Analysis of a Surprise Western New York Snowstorm

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ABSTRACT

Although Rochester, New York (ROC), is not located in a climatologically favored region for extreme [i.e., $\geq 30$ cm (12 in.) 24 h$^{-1}$] lake-effect snow (LES), significant [i.e., $\geq 15$ cm (6 in.) 24 h$^{-1}$] LES can occur there under specific synoptic regimes. The purposes of this study are to document synoptic conditions that are associated with significant LES in ROC and to examine a specific event in which the passage of an upper disturbance combined with a lower-tropospheric trough to produce a surprise western New York snowstorm on 26–27 November 1996.

A database of 127 events in which 2-day ROC snowfall exceeded 15 cm (6 in.) was constructed for the years 1963 through 1992, inclusive. Each event was categorized as “LES” or “non-LES” on the basis of air–lake temperature difference, wind direction, and synoptic setting. Of the 127 events, 32 were classified as LES. Composites based on this 32-case sample reveal a mobile upper trough that moves from the western Great Lakes 48 h prior to the snowfall event to northern Maine 24 h after the event. All 32 cases were accompanied by either a mobile upper trough or a closed low at the 500-hPa level.

An unexpected snowstorm on 26–27 November 1996 resulted in accumulations of up to 30 cm (12 in.) in parts of western New York. Nonclassical LES structures developed in a rapidly changing synoptic environment that was characterized by the passage of an intense upper-tropospheric disturbance. Model forecasts underestimated the strength of this disturbance and also the intensity of lower-tropospheric troughing over and north of Lake Ontario. The upper trough is hypothesized to have increased the inversion altitude and relative humidity in the lower troposphere, and likely contributed to the strength of lower-tropospheric troughing near Lake Ontario. Cyclonic isobaric curvature accompanying the surface trough enhanced lower-tropospheric ascent through Ekman pumping and increased the overwater fetch for boundary layer air parcels traversing Lake Ontario. Comparison of Eta Model forecasts with analyses suggests that problems with model initialization and diabatic boundary layer processes both contributed to forecast errors.

1. Introduction

Wintertime forecasting in the vicinity of the Great Lakes is complicated by interactions between synoptic-scale disturbances and mesoscale circulations that develop as cold air passes over relatively warm lake surfaces. Destabilization of arctic air by the warm underlying surface can lead to convective bands that produce highly localized snowfalls; the convective nature of the bands can allow significant [i.e., $\geq 15$ cm (6 in.) 24 h$^{-1}$] snow accumulation even with small liquid-water equivalents (e.g., Remick 1942; Jiusto et al. 1970). Forecasting difficulties are exacerbated by the sensitivity of snowband character to subtle changes in environmental parameters (Niziol 1987; Hjelmfelt 1990). Such complications led Dole (1928) to comment that lake-effect precipitation is “most exasperating to forecasters as well as to the public.” An area of high population density that is occasionally affected by lake-effect snow (LES) is found in western New York; LES represents a leading cold-season forecasting problem in the cities of Buffalo, Rochester, and Syracuse, New York (e.g., Niziol et al. 1995).

Variations in shoreline shape, lee topography, extent of ice cover, and in the orientation of prevailing winds relative to lake axes result in location-specific LES behavior (e.g., Eichenlaub 1970). For example, synoptic conditions that are conducive to LES in Rochester, New York (ROC), differ from those associated with LES in nearby Buffalo or Syracuse (Weinbeck 1983; Ellis and Leathers 1996). As noted by Niziol (1987) and others, the prevailing wind direction favorable for heavy LES at Buffalo would not support LES in ROC, for example.
Rochester does not lie in a climatologically favorable location for extreme [i.e., ≳30 cm (12 in.) 24 h−1] LES due to the lack of upstream overwater fetch during many arctic air outbreaks (Fig. 1). Ellis and Leathers (1996) identified five synoptic patterns that are associated with LES in western New York and Pennsylvania. Spatial plots of snowfall amount and frequency accompanying each of the five patterns reveal that ROC is located in a climatological minimum for LES relative to the remainder of the region (see their Figs. 3–7). However, significant LES is occasionally observed at ROC, and differences in synoptic conditions that distinguish these events from those that produce a light dusting can be subtle.

Niziol et al. (1995) present a classification scheme depicting five different types of lake-effect snow structure. In their scheme, type I bands arise when the prevailing flow is aligned with the major axis of a lake, resulting in an intense midlake band that is 20–50 km in width. The location of Rochester is protected from type I bands over Lake Ontario, but in certain situations type I bands can extend from Lake Erie into the vicinity of ROC. Type II bands are also wind parallel, but develop when the prevailing flow is perpendicular to the major axis of the lake and are typically 5–20 km in width. Bands of this type frequently affect the southern shore of Lake Ontario, including the Rochester area, but rarely produce heavy accumulations. On occasion, multiple type II bands may combine into a single band along the downwind lakeshore. Type III bands are wind-parallel bands that involve more than one lake. For instance, when a type I band sets up over Lake Huron, it may extend across Ontario and reintensify over Lake Ontario. The upstream presence of Lake Huron and Georgian Bay can result in significant moistening and destabilization of the air mass upstream of Lake Ontario (Byrd et al. 1992; Niziol et al. 1995); type III bands originating over these water bodies have been observed to affect ROC and vicinity (e.g., Byrd et al. 1992). Type IV bands are shore-parallel bands that set up in response to a land-breeze circulation during weak synoptic flow, such as when an arctic high pressure system resides over the area. Snowfall associated with type IV bands is usually limited in intensity by the presence of the subsidence inversion that typically accompanies arctic high pressure systems. Snowfall, which can be substantial during persistent type IV bands, is spatially restricted to within a few miles of the shoreline. Type V bands are actually mesoscale vortices that develop during weak synoptic forcing. They are often confined to the offshore waters and seldom affect the Rochester area.

Due to the mesoscale nature of LES, the density of observations in the traditional synoptic data network is insufficient to resolve many important LES features. Even early field projects designed to describe the physical processes acting in LES (e.g., Falconer et al. 1964; Peace and Sykes 1966; Sykes 1966) were limited by insufficient observational data resolution (Peace and Sykes 1966). More recently, studies utilizing Doppler radar, satellite imagery, and instrumented aircraft have documented the structure and dynamics of LES systems (e.g., Passarelli and Braham 1981; Schoenberger 1983; Chang and Braham 1991; Reinking et al. 1993; Kristovich et al. 2000).

With the advent of high-speed computing, meteorologists have increasingly turned to numerical models to provide the requisite spatial and temporal resolution for LES analysis (e.g., Lavoie 1972; Boudra 1981; Ballentine 1982; Hjelmfelt and Braham 1983; Ballentine et al. 1998). Concomitant with computational advances, the sophistication and accuracy of operational LES forecasting has likewise increased (Dewey 1979a,b; Dockus 1985; Niziol 1987; Niziol et al. 1995; Waldstreicher et al. 1996; Mahoney and Niziol 1997). Recent high-resolution numerical simulations have proven capable of resolving many of the mesoscale characteristics of LES bands (e.g., Hjelmfelt 1990; Bates et al. 1993), and mesoscale model simulations are now available in real time to some operational forecasters (Ballentine et al. 1998; Waldstreicher et al. 1998). The role of mesoscale model output in the LES forecast process continues to evolve as operational forecasters gain experience with this new tool. Although the National Centers for Environmental Prediction operational models such as the Eta Model (e.g., Rogers et al. 1996) do not currently resolve individual LES bands, forecasters still rely heavily on

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For context, the 30-yr average annual snowfall at ROC for the period 1967–96 is 244 cm (96.2 in.), with a standard deviation of 66 cm (26.6 in.). On average, there are approximately 66 days per season with measurable snow. This implies that there are a relatively large number of days with light snowfall reported.
these models to accurately predict the synoptic-scale setting for LES events (Niziol et al. 1995).

During the afternoon of 26 November 1996, light precipitation in advance of an upper-tropospheric disturbance overspread much of southern Ontario and parts of western New York State. Forecasts called for a dusting to an inch of snow accumulation. However, shortly after nightfall, the snowfall increased from light to moderate in intensity, and by 0322 UTC 27 November, heavy snow was reported at ROC and the surrounding areas. Snow continued to fall over the southern and central portions of New York during the morning of 27 November. Morning commuters were faced with up to 30 cm (12 in.) of accumulated snowfall in the Rochester area. The goals of this paper are to (i) examine the mechanisms responsible for this storm within the context of a cursory climatology of historical Rochester LES events and previous LES classification studies, and (ii) conduct a preliminary investigation into the causes of synoptic-scale errors in the operational Eta Model for the 26–27 November 1996 event. Following an outline of data and methodology in section 2, a brief climatological summary of LES in the Rochester area is contained in section 3. Analysis of synoptic and mesoscale aspects of the November 1996 snowstorm is presented in section 4, followed in section 5 by an evaluation of operational Eta Model forecasts for that event. Conclusions and suggestions for future research are provided in section 6.

2. Data and methodology

In order to provide context for the November 1996 event, a sample of historical ROC LES cases was identified. The methodology for this cursory climatology is presented below, followed by a description of data sources and analysis techniques for the November 1996 case study.

a. Background climatology

Daily precipitation and temperature data were obtained from the National Climatic Data Center (NCDC) dataset TD-3220, stored on compact disc (CD). This dataset was used to identify all events from January 1963 through December 1992 in which 1 or 2-day snowfall accumulations exceeded 15 cm (6 in.) at ROC. This search yielded 127 events. The 2-day accumulation period allows the inclusion of 62 events that spanned consecutive days with 1-day totals of less than 15 cm (6 in.). Events in which lake-induced convection was primarily responsible for the snowfall were categorized as ‘‘LES,’’ and all other events were dubbed ‘‘non-LES.’’ The criteria used in this categorization are discussed in detail below.

Rothrock (1969) and Holroyd (1971) identified conditions that are conducive to LES, including a threshold temperature difference between the lake surface and 850-hPa level of 13°C. This value is often cited operationally as a threshold value for LES occurrence (e.g., Niziol 1987, section 3) and was thus utilized in the identification of ‘‘LES candidate’’ events from the original 127-case sample. Monthly climatological lake surface temperature distributions for Lake Ontario were taken from Webb (1970), and linearly interpolated in time and spatially averaged to form single biweekly climatological values for the Lake Ontario temperature. Next, 850-hPa temperature values were obtained from Buffalo, New York (BUF), rawinsonde profiles, and lake surface to 850 hPa temperature differences were computed. Cases in which at least one of the synoptic times during the event exhibited a temperature difference ≥13°C were identified as LES candidates.

Next, ROC LES events were restricted to include only those cases where the lower-tropospheric wind direction (as determined via examination of the BUF wind profile between the surface and inversion base) was northwesterly for at least one of the times that met the temperature difference criterion. Northeasterly flow can produce ‘‘lake enhanced’’ snowfall during the passage of a cyclone to the south of Lake Ontario if the air mass is sufficiently cold [e.g., Eichenlaub (1970) discusses this phenomenon for Lake Michigan]. However, for Lake Ontario, ‘‘pure’’ LES with northeasterly flow in the absence of a synoptic-scale cyclone, though not unheard of [an example occurred at ROC on 14 February 1998 which resulted in 7.1 cm (2.8 in.) of snowfall], is unusual and rarely significant. This is due in part to the low inversion that generally accompanies anticyclonic northeasterly flow in these situations. Accumulating snowfall from type I Lake Erie bands can reach ROC during events with strong southwesterly winds in the boundary layer and high inversion altitudes, but these events are also unusual. Therefore only northwesterly flow cases were considered as ROC LES events here.

Gridded analyses of sea level pressure, 850-hPa geopotential height and temperature, and 500-hPa geopotential height and geostrophic relative vorticity ($\zeta_g$) were generated from the NCEP CD, which contains an archive of the ‘‘octagonal grid’’ analyses with horizontal resolution of 381 km at 60°N (Mass et al. 1987). Grids read from the CD were ingested into the General Meteorological Package (Koch et al. 1983) for storage and display. Sequences of analyses were subjectively examined in order to determine if synoptic-scale cyclones (as defined by at least two closed 4-hPa sea level pressure contours in the NCEP analysis) were located within 250 km of ROC during the period of snowfall. Candidate events that were not accompanied by a cyclone were then classified as LES.

Of the 127 original ROC snow events, 47 were identified as LES candidates. Of these, 15 were eliminated due to the presence of synoptic-scale cyclones, leaving 32 ‘‘true’’ LES events with a monthly breakdown as follows: November, 1; December, 10; January, 9; Feb-
ditionally used in weather forecasting, the 500-hPa geodynamics (e.g., Morgan and Nielsen-Gammon 1998). In a compact means of summarizing upper-tropospheric data is stored on a Lambert conformal grid projection with the 26±27 November 1996 event from the Unidata local data manager, at the State University of New York the operational Eta Model, at the 500-hPa geopotential height field indicates a ridge over the west coast of North America, with a trough over the central United States (Fig. 2a). An arctic anticyclone is evident in the sea level pressure field over the Canadian Northwest Territories, Alberta, and Saskatchewan and a weak trough resides over the Great Lakes. By the arctic high pressure system has intensified and moved southward (Fig. 2b). A surface cyclone is centered over southern Ontario ahead of the advancing trough at the 500-hPa level. A cyclonic geostrophic vorticity maximum is centered over Michigan at this time. At the surface cyclone has intensified and moved to the coast of Maine (Fig. 2c). A strong gradient in the sea level pressure field exists between this cyclone and the arctic high to the west, and a trough extends from northern Michigan eastward to southern Ontario. At the 500-hPa level a closed low is centered over southern Quebec (Fig. 2c). By the arctic anticyclone has weakened and moved southeast while the Maine low has intensified and moved northeast into the Canadian Maritime Provinces (Fig. 2d).

Operational forecasters have long recognized that a slow-moving closed low at the 500-hPa level centered near James Bay often accompanies prolonged LES in the eastern Great Lakes (Niziol 1987). This composite exhibits a mobile upper trough passing over the eastern Great Lakes and, by the arctic high over southern Quebec. The appearance of a mobile upper-tropospheric vorticity maximum in the composite is significant. Such disturbances act to lift the base of the inversion layer typically found in arctic air masses, and are associated with ascent and cyclonic flow curvature on the synoptic scale. Such an environment is more conducive to the development of lake-induced convection than would otherwise be the case (e.g., Niziol 1987). Inspection of the 500-hPa geopotential height and geostrophic relative vorticity fields for each of the 32 individual LES cases indicates that all of these events were accompanied by either a mobile trough or a closed low at the 500-hPa level. In light of the fact that each case was characterized by a synoptic-scale upper disturbance, it can be argued that these are not pure LES events. However in situations where an upper trough is moving over a cold, stable arctic air mass, there is often little in the way of accumulating snow until the disturbance passes over the Great Lakes. Thus, it appears that upper-tropospheric disturbances are a ubiquitous feature accompanying significant ROC LES.

3 The dynamic tropopause is defined here as the first encounter with the 1.5 potential vorticity unit (Hoskins et al. 1985) surface during a downward search beginning at the 100-hPa level.

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Troughing in the composite sea level pressure field over the Great Lakes between the cyclone to the east and arctic anticyclone to the west is similar to that documented by Petterssen and Calabrese (1959), who determined that the upper limit to lake-induced trough intensity was approximately 6 hPa for the Great Lakes. The troughing due to the combined effects of the Great Lakes has been investigated numerically by Sousounis and Fritsch (1994), and dubbed the “lake aggregate” trough. Such troughing can act to increase the westward deflection of boundary layer air parcels moving south from Canada, and may allow airflow to cross Lake Huron or Georgian Bay prior to reaching Lake Ontario. As noted previously, significant LES at ROC can develop in association with type III snowbands when air flows over Georgian Bay, across southern Ontario, and then across Lake Ontario, thereby effectively increasing the overwater fetch beyond what Lake Ontario alone could provide.

The composite sea level pressure field for $\tau_0$ presented in Fig. 2c is broadly similar to the WNW–S, W–S, and NW–S patterns of Ellis and Leathers (1996, their Figs. 3, 5, and 7). Each of these patterns is characterized by a departing cyclonic system to the northeast and an arctic anticyclone to the west and southwest. The spatial snowfall frequency diagrams presented for each of these synoptic types (Ellis and Leathers 1996; Figs. 3c, 5c, and 7c) indicate that snowfall in the vicinity of ROC was relatively light and infrequent for all of these patterns, with the WNW–S pattern producing the most frequent snowfall at ROC (approximately 30%). The results of the current study suggest that the presence of a mobile upper trough may be a distinguishing characteristic for significant ROC LES.

4. Analysis of the 26–27 November 1996 event

During the evening of 26 November 1996, an intense upper trough accompanied by lake-induced convection produced locally heavy snow accumulations south of
Aware of an intense upper trough passing over the region, NWS forecasters posted winter weather advisories in advance of the event. Numerical forecast guidance predicted a straight northerly flow over Lake Ontario; thus, it appeared that the overwater fetch would be insufficient to support heavy LES in western New York. In light of the synoptic-scale guidance, no winter storm warnings or watches were posted until after the event was under way (NWS Buffalo 1997, personal communication). This section contains an analysis of this event, followed by an investigation of potential sources of model errors for this case in section 5.

**a. Synoptic-scale environment**

The Eta-analyzed $p_{trop}$ and $V_{trop}$ fields for 1200 UTC 26 November 1996 indicate an upper trough extending from northwestern Quebec to Wisconsin (Fig. 4a). Tropopause pressure is greater than 500 hPa along the northern portion of the trough axis, with strong cyclonic curvature evident in the $V_{trop}$ field. The region of larger $p_{trop}$ coincides directly with a 500-hPa absolute vorticity ($\xi_a$) maximum (Fig. 4b). A surface cyclone is centered over northern Virginia, and an arctic anticyclone extends from the Canadian Prairie Provinces southward to Oklahoma (Fig. 4c). A complex pattern of troughing associated with cold air flowing over relatively warm water is evident over Lakes Superior and Huron, reminiscent of the case documented by Petterssen and Cabarese (1959).

At 0000 UTC 27 November, the portion of the upper trough that was previously centered over the northern Great Lakes has moved east-southeastward and has evidently intensified (Figs. 4a and 5a). Pressure on the dynamic tropopause near the center of the trough has increased to over 650 hPa at this time, while the 500-hPa $\xi_a$ is analyzed in excess of $32 \times 10^{-5}$ s$^{-1}$ there (Figs. 5a,b). The surface cyclone has moved northeastward toward the Canadian Maritime Provinces (Fig. 5a).

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**Table 1.** Hourly METAR and selected SPECI observations from Rochester, NY, from 2354 UTC 26 Nov through 1154 UTC 27 Nov 1996. Reports between 0554 and 1154 UTC are omitted for brevity.

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>Sky</th>
<th>VIS (mi)</th>
<th>WEA</th>
<th>SLP (mb)</th>
<th>$T$ (°C)</th>
<th>DP (°C)</th>
<th>WDIR (%)</th>
<th>WSPD (kt)</th>
<th>GST (kt)</th>
<th>6-h max (°C)</th>
<th>6-h min (°C)</th>
<th>P3/6 (in.)</th>
<th>SD (in.)</th>
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<tbody>
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<td>OVC</td>
<td>8</td>
<td>OVC</td>
<td>1019.9</td>
<td>-2.2</td>
<td>-2.8</td>
<td>240</td>
<td>5</td>
<td>-0.6</td>
<td>-2.8</td>
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<td>KROC 0054</td>
<td>OVC</td>
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<td>-SN</td>
<td>1020.2</td>
<td>-2.2</td>
<td>-2.2</td>
<td>250</td>
<td>7</td>
<td>270</td>
<td>8</td>
<td>0.05</td>
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<td>270</td>
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<td>270</td>
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<td>0.04</td>
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</tr>
</tbody>
</table>

Legend: VIS (visibility), WEA (current weather), SLP (sea level pressure), $T$ (temperature), DP (dewpoint temperature), WDIR (wind direction), WSPD (wind speed), GST (wind gust), Max (maximum temperature), Min (minimum temperature), P3/6 (3- or 6-h precipitation amounts), and SD (snow depth). Six-hourly precipitation is given at 2354, 0554, and 1154 UTC.
The arctic anticyclone has intensified, with central sea level pressure values in excess of 1042 hPa over South Dakota (Fig. 5b). The sea level pressure pattern is complex in the vicinity of the Great Lakes, with troughing evident over Lakes Ontario, Huron, and Superior. The troughing over the individual lakes is likely forced by diabatic processes in the boundary layer; the trough over Lake Ontario may be additionally influenced by the nearby upper-level disturbance. The sea level pressure pattern at this time is broadly similar to the $t_0$ composite—a surface cyclone is located over the Canadian Maritimes with an arctic anticyclone in the Midwest, and complex lake-induced troughs located between these two pressure centers (cf. Figs. 2c and 5c).

The Buffalo sounding from 0000 UTC 27 November indicates a relatively deep arctic air mass, with an inversion located near the 700-hPa level (Fig. 6). This inversion altitude is relatively high for an LES event (the mean inversion pressure for the 32-case climatological sample of ROC LES events was approximately 825 hPa); this allowed sufficiently deep convective motions in the unstable air mass below the inversion to produce heavy snow. It is likely that some lifting of the inversion occurred locally in association with the convection, as was observed in the case documented by Byrd et al. (1991). The moist air below the inversion and high inversion base are also consistent with synoptic-scale ascent associated with the passage of the upper trough. The temperature of Lake Ontario was reported as 7.2°C in the ROC observation of 1400 UTC 27 November. The 850-hPa temperature measured at Buffalo was $-10.5\,^\circ\text{C}$ at 0000 UTC 27 November and $-16.5\,^\circ\text{C}$ at 1200 UTC 27 November. The corresponding lake surface to 850 hPa temperature differences of 17.7°C and 24.3°C are indicative of moderate instability and are consistent with the development of convection along the southern shore of Lake Ontario. Winds below the inversion indicate slight vertical shear (Fig. 6), but less than the 30° indicated by Niziol (1987) to be disruptive to LES.

b. Mesoscale troughing and precipitation bands

Diabatically forced troughing in the vicinity of the Great Lakes has been studied within the context of modifications to preexisting synoptic-scale cyclones (e.g.,
Troughing is evident over each of the Great Lakes in the Eta analysis of 0000 UTC 27 November (Fig. 5c), and it is likely that a collective mesoscale aggregate lake vortex is present. In addition, a regional sea level pressure analysis for the eastern Great Lakes at 0000 UTC 27 November indicates troughing over Lake Ontario (Fig. 7). This is supported by the observed westerly wind at ROC, which is consistent with a land breeze circulation. By 0300 UTC, winds at ROC are still westerly (Table 1), but by 0400 UTC the wind has shifted abruptly to a northerly direction with gusts to 21 kt and heavy snow, suggestive of a trough passage (Table 1 and Fig. 8). Examination of the base reflectivity from the Binghamton, New York (BGM), Weather Surveillance Radar-1988 Doppler (WSR-88D) at 0300 UTC 27 November reveals a band of heavier precipitation in a north–south-oriented band extending from the southern shore of Lake Ontario southward into northern Pennsylvania (Fig. 9a). A secondary east–west-oriented band of enhanced reflectivity is evident along the southern shore of Lake Ontario in a location consistent with the axis of the aforementioned trough. This band extends farther west along the lakeshore than is evident from

Danard and Rao 1972; Danard and McMillan 1974; Angel and Isard 1999) and as an independent consequence of cold air passing over a locally warmer surface (e.g., Petterssen and Calabrese 1959; Sousounis and Fritsch 1994; Sousounis 1998). Sousounis (1998) noted that the collective lake-aggregate troughing includes nonlinear effects; the perturbation velocity associated with the lake-aggregate cyclone advects the synoptic-scale temperature field, leading to nonzero height tendencies. A consistent but alternate view, in keeping with potential vorticity (PV) concepts, is to consider that a warm thermal anomaly on the lower boundary is dynamically equivalent to a cyclonic PV anomaly (Bretherton 1966). Thus, the Great Lakes can be viewed as a set of surface-based cycloic PV anomalies that exert a collective negative tendency on the lower-tropospheric geopotential height field in their vicinity. The flow field associated with this boundary PV feature could then advect the background temperature field, resulting in secondary height change features via quasigeostrophic processes. Using numerical sensitivity experiments, Sousounis and Mann (2000) have shown that the spatial distribution of precipitation is sensitive to the lake aggregate.
Fig. 9a because the BGM radar overshoots the top of the band west of ROC. Immediately south of the shore-parallel band, ROC is reporting a westerly wind with light snow at this time (Fig. 9a), but by 0400 UTC the wind has shifted to northerly with heavy snow reported (Fig. 9b). These images indicate that the passage of the band of heavy snow at ROC is coincident with the wind shift and trough passage.

Although it is likely that diabatic processes played a significant role in the formation of the Lake Ontario trough, diabatic forcing cannot explain the southward propagation of this feature after 0300 UTC. Because Lake Ontario is located east of the centroid of the lake-aggregate trough, the advective tendency of the lake aggregate would be to shift the trough axis to the north, opposite the observed movement. Therefore, it is suggested that the trough movement was due to an increasing synoptic-scale northerly flow component as high pressure built to the west of the mobile upper trough. Additional high-resolution numerical model simulations and PV inversion diagnostics will be necessary to test this speculation.

The radar imagery presented in Fig. 9 suggests a superposition of two axes of heavier precipitation; additional information into the structure of these bands is provided through examination of Geostationary Operational Environmental Satellite-8 (GOES-8) enhanced infrared satellite imagery. At 0000 UTC 27 November, an axis of locally colder cloud tops extends from southern Quebec southwestward into western New York State (Fig. 10a). This cloud feature is likely associated with

the upper trough depicted in Figs. 5a and 5b. By 0200 UTC 27 November, convection has developed in a localized band immediately upwind of the southern shore of Lake Ontario (not shown), and at 0300 UTC, this band has moved onshore and intensified, with estimated cloud-top temperatures as low as $-29^\circ$C over eastern Monroe County (Fig. 10b). Consistent with the radar imagery presented in Fig. 9a, the signature of the north-south-oriented band is also evident in the satellite imagery (Fig. 10b).

The heavy snowfall during this event was initiated along the south shore of Lake Ontario as an intense shore-parallel band (Fig. 10b), in association with an additional wind-parallel band extending from Lake Ontario southward to near the Pennsylvania border.
One possible explanation for this band structure is that frictional convergence near the southern shore of Lake Ontario helped to initially release convective instability there; once initiated, advective transport of convective circulations and hydrometeors led to the development of a wind-parallel band downwind of the most intense convective activity. The particular alignment may also be related to the concave notch of Irondequoit Bay, which could produce a local enhancement of convection through land-breeze convergence and extended fetch. This mechanism may have been responsible for the snowfall maximum in eastern Monroe County (Fig. 3).

c. Synopsis

Convection developed in a rapidly evolving synoptic setting as an upper-tropospheric trough moved southeast over Ontario and surface troughing over Lake Ontario shifted southward. Ascent associated with both the upper- and lower-tropospheric troughs helped to increase the lower-tropospheric relative humidity as well as the depth of the cold air mass. Both diabatic boundary layer processes over the eastern Great Lakes and the upper trough likely contributed to the strength of surface troughing near Lake Ontario. The lower-tropospheric troughing effectively increased overwater fetch, and
may have allowed boundary layer air parcel trajectories to traverse both Georgian Bay and Lake Ontario prior to reaching the southern shore of Lake Ontario. Cyclonic isobaric curvature leads to upward vertical motion via Ekman pumping (e.g., Holton 1992, chapter 5), which can also contribute to the lifting of the inversion and increasing relative humidity in the lower troposphere.

The radar and satellite imagery presented in Figs. 9 and 10 describe a complex structure that does not fit cleanly into any of the five LES categories of Niziol et al. (1995). There were two prominent axes of enhanced precipitation: a shore-parallel band that developed immediately south of Lake Ontario and migrated southward, and a wind-parallel band that extended southward from Lake Ontario into northern Pennsylvania. The shore-parallel band may have been similar to the type IV band in that frictional processes at the shoreline played a role in its development. However, it was associated with a mobile trough, and was accompanied by heavy snowfall that was not restricted to the immediate vicinity of the shoreline; these characteristics are not consistent with the type IV band classification. The north–south (wind parallel) band may have involved multilake (type III) interactions initially, other-
wise it must be interpreted as a locally enhanced type II band, as the prevailing flow was perpendicular to the major axis of the lake.

The rapidly evolving lower-tropospheric wind field in the presence of an intense, mobile upper disturbance provides a complex LES environment. The ROC LES climatology (section 3) indicates that mobile upper troughs frequently accompany LES in this region; therefore, it is suggested that the LES classification scheme of Niziol et al. (1995) could be extended to include time-dependent band configurations associated with mobile synoptic-scale forcing in the form of an upper-tropospheric disturbance. The relative importance of lake-induced troughing, the mobile upper trough, lower-tropospheric instability, and local orography during this event will be determined through subsequent numerical modeling experiments.

5. Model forecast errors

Limitations in the horizontal resolution of operational forecast models preclude representation of individual lake-effect bands. Therefore, it is not surprising that operational model forecasts of the November 1996 event failed to represent the lake-induced convection. The horizontal resolution of the operational Eta Model was 48 km at the time of this event, which is far too coarse to represent the axis of heavy convective precipitation indicated in Fig. 9. The Eta Model run initialized at 1200 UTC 26 November produced 3.6 mm (0.14 in.) of liquid equivalent precipitation at ROC for the 24-h period ending at 1200 UTC 27 November, while 20.3 mm (0.80 in.) was measured during that period. Forecasters are cognizant of the limitations of relatively coarse-resolution numerical forecasts in LES situations. However, operational forecasters do rely on numerical guidance to accurately portray the synoptic-scale environment in which LES develops (e.g., Niziol et al. 1995). Synoptic-scale forecast information is used to predict the extent of overwater fetch and the potential instability. This section will investigate errors in the synoptic-scale forecasts of the operational Eta Model for this event.

a. Upper- and lower-tropospheric forecast errors

Figure 11 displays the 12-h forecast fields valid at 0000 UTC 27 November from the operational Eta Model run. Comparison of the forecast fields with the analysis valid at this time in Fig. 5 reveals that the upper trough centered over Ontario was more intense than forecast (cf. Figs. 11a,b and 5a,b). The analyzed 500-hPa $\zeta_v$ value in the trough center was approximately $8 \times 10^{-5}$ s$^{-1}$ larger than in the 12-h forecast, while $p_{new}$ was over 200 hPa higher in the analysis. Figure 11d presents the difference between the 12-h forecast $p_{new}$ and the analysis valid at 0000 UTC 27 November. The $p_{new}$ differences are due to a combination of errors in forecast intensity of the upper trough and errors in the forecast location. The forecasted trough center was approximately 300 km to the southwest of the analyzed position, as indicated by the relative positions of the 500-hPa $\zeta_v$ maxima in the forecast and analysis (Figs. 5b and 11b). Comparison of Figs. 11c and 5c also reveals that the surface trough over Lake Ontario was underpredicted, as was the general extent of troughing over the Great Lakes. Forecasted values of surface potential temperature were too cold, especially over Lake Superior (Figs. 5c and 11c). This cold bias is consistent with the positive bias in forecasts of the lower-tropospheric geopotential height and sea level pressure over the lakes.

Figure 12 presents a more detailed examination of the differences between the 12-h forecasts and analyses valid at both 1200 UTC 26 November and 0000 UTC 27 November. The analyzed 500-hPa geopotential height at 1200 UTC 26 November indicates that the trough axis is located over the central Great Lakes (Fig. 12a). Difference fields (12-h forecast minus analysis) exhibit positive values east and south of the trough axis, and negative values to the northwest. This spatial pattern indicates that the forecasted trough propagated eastward too slowly. The difference in the magnitude of the positive and negative centers indicates that the forecasted trough intensity was weaker than that analyzed; the positive errors east of the trough are larger than the negative errors to the west. A similar analysis for the 12-h forecast valid at 0000 UTC 27 November reveals large positive difference values throughout the vicinity of the upper trough, which is analyzed over the Ontario–Quebec border at this time (Fig. 12b). The difference field is consistent with a continued underforecast of trough intensity, and with data presented in Fig. 11d. Positive differences are also found over the western Great Lakes, especially Lake Superior.

At the 850-hPa level, the difference field for 1200 UTC 26 November exhibits maxima centered over Lakes Superior and Huron, in addition to central Quebec and western Ontario (Fig. 12c). The maxima over the lakes are suggestive of model underprediction of diabatic heating as cold arctic air flowed over the warmer lake surfaces, while the maximum to the northeast of the lakes is likely related to the errors in the upper trough seen in Fig. 12a. At 0000 UTC 27 November, the 850-hPa geopotential height differences are generally smaller than at 500 hPa, with maximum values of approximately 2 dam over Lake Superior (Fig. 12d). Again, the errors over the lakes are suggestive of problems with the model handling of diabatic boundary layer processes over the lakes, while the error over southeastern Ontario is probably due to a combination of diabatic processes and large errors in the upper trough in this region (Figs. 12b,d). Examination of the 12-h Eta Model 850-hPa geopotential height forecast valid at 0000 UTC 27 November confirms what is suggested by Fig. 12d, namely that the trough over southern Ontario is almost completely absent (not shown). Diabatic boundary layer processes alone cannot explain the 850-hPa forecast errors.
over southwestern Quebec and eastern Ontario, given their location upwind of the lakes (Figs. 12c,d). This is especially true for the distinct maximum centered over Quebec at 1200 UTC 26 November (Fig. 12c). Additional research, involving model sensitivity experiments, would be required to isolate the relative contribution of diabatic processes and synoptic-scale errors in the upper trough to the differences shown in Fig. 12. On the basis of this figure, neither error source can be ruled out, and it appears that both may have been significant.

b. Consequences of the poorly forecast surface trough

As indicated in Table 1 and Figs. 7 and 8, a surface trough passage coincided with the period of heavy snow in the Rochester area. A broader trough with a similar orientation is seen at the 850-hPa level immediately north of Lake Ontario (Fig. 12d); the operational Eta Model largely failed to predict this feature, even at 12 h. The horizontal wavelength of the manually analyzed surface trough is on the order of 200 km, and is marginally resolved by the 48-km Eta forecast and poorly represented on the 80-km output grid displayed here. However, careful comparison of Figs. 5c and 11c indicates that even the coarse grid was able to represent differences in the location of the axis of this trough. At the 850-hPa level the trough is broader, and the feature is more adequately resolved in the Eta analysis (Fig. 12d). The consequences of the model misrepresentation of this feature will now be considered.

Due to the cyclonic curvature of the windflow in the vicinity of the lower-tropospheric trough, lower-tropospheric trajectories passing over the western portion of Lake Ontario experienced a greater overwater fetch than would have occurred without the trough, and may have been sufficiently curved to allow air that had previously tracked over Georgian Bay to be advected southeastward over Lake Ontario and subsequently into western New York. Without the lower trough, boundary layer windflow would have been more northerly, and would
have exhibited a substantially shorter overwater fetch. To provide some rudimentary support for this contention, Fig. 13a displays streamlines based on objectively analyzed surface observations from 0000 UTC 27 November 1996, while Figs. 13b and 13c display forecast streamlines from the Eta analysis and 12-h forecast from 1200 UTC 26 November. This comparison indicates that the Eta Model forecast failed to produce sufficient cyclonic curvature in the boundary layer wind field, resulting in underestimation of overwater fetch for boundary layer trajectories. Note that the Eta analysis indicates a more favorable northwesterly flow relative to the 12-h forecast (Figs. 13b,c). The objectively analyzed streamlines also appear to capture the confluence along the axis of the developing wind-parallel band, and support troughing and the shore-parallel band as well (Fig. 13a).

As expected, a comparison of forecast and analyzed vertical motion for the same times as presented in Fig. 12 indicates that the more pronounced troughing in the analysis is accompanied by stronger ascent over most of southern Ontario and western New York (not shown). This is consistent with increased Ekman pumping in the boundary layer in the presence of cyclonic geostrophic relative vorticity. A related consequence would be a tendency for model underprediction of the height of the capping inversion atop the planetary boundary layer.

The discussion issued by the NWS forecast office in Buffalo at 1452 UTC (0952 EST) 26 November noted a growing precipitation shield over southern Ontario and LES developing downwind of Lake Huron and Georgian Bay. The discussion commented on the intense upper trough dropping southeastward over Lake Huron. Recognition of these factors eventually led to a revised New York State zone forecast issued at 2001 UTC (1501 EST) 26 November, which called for snow showers with “a fresh coating to an inch” accumulation by evening, followed by a 40% chance of snow showers overnight. Forecasts of light accumulations in western New York

![Fig. 12. Comparison of 12-h Eta Model forecasts with analyses: (a) 500-hPa geopotential height analysis valid 1200 UTC 26 Nov 1996 (solid, contour interval 3 dam), and geopotential height difference (forecast minus analysis, contoured and shaded above and below 1 dam, shade and contour interval 0.5 dam) between 12-h forecast valid at 1200 UTC 26 Nov and the analysis, (b) as in (a) except analysis is for 0000 UTC 27 Nov and the difference field is based on the 12-h forecast initialized 1200 UTC 26 Nov, (c) as in (a) except for 850 hPa, and (d) as in (b) except for 850 hPa.](http://journals.ametsoc.org/doi/pdf/10.1175/1520-0434(2001)016<0099:AOASWN>2.0.CO;2?cookieSet=1)
FIG. 13. Streamline comparison for 0000 UTC 27 Nov 1996: (a) based on objectively analyzed surface wind observations, (b) based on Eta 10-m wind analysis, and (c) based on 12-h Eta forecast initialized 1200 UTC 26 Nov.

were consistent with the Eta Model predictions of northerly lower-tropospheric flow and a correspondingly short overwater fetch for Lake Ontario (Fig. 13c). Clearly, the model underprediction of the lower-tropospheric trough fed back into operational forecast errors.

c. Possible causes of model error

One factor that may have contributed to the Eta mis-forecast of the upper trough is that the upper disturbance was located over central Ontario at the time of model initialization. The closest rawinsonde site to the upper trough at 0000 UTC 26 November was Pickle Lake, Ontario (Fig. 14a), but there were no rawinsonde sites near the core of the upper trough. At 1200 UTC 26 November, the closest rawinsonde site was Alpena, Michigan (Fig. 14b). The compact nature of the trough suggests the possibility that the trough center fell between rawinsonde sites, with the result being that the model did not capture its initial intensity. It is true that the sophisticated data assimilation systems currently utilized by the operational centers are no longer heavily reliant on rawinsonde data over large portions of the forecast domain. However, examination of an archived database of commercial aircraft reports (ACARS), obtained from the NOAA Forecast Systems Laboratory (FSL), indicates that no aircraft data were available in the vicinity of the upper trough on 26 November. It is likely that the model first guess represented the largest contribution to the analysis in the vicinity of the upper trough at both 0000 and 1200 UTC 26 November, although it is extremely difficult to quantify the contributions of various data sources in these analyses retrospectively.

The analysis presented above also suggests that model errors relating to the representation of diabatic boundary layer processes over the Great Lakes were significant. Comparisons of forecast and analyzed model temperature profiles, and difference fields between forecast and observed near-surface temperature fields in the vicinity of the Great Lakes are consistent with model underprediction of the troughing over the western lakes. However, lower-tropospheric temperature errors are much smaller over the eastern lakes, and vertical error profiles are not maximized in the boundary layer, suggesting that errors in the forecast of the upper trough were more
critical there. It is beyond the scope of the current study to speculate as to the cause of diabatic errors in the Eta Model; this topic is left to future research.

6. Summary, conclusions, and future research directions

On 26–27 November 1996, an unexpected LES event dropped up to 30 cm (12 in.) of snow over parts of western New York. In order to provide context for this case study, a 32-case sample of historical LES events in which ROC experienced snowfall in excess of 15 cm (6 in.) was analyzed. Composites were constructed with the purpose of identifying synoptic-scale common denominators prior to and during ROC LES. Three prominent features of the composite (an arctic high pressure system in the Midwest, a cyclone over eastern Canada, and troughing over the Great Lakes) have been observed and analyzed in previous studies (e.g., Wigginton 1950; Petterssen and Calabrese 1959; Jiusto et al. 1970; Ellis and Leathers 1996). A prominent feature of the composite that has not been emphasized in earlier studies is a mobile upper trough. Although the role of mobile upper troughs in enhancing LES is known (e.g., Niziol et al. 1995, section 4c), these results suggest that such disturbances are of primary importance to significant ROC LES events.

On 26–27 November 1996, a mobile upper trough associated with lake-induced convection produced an unexpected snowstorm that deposited 20.8 cm (8.2 in.) of snow at ROC, with accumulations up to 30 cm (12 in.) in parts of Monroe County. Synoptic-scale ascent due to the approaching upper trough produced a widespread cloud and precipitation shield that contained embedded convective precipitation bands. The upper trough is hypothesized to have increased the inversion altitude and relative humidity in the lower troposphere, and likely contributed to the strength of surface troughing over Lake Ontario during the event. Cyclonic isobaric curvature accompanying the surface trough enhanced lower-tropospheric ascent through Ekman pumping and increased the overwater fetch for boundary layer air parcels traversing Lake Ontario. Examination of GOES-8 satellite and WSR-88D radar imagery reveals two primary axes of heavy precipitation: a shore-parallel band with heavy snow that was located ahead of a mobile surface trough, and a wind-parallel band of moderate and heavy snow that extended from Lake Ontario southward into northern Pennsylvania. It is probable that both bands were strongly influenced by the presence of the mobile upper trough and a rapidly changing synoptic-scale wind field. This complex LES structure does not fit cleanly into the LES categorization scheme of Niziol et al. (1995).

Although forecasters do not look to operational numerical forecast models to forecast individual lake-effect bands, they do rely on these models to accurately predict the synoptic-scale environment in which LES develops. In the November 1996 event, Eta Model forecasts indicated a general northerly flow across the minor axis of Lake Ontario and a short overwater fetch upwind of western New York. Lower-tropospheric troughing over and north of Lake Ontario was more pronounced than forecast, leading to increased overwater fetch. Forecast errors are hypothesized to be the result of a combination of surface-based diabatic processes and poor model initialization in the vicinity of the mobile upper trough. The upper trough was located in a data-sparse region during the initialization. The underforecast of the upper trough intensity may have contributed to errors in forecasts of the lower-tropospheric troughing over Lake Ontario. The surface trough allowed an effectively increased overwater fetch to be realized, and sufficient moistening and destabilization occurred for lake convection to erupt.

Future research is required in order to isolate the relative contributions of diabatic boundary layer processes over the Great Lakes and the mobile upper trough. Preliminary examination of a simulation from the National Center for Atmospheric Research–Pennsylvania State
University fifth-generation Mesoscale Model initialized at 0000 UTC 27 November indicates that a lobe of stronger vorticity advected over the south shore of Lake Ontario during the time of heavy precipitation (not shown). Model initialization of this feature could be tested using sensitivity studies involving the inclusion of initial sounding data over Ontario. Trajectory analyses could be used to determine the degree to which lake–lake interactions were at work in this case. A model sensitivity test could also aid in determination of the degree to which the Lake Ontario trough was forced by the mobile upper trough versus diabatic boundary layer processes.

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