fluxes from the surface of the lakes (these fluxes are greater early in the season) greatly influence the growth rate of the convective boundary layer. The warmer profiles also correspond with higher $-20^\circ$C and lake-induced equilibrium levels. These are both indicators of a deeper

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FIG. 8. Time–height model BUFKIT cross section for a location over eastern Lake Erie (time increases from 0000 UTC 13 Oct on the right to 1200 UTC 15 Oct 2006 on the left). Isotherms (°C, thin black), lake-induced LCL (thick black), and lake-induced EL (thick gray) are plotted with pressure (hPa) as the vertical coordinate. The cloud layer is defined as being between the LCL and EL.
mixed-phase cloud region, which can allow for sufficient charge separation to occur.

The lake-effect snowstorms that did not produce any lightning were 37% shallower than their counterparts. Lightning events were associated with lake-induced equilibrium levels in excess of 3.6 km AGL. These observations are supported by Niziol (1987), who showed that lightning and thunder accompanied heavy lake-effect snowfall occurring with inversion heights of greater than 3 km.

With respect to atmospheric instability, modified thermal profiles show that lake-effect thundersnow events are more likely when the lake-induced CAPE values are in excess of 500 J kg$^{-1}$. These values are approximately 30% larger than for lake-effect snow events without lightning. Nearly 100% of the events with lake-induced CAPE values in excess of 1000 J kg$^{-1}$ produced lightning. Hence, significant convective vertical motion in the presence of a mixed-phase lake-effect cloud is integral to separating electrical charge to the point of initiating lightning. Lake-induced CAPE was not used in earlier studies (e.g., Schultz 1999) because of the inability to incorporate the effects of heat and moisture fluxes from the lake surface into observational and model sounding data (e.g., soundings were taken over land).

A nomogram was developed based on our findings (Fig. 10) that was experimentally used to forecast the occurrence of lightning within lake-effect snowstorms during the 2006/07 winter season. The nomogram uses two main variables: the height of the −10°C level as the x axis and the height of the lake-induced EL as the y axis. The height of the −10°C level was used to represent the availability of graupel sufficient for separating electrical charge and producing lightning within the mixed-phase region of the cloud. The height of the lake-induced EL was found to best represent convective cloud depth that can allow for charge separation. Given a single plume of lake-effect snow, these two variables were found to work best in a basic two-element nomogram for the prediction of thundersnow. Various other parameters were examined during the development of the nomogram, but none were able to make as clear of a distinction between events that contained lightning and those that did not as do the parameters used in Fig. 10. These other parameters included CAPE, depth of the convective cloud, depth of the mixed-phase region (the layer from −10°C to −25°C) within the cloud, shear, and wind direction (fetch length). The use of mixing ratio or surface dewpoint temperature may have proven to be more useful, but reliable values for these parameters over the lake surfaces were not available in our data.

The nomogram was tested in the Buffalo National Weather Service Forecast Office during the winter of 2006/07 with favorable results (Fig. 11). Four out of five lake-effect events that were in the “likely” portion of the nomogram produced lightning, and two out of nine events in the “chance” section contained lightning. All of the remaining events in the “not expected” portion of the nomogram occurred without producing any reported (spotter and NLDN) lightning.

The wide range of values associated with lightning events within the chance section of the nomogram shows that a greater understanding is needed about the development of lightning within lake-effect snowbands. The future implementation of total lightning detection systems [ground and satellite based, e.g., the planned...
Geostationary Operational Environmental Satellite-R (GOES-R) program and improvements in numerical modeling (e.g., higher resolution and improved microphysics schemes) will lead to better forecast methods.

5. Conclusions

Cloud-to-ground lightning occurs in intense lake-effect storms (rain and snow) during the autumn and winter seasons across the lower Great Lakes. The majority of these storms occurred in the September–December period; however, 27% of observed lightning events in the study period occurred in January, February, or March. Seventy-five percent of the thundersnow events occurred in November and December over the region. An average of about 8 days with CG lightning in lake-effect storms and 5.8 storm events (an event’s duration was defined as the number of consecutive days with CG lightning in lake-effect storms) occur during the lake-effect season. These storms often last only 1 day and have only a few CG flashes. The longest storm lasted 4 days, and the most intense (defined by the number of CG flashes indicated by the NLDN) storm consisted of 1551 CG flashes. Updates to this study are recommended in future years, after data from more lake-effect seasons have been collected. In addition, lake-effect thunderstorms may allow scientists to determine the minimum requirements for lightning to occur in storms because conditions are marginal for significant charge separation in these clouds (i.e., most lake-effect storms do not produce lightning because they represent weak convection relative to warm season thunderstorms).

Mobile upper-air facilities were recently acquired by the faculty at SUNY Oswego through a National Science Foundation grant. Our objectives with this equipment include model verification of lake-effect snowstorms and their environments and the testing of some of our hypotheses for lake-effect thunderstorm development (e.g., the presence of the layer from \(-10^\circ\) to \(-25^\circ\)C in the cloud). The mobile facilities will be important, because Fig. 2 shows that the majority of the lightning occurs near the shoreline, a location not well sampled by the nearby fixed sounding sites at Buffalo and Albany, New York.

While forecasting lake-effect snow in western New York for more than 10 years, the second author (RH) has personally experienced that the large majority of lake-effect lightning has been intracloud. This personal experience has been supported by comparing NWS spotter reports with the information received from the NLDN. There are frequent reports of lightning with no corresponding data from the NLDN. Because the NLDN is designed to detect mostly CG lightning, it is likely missing the majority of activity within lake-effect snowbands. Total lightning detection networks [e.g., the Lightning Mapping Array (LMA); Krehbiel et al. (2000)] are necessary to capture the intracloud lightning and charge structure in these clouds. Volunteer snow-spotter reports that are not limited to observing just CG flashes like the NLDN will continue to assist NWS meteorologists in forecasting lake-effect thunderstorms until a system like the LMA is deployed in the area.

Future research also entails examining the radar reflectivity structure of these storms and determining whether there are any thresholds in the data that are indicative of charge separation and lightning (e.g., presence of values of >35 dBZ).

BUFKIT software is a useful tool in forecasting lightning in lake-effect storms. The presence of the layer from \(-10^\circ\) to \(-25^\circ\)C within the predicted lake-effect cloud layer (lake-induced LCL to EL analyzed within BUFKIT) appears to be a necessary, but not sufficient, condition for CG lightning to occur; values well over 500 J kg\(^{-1}\) of lake-induced CAPE (also calculated by the BUFKIT software) must also be present for CG lightning initiation. In agreement with the results of Michimoto (1993), greater heights of the \(-10^\circ\)C isotherm (>1 km AGL) are also conducive to lake-effect lightning because this condition allows more graupel to form in the warmer cloud. Some other environmental characteristics that favored electrified lake-effect storms include a higher equilibrium level (>3.6 km) and hence greater cloud depth, small wind shear values, and single lake-effect bands as opposed to multiple-band events. More case studies are recommended to test the statistical significance of these hypotheses.

A nomogram was developed using the heights of the lake-induced equilibrium level and \(-10^\circ\)C level to predict lake-effect thundersnowstorms (Fig. 10). The
nomogram was used successfully in the Buffalo NWSFO to forecast the occurrence of these storms during the 2006/07 winter season.

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