Factors Affecting the Inland and Orographic Enhancement of Lake-Effect Precipitation over the Tug Hill Plateau

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(Manuscript received 15 December 2017, in final form 6 April 2018)

ABSTRACT

The factors affecting the inland and orographic enhancement of lake-effect precipitation are poorly understood, yet critical for operational forecasting. Here we use nine cool seasons (16 November–15 April) of radar data from the Montague/Ft. Drum, New York (KTYX), WSR-88D, the North American Regional Reanalysis (NARR), and observations from the Ontario Winter Lake-effect Systems (OWLeS) field campaign to examine variations in lake-effect precipitation enhancement east of Lake Ontario and over the Tug Hill Plateau (hereafter Tug Hill). Key factors affecting the inland and orographic enhancement in this region include the strength of the incident boundary layer flow, the intensity of the lake-induced convective available potential energy (LCAPE), and the mode of the lake-effect system. Stronger flow favors higher precipitation rates, a precipitation maximum displaced farther downwind, and greater inland and orographic enhancement. The effects of LCAPE depend upon the strength of the flow. During periods of weak flow, higher LCAPE favors lower precipitation rates, a maximum closer to the shoreline, and lesser inland and orographic enhancement. During periods of strong flow, higher LCAPE favors higher precipitation rates, a maximum displaced farther downwind, and greater inland and orographic enhancement. Banded (nonbanded) modes favor higher (lower) precipitation rates, lesser (greater) inland and orographic enhancement, and a maximum closer to the shoreline (over Tug Hill). These results, for both manually measured and radar-estimated precipitation, are robust when many lake-effect events are considered, but substantial variability exists during individual events.

1. Introduction

The frequent inland and orographic enhancement of lake-, sea-, and ocean-effect (hereinafter lake effect) precipitation results in climatological precipitation maxima over downstream hills, mountains, and upland regions. As summarized by Niziol et al. (1995, p. 62) for the Laurentian Great Lakes, “the greatest snowfall occurs where the prevailing winds blow [along] the longest fetch of the lake, particularly where orographic features enhance precipitation processes.” Veals and Steenburgh (2015) show that a prominent maximum in mean cool-season lake-effect liquid precipitation equivalent (LPE) exists over the Tug Hill Plateau (hereinafter Tug Hill), a broad upland region east of Lake Ontario that rises 500 m above lake level. Similarly, downstream of the Great Salt Lake of northern Utah, mean cool-season LPE during lake-effect periods increases fourfold from the lake shore to the Wasatch Mountains (Yeager et al. 2013). Additionally, high elevation sites along the west coast of Japan, where the Sea of Japan produces heavy snowfalls during the Asian winter monsoon, observe seasonal snowpacks of immense depth, with snow-water equivalent frequently exceeding 300 cm (Yamaguchi et al. 2011).

During individual lake-effect storms, however, the inland and orographic enhancement of LPE can vary widely. In some cases, the ratio of upland to lowland LPE can greatly exceed that expected from climatology, with a dramatic increase in LPE with elevation, whereas in others it may be <1, with lowland snowfall exceeding that at upper elevations (e.g., Magono et al. 1966; Ishihara et al. 1989; Nakai and Endoh 1995; Eito et al. 2005; Nakai et al. 2008; Campbell et al. 2016).
In this paper, we examine the factors affecting such inland and orographic variations in lake-effect precipitation east of Lake Ontario and over Tug Hill. This represents a unique problem, requiring a synthesis of knowledge of lake-effect precipitation, coastal and inland effects, and orographic processes. To provide a foundation for our analysis, we begin with a synthesis of prior literature in these areas.

2. Literature synthesis

a. Environmental conditions and lake-effect mode

Lake-effect precipitation occurs when a cold air mass flows over a relatively warm body of water, initiating moist convection (e.g., Peace and Sykes 1966; Hjelmfelt and Braham 1983; Niziol et al. 1995). Sensible and latent heat fluxes over the water body generate a convective boundary layer that deepens with downwind extent, is frequently surmounted by a capping inversion or stable layer (hereafter cap), and contains precipitating clouds. The depth of the lake-effect boundary layer is modulated by the characteristics of the upstream air mass, the strength of the sensible and latent heat fluxes, and the fetch, with a larger lake–air temperature difference, stronger winds, and longer fetch producing a deeper boundary layer and higher cap (Hill 1971; Kristovich et al. 1999; Laird and Kristovich 2002). The strength of the fluxes and height of the cap in turn affect the behavior and intensity of the lake-effect convection. Larger fluxes and a higher cap enable deeper, stronger convection and greater LPE downwind of the lake (e.g., Braham 1983; Niziol 1987; Hjelmfelt 1990; Byrd et al. 1991; Smith and Boris 2017).

For operational forecasting, the potential for boundary layer growth and lake-effect convection is often assessed using estimates of the lake-induced instability such as the lake–850-hPa temperature difference, \( \Delta T_{\text{Lake}-850} \) (\( \Delta T_{\text{Lake}-700} \) for higher-elevation lakes; Niziol 1987; Niziol et al. 1995; Alcott et al. 2012) or the lake-induced convective available potential energy, LCAPE, defined as the surface-parcel CAPE calculated from an observed, analysis, or forecast sounding using the lake temperature for the surface air temperature and dewpoint (e.g., Miner and Fritsch 1997; Steiger et al. 2000). Higher values of lake-induced instability are thought to yield a higher probability of convective initiation, stronger updrafts, and higher LPE rates, though this has not been evaluated explicitly (Niziol 1987; Hjelmfelt 1990; Niziol et al. 1995).

Lake-effect convection frequently assumes the mode of disorganized open cells or horizontal roll convection, referred to collectively as broad coverage in some studies (Hjelmfelt 1990; Steenburgh et al. 2000; Veals and Steenburgh 2015; Campbell et al. 2016). The likelihood of roll convection increases with wind speed and vertical wind shear (Kristovich et al. 1999). The resulting cloud and precipitation bands can align either parallel to or normal to the mean boundary layer wind and are sometimes referred to as longitudinal (“L” mode) and transversal (“T” mode) bands, respectively (Nakai et al. 2005). The band orientation is determined by the orientation of the wind shear vector relative to the mean wind vector, with longitudinal (transversal) bands favored when the two are approximately parallel (normal) (Asai 1972; Kelly 1986; Kristovich 1993; Kristovich et al. 1999; Cooper et al. 2000; Eito et al. 2010).

In addition to boundary layer circulations, convection generated by thermally driven flows can influence lake-effect mode. Areas where the upstream shore features a large bay or channel favor thermally driven convergence and lake-effect initiation, with the resulting bands often broader and more intense than neighboring bands produced by horizontal roll convection (e.g., Anderson and Gustafsson 1994; Norris et al. 2013; Mazon et al. 2015). When the wind blows along the major axis of an elongated lake (e.g., Lakes Michigan, Erie, and Ontario) land-breeze convergence can generate an intense, mesoscale band that produces especially large snowfall rates and accumulations. Although called midlake or type-1 bands by some authors (e.g., Braham 1983; Kelly 1986; Niziol et al. 1995; Steenburgh et al. 2000), following more recent studies (Steiger et al. 2013; Veals and Steenburgh 2015; Steenburgh and Campbell 2017), we refer to these as long-lake-axis parallel (LLAP) bands. LLAP bands are favored during periods of large lake–land temperature differences, \( \Delta T_{\text{Lake}-\text{Land}} \) (e.g., Braham 1983; Veals and Steenburgh 2015), and typically occur during periods of weak directional wind shear (Niziol 1987; Niziol et al. 1995). During periods of large \( \Delta T_{\text{Lake}-\text{Land}} \) and weak large-scale flow, convection along the land-breeze front may produce a quasi-stationary near or offshore shoreline band (e.g., Passarelli and Braham 1981; Hjelmfelt 1990; Niziol et al. 1995; Laird et al. 2003a,b; Laird and Kristovich 2004).

b. Coastal and inland effects

Lake-effect systems undergo significant changes as they approach the downstream coast and penetrate inland, in some cases interacting with a land breeze or land-breeze front. Campbell and Steenburgh (2017) showed that a key contributor to the enhancement of LPE during a lake-effect event over Tug Hill was a land-breeze front that formed along the southeast shore and cut obliquely across the lake-effect system. Along the
west coast of Japan, offshore katabatic flow from the coastal topography can oppose the large-scale flow, re-inforcing convergence along the land-breeze front and producing a coast-parallel precipitation band (e.g., Ishihara et al. 1989; Eito et al. 2005).

As the lake-effect system penetrates inland, there is a transition in the character of lake-effect convection. Downstream of Lake Ontario, Minder et al. (2015) and Welsh et al. (2016) showed that lake-effect systems become less turbulent, more spatially continuous, and shallower with inland extent, consistent with a convective to stratiform transition. The inland penetration of lake-effect precipitation varies, with Villani et al. (2017) concluding that higher boundary layer wind speed, a connection of the lake-effect system to upstream lakes, and a lower $\Delta T_{\text{Lake--850}}$ lead to greater inland penetration.

c. Orographic precipitation processes

The combined influence of wind speed and static stability on orographic flows and precipitation is frequently evaluated using the nondimensional mountain height (or inverse Froude number), defined as

$$H = \frac{N_d h_m}{U},$$

where $N_d$ is the dry Brunt–Väisälä frequency, $h_m$ is the height of the barrier, and $U$ is the component of the wind speed normal to the barrier (e.g., Pierrehumbert and Wyman 1985; Baines 1987; Smith 1988; Durran 1990). Values of $H \gg 1$ are associated with windward blocking and flow along or around the barrier. As $H$ decreases, the blocking and deceleration decrease, with unblocked flow for $H \ll 1$ (e.g., Pierrehumbert and Wyman 1985; Smolarkiewicz and Rotunno 1989; Smith 1989; Durran 1990; Galewsky 2008). This relationship assumes dry flow, however. For saturated flows, the moist Brunt–Väisälä frequency $N_w$ can be used to characterize stability (e.g., Fraser et al. 1973; Barcilon et al. 1979; Durran and Klemp 1982; Jiang 2003).

The flow regime has a profound influence on the distribution of LPE upstream and over a barrier, with higher (lower) values of $H$ associated with a larger (smaller) ratio of upland to lowland precipitation (e.g., Neiman et al. 2002; Colle 2004; Hughes et al. 2009; Yuter et al. 2011). However, a single value of $H$ is difficult to define in real-world situations where the static stability and $U$ are nonuniform (Reinecke and Durran 2008), and the application of an appropriate $N_d$ and $N_w$ is problematic because of factors such as uncertainty in the distribution of moisture and location of the saturated region (Barcilon et al. 1979; Barcilon and Fitzjarrald 1985; Rotunno and Ferretti 2001; Jiang 2003). Microphysical processes and mountain wave dynamics also contribute to a highly nonlinear response in orographic precipitation, so caution must be used with simplistic treatment of these parameters (e.g., Jiang 2003; Jiang and Smith 2003; Colle 2004; Smith and Skillingstad 2011).

The effects of $U$ are not limited to the determination of flow over or flow around the barrier. Increasing wind speed for a given moisture profile results in greater up slope moisture flux, potentially increasing LPE amounts over the barrier (e.g., Neiman et al. 2002). Stronger flows can also advect hydrometeors farther downwind, in some cases increasing spillover into the lee (e.g., Sinclair et al. 1997).

The orographic enhancement can also be affected by the presence of nonorographic ascent. Steenburgh (2003) examined a storm cycle over the Wasatch Range and found that the lowest ratio of upland to lowland precipitation occurred during the passage of a cold-frontal precipitation band, when strong mesoscale ascent produced similar LPE in both the lowlands and uplands. Similarly, Mass et al. (2015) found that periods of precipitation forced by strong synoptic scale ascent were associated with little orographic enhancement over the Cascade Mountains.

d. Summary

Advancing knowledge of the inland and orographic enhancement of lake-effect storms requires the synthesis of concepts from the preceding sections. The cap height likely has some control over the depth and intensity of the convection, while also influencing the characteristics of the flow over the downstream barrier (e.g., a cap near or below crest level might induce blocking). High wind speeds may be associated with more intense lake-effect precipitation (Smith and Boris 2017), while simultaneously increasing the cross-barrier moisture flux and inland hydrometeor transport (Sinclair et al. 1997; Neiman et al. 2002). Similar to the effects of mesoscale and synoptic-scale forcing on orographic precipitation, the mesoscale organization of lake-effect precipitation systems may affect the ratio of upland to lowland LPE. Single, organized bands (LLAP and shoreline) generally feature the most intense LPE rates and ascent, and the location of their associated LPE/ascent maxima may supersede any orographically induced ascent to produce a lowland LPE maximum or a nearly equal distribution between upland and lowland sites (e.g., Ishihara et al. 1989; Eito et al. 2005; Campbell et al. 2016).

The remainder of this paper examines the inland and orographic enhancement of lake-effect precipitation east of Lake Ontario and over Tug Hill. In section 3, we describe the datasets and methods used in the analysis, which is based on a multicool-season radar climatology.
and observations from the Ontario Winter Lake-effect Systems (OWLeS) project, conducted from December 2013 to January 2014 [see Kristovich et al. (2017) for a full description]. The radar climatology is used to examine the factors affecting the inland and orographic enhancement of lake-effect precipitation in section 4, while the OWLeS observations are used to examine a smaller subset of lake-effect periods in section 5. A summary and conclusions are presented in section 6.

3. Data and methods

a. Radar climatology

Development of the radar climatology for this study began with the lake-effect periods identified east of Lake Ontario by Veals and Steenburgh (2015) using lowest-level reflectivity data from the Montague/Ft. Drum, New York (KTYX), WSR-88D during the 13 cool seasons during 2001–14. Lake-effect periods during the subsequent cool seasons (2014–15, 2015–16, and 2016–17) were then identified following Veals and Steenburgh (2015) and added to the analysis. We then focused on the period of continuous level-II radar data, which spans nine cool seasons during 2008–17, because the level-II format includes full-volume scans, allowing for the plotting and analysis of higher-quality vertical cross sections. We also focused on periods from 16 November to 15 April to minimize the inclusion of lake-effect rain events that could include a radar bright band.

To simplify the analysis, we concentrated on times during these periods when lake-effect features extend along a transect that runs across the center of Tug Hill, through the climatological lake-effect precipitation maximum identified by Veals and Steenburgh (2015), and approximately over two OWLeS snow measurement sites, North Redfield (NR) and Sandy Creek (SC; Fig. 1). Following Campbell et al. (2016), we classified the lake-effect mode along the transect during these periods as banded, weakly banded, nonbanded, or other based on visual inspection of lowest-level radar reflectivity data. Thus, our analysis concentrates on 311 periods (683 h) with either banded, weakly banded, or nonbanded lake-effect features along the transect.

For plotting purposes, we interpolated the level-II data to a Cartesian grid using the Radx C++ software package. We also produced radar-estimated LPE using the $Z-S$ relationship described by Vasiloff (2001) and Campbell et al. (2016) and given by

$$Z = 75S^2,$$

where $Z$ is the lowest-level radar reflectivity factor (mm$^6$ m$^{-3}$) and $S$ is the LPE rate (mm h$^{-1}$). Compared to manual OWLeS LPE observations for the 160 h they are available, the radar-derived LPE overestimates the manual LPE by 10% at NR and underestimates the manual LPE by 15% at SC. KTYX was upgraded to dual-polarization capability in July 2012, and comparison of results before and after the upgrade showed no discernable difference.

b. Environmental conditions

To examine the relationship between inland/orographic enhancement and environmental conditions during lake-effect periods, we focused on three parameters that were selected based on a review of the lake-effect and orographic precipitation literature. These parameters are as follows: 1) the lake-effect mode along the transect, 2) the mean 950–850-hPa zonal wind speed $U$, and 3) the LCAPE. The lake-effect mode was identified following Campbell et al. (2016). Given a lack of regular upper-air sounding data directly upstream of Lake Ontario and the low temporal resolution (12 h) of sounding data, the data for $U$ and LCAPE came from
the North American Regional Reanalysis (NARR; Mesinger et al. 2006), produced at a horizontal grid spacing of 32 km and available with a vertical spacing of 25 hPa in the lower troposphere and 50 hPa in the mid-troposphere. These horizontal and vertical grid spacings likely result in smoother representations of the lake-effect environment and boundary layer structure than observed. NARR analyses are available every 3 h, and the values from the nearest analysis time were utilized in all calculations.

During lake-effect events, mesoscale circulations induced by Lake Ontario result in low-level flow convergence over eastern Lake Ontario and Tug Hill (e.g., Steenburgh and Campbell 2017; Campbell and Steenburgh 2017). As a result, no single location serves as the source region for air impinging on Tug Hill. Therefore, all variables are averaged across all grid points within a rectangle spanning the study area (Fig. 1). Results obtained using a grid point immediately upstream of the eastern shore of Lake Ontario were generally similar and are not presented here.

The zonal wind speed \( U \) represents the mean zonal boundary layer wind along the transect and the flow incident on Tug Hill. It was chosen because the strength of the incident flow is important in determining the distribution of orographic precipitation, and it may influence the inland evolution and penetration of lake-effect precipitation (e.g., Sinclair et al. 1997; Neiman et al. 2002; Alcott and Steenburgh 2013; Villani et al. 2017). The 950–850-hPa layer was chosen because this
layer is likely contained within the lake-effect boundary layer in most events. For example, Byrd et al. (1991) examined mobile soundings from lake-effect events near and around Lake Ontario and reported mixed-layer depths of 1.5–1.9 km in weak events and 2.3–3.1 km in strong events.

LCAPE represents an estimate of the thermodynamic forcing and environment for lake-effect convection. It likely affects the intensity of the convection and resulting snowfall rates (e.g., Niziol 1987; Steiger et al. 2009). To calculate LCAPE, we utilized NARR temperature and dewpoint profiles and the horizontal mean lake-surface temperature from the Great Lakes Environmental Research Laboratory’s Great Lakes Coastal Forecasting System (GLCFS), also available at 3-h intervals. LCAPE is defined as

\[
LCAPE = \int_{SFC}^{\text{LNB}} g \left( \frac{T_{\text{lake}} - T_{\text{env}}}{T_{\text{env}}} \right) dz,
\]

where \( T_{\text{lake}} \) is the parcel temperature (K), initialized to the mean GLCFS lake-surface temperature, \( T_{\text{env}} \) is the environmental temperature (K) from the NARR, \( g = 9.81 \text{ m s}^{-2} \), SFC is the surface, and LNB is the level of neutral buoyancy (m). The initial parcel dewpoint was set to \( T_{\text{lake}} \).

We also evaluated the influence of the \( \Delta T_{\text{Lake-Land}} \), \( \Delta T_{\text{Lake-850}} \), planetary boundary layer (PBL) depth, and \( \bar{H} \); where \( \Delta T_{\text{Lake-Land}} \) was defined as the difference between the mean temperature at 12 surface sites surrounding Lake Ontario and the mean GLCFS lake-surface temperature, and \( \Delta T_{\text{Lake-850}} \) is the difference between the mean GLCFS lake-surface temperature and the NARR 850-hPa temperature. PBL depth was obtained from the NARR, in which it is calculated following Janjić (1990). We found, however, that \( \Delta T_{\text{Lake-Land}} \) and \( \Delta T_{\text{Lake-850}} \) have a smaller influence on the distribution of lake-effect precipitation east of Lake Ontario than LCAPE, which is strongly correlated with \( \Delta T_{\text{Lake-Land}} \) and implicitly accounts for \( \Delta T_{\text{Lake-850}} \). The influence of PBL depth was comparable to LCAPE, but we elected to focus on the latter because it is more prognostic and less parameterized, making it more easily applied across a variety of gridded datasets. The \( \bar{H} \) for dry and saturated flows was calculated using an \( h_n \) of 500 m, \( \bar{U} \), and the mean NARR temperature and moisture profiles from 950–850 hPa for \( N_d \) and \( N_w \), respectively. However, nearly every case yielded \( \bar{H} \ll 1 \) for dry flow and imaginary values of \( \bar{H} \) for saturated flows because of the large boundary layer lapse rates. Therefore, in this region where the topography is modest and is typically confined to within the lake-effect boundary layer, \( \bar{H} \) was not useful.

c. OWLeS data

To augment the radar-based climatological analysis, we also examined events during OWLeS, when LPE observations at SC and NR were available. These two sites measured (among other variables) LPE with an automated gauge, automated snow depth, 6-hourly manual LPE, and 6-hourly manual snowfall (Campbell et al. 2016).

The automated gauge LPE observations at NR and SC have high (5 min) temporal resolution, but comparison with manual observations taken at 6-h intervals revealed
some problems with the gauge observations, principally undercatch and the adherence of snow to the walls of the gauge. Thus, following Campbell et al. (2016), we utilized the radar-based technique of Wüest et al. (2010) to disaggregate the 6-hourly manual precipitation totals into accumulations with the high temporal resolution of KTYX scans (~5 min). Specifically, the disaggregated LPE amounts are calculated using

\[ G_t = \frac{S}{S_{6h}} G_{6h}, \]

where \( G_t \) is the radar-disaggregated LPE estimate for the period of interest, \( S \) is the radar-derived LPE for that period, \( S_{6h} \) is the 6-h radar-derived LPE estimate, and \( G_{6h} \) is the manually measured 6-h LPE. These surface precipitation observations complement the long-term KTYX statistics, with smaller errors in radar-estimated precipitation rates.

d. Quantification of inland and orographic enhancement

To objectively compare the relationship of the three parameters to the LPE distribution downstream of Lake Ontario and over Tug Hill, we define six variables, all calculated along a portion of the transect stretching from the Lake Ontario shoreline through Tug Hill to the base of the Adirondack Mountains (hereafter Adirondacks; Fig. 1). LPE is the transect-mean radar-estimated precipitation rate (mm h\(^{-1}\)). The inland displacement, InDisp, is the distance (km) from the shore to the LPE rate maximum. The enhancement ratio, ER, is ratio of the maximum LPE rate to the LPE rate at the shoreline. The absolute enhancement, AE, is
the difference between the maximum LPE rate and the LPE rate at the shore line. InDisp and AE are zero if the maximum LPE rate occurs at the shoreline. Finally, for the analysis of the OWLeS data, ERNR/SC is the ratio of disaggregated LPE rate at NR to SC and AENR–SC is the difference.

4. Radar-based climatology

4.a. Univariate statistics

To examine the influence of $\mathit{U}$ on the lake-effect precipitation, we divided $\mathit{U}$ values during lake-effect periods into quintiles, with the lower quintile (0–20th percentile; $\leq 7.7 \text{ m s}^{-1}$), middle quintile (40th–60th percentile; 10.2–12.1 m s$^{-1}$), and upper quintile (80th–100th percentile; $\geq 14.2 \text{ m s}^{-1}$) referred to hereafter as low, moderate, and high $\mathit{U}$, respectively. Low $\mathit{U}$ periods feature relatively low LPE rates and weak inland penetration of lake-effect precipitation, with the LPE-rate maximum near the initial windward slope of Tug Hill (Figs. 2a,b). As $\mathit{U}$ increases to moderate (Figs. 2c,d) and high (Figs. 2e,f) values, the LPE rates increase, echoes deepen, inland penetration increases, and the LPE-rate maximum moves over upper Tug Hill. Relatively high LPE rates also spread downstream of Tug Hill into the western Adirondacks. Cross sections show that radar echoes for low $\mathit{U}$ periods tend to be shallow and decrease quickly in depth with distance from the shoreline, but as wind speeds increase to moderate and high, echoes become deeper throughout the transect, with the depth decreasing abruptly in the immediate lee of Tug Hill, followed by a rebound over the Adirondacks, consistent with a mountain-wave response (Figs. 2b,d,f).

Quantifying these effects, an increase from low to high $\mathit{U}$ approximately doubles the median LPE and median InDisp, increases the median ER from 2.1 to 5.0, and increases the median AE from 0.7 to 2.0 mm h$^{-1}$ (Fig. 3).

Although there is variability by period, these median differences are all statistically significant at the 95% confidence level.

LCAPE was similarly divided into lower ($\leq 802 \text{ J kg}^{-1}$; hereinafter low), middle (1066–1454 J kg$^{-1}$; hereinafter moderate), and upper quintile ($\geq 2055 \text{ J kg}^{-1}$; hereinafter high) values. Low LCAPE periods feature relatively low LPE rates throughout the domain, with a pronounced maximum over Tug Hill (Figs. 4a,b). LPE rate contours closely follow terrain contours on Tug Hill, indicating a tendency for lake-effect precipitation to intensify or develop over the western (windward) slope. As LCAPE increases to moderate (Fig. 4c) and high (Fig. 4e) values, the LPE rates increase. In addition, the area of lake-effect precipitation extends farther upstream and exhibits a banded structure over eastern Lake Ontario. Radar echoes during low LCAPE periods tend to be shallow and most frequent near the mid slope of Tug Hill, decreasing sharply in frequency and depth in the lee. As LCAPE increases to moderate and high values, echo frequencies increase over the coastal lowlands and eastern Lake Ontario (Figs. 4b,d,f). At high LCAPE, echo frequencies suggest a decrease in the
depth of echoes as one moves inland over Tug Hill. In terms of overall statistics, an increase from low to high LCAPE increases the median LPE from 0.61 to 0.80 mm h\(^{-1}\), and as a result of the increased LPE rates in the lowlands, decreases the median ER from 4.9 to 2.4 (significant at the 95% level; Figs. 5a,c). The median InDisp decreases from 37 km during low LCAPE to 31 km during high LCAPE periods, though there is no statistically significant difference between moderate and high LCAPE periods (Fig. 5b). There is no statistically significant effect on AE (Figs. 5d).

The lake-effect mode (nonbanded, weakly banded, and banded) along the transect is the third variable presented. Nonbanded periods feature relatively low LPE rates throughout the domain, with precipitation confined mainly to Tug Hill (Figs. 6a,b). LPE rates increase throughout the transect for weakly banded periods (Figs. 6c,d) and further still for banded periods, especially in the lowlands near the shoreline (Figs. 6e,f). Radar echoes tend to be shallow and confined to the plateau for nonbanded periods, but the transition to weakly banded and especially banded periods brings deep echoes over the shoreline, with depth gradually increasing with inland extent (cf. Figs. 6b,d,f). Quantifying these results, LPE increases as banding increases, with the lowest, intermediate, and highest values corresponding to nonbanded, weakly banded, and banded periods, respectively (significant at the 95% level; Fig. 7a). Stronger banding corresponds to a decrease in InDisp as the maximum moves closer to the shoreline (significant at the 95% level; Fig. 7b). ER decreases with increasing banding, but there is no statistically significant trend in AE (Figs. 7c,d).
b. Multivariate statistics

To examine the interactions between $U$, LCAPE, and mode, we selected two values of each variable, resulting in eight different combinations of conditions, and evaluated the effect of each combination upon KTYX LPE rate. The two values of $U$ ($\approx 10.2 \text{ m s}^{-1}$ and $\approx 12.1 \text{ m s}^{-1}$; hereinafter “low” and “high,” respectively) and LCAPE ($\approx 1066 \text{ J kg}^{-1}$ and $\approx 1454 \text{ J kg}^{-1}$; hereinafter “low” and “high,” respectively) correspond to the lower two (0–40th percentile) and upper two quintiles (60th–100th percentile), with these broad ranges used in order to maximize sample sizes, which are generally lower than those in the univariate analysis. The first mode is simply nonbanded, but weakly banded and banded periods were merged into a combined “banded” mode to maintain a sample size comparable to nonbanded periods. The eight different combinations are therefore as follows:

1) nonbanded/low LCAPE/low $U$ (NB/LL/L$U$)
2) nonbanded/high LCAPE/low $U$ (NB/HL/L$U$)
3) nonbanded/low LCAPE/high $U$ (NB/LL/H$U$)
4) nonbanded/high LCAPE/high $U$ (NB/HL/H$U$)
5) banded/low LCAPE/low $U$ (B/LL/L$U$)
6) banded/high LCAPE/low $U$ (B/HL/L$U$)
7) banded/low LCAPE/high $U$ (B/LL/H$U$)
8) banded/high LCAPE/high $U$ (B/HL/H$U$)

The NB/HL/L$U$ and B/HL/L$U$ regimes feature precipitation confined primarily to the lowlands near the shoreline (Figs. 8a,b). These regimes produce little inland penetration and lower echo frequencies and precipitation rates over upper Tug Hill, with low median values of InDisp, ER, and AE (Figs. 9b–d). The combination of high LCAPE and low $U$ results in frequent confinement of lake-effect precipitation to the coastal lowlands and lower slopes of Tug Hill. Such confinement occurs for both nonbanded and banded lake-effect modes, with the latter featuring higher LPE rates. Under such conditions, the lake forcing is strong, but the flow is weak.

In contrast, the NB/LL/H$U$ and NB/HL/H$U$ regimes produce a large increase in LPE rate from the shoreline to a maximum on upper Tug Hill (Figs. 8g,h). These regimes feature the strongest inland penetration and enhancement over Tug Hill, with the highest median values of InDisp and ER (significant at the 95% level; Figs. 9b,c). NB/HL/H$U$ also features the highest median AE of any regime (though not significant at the 95% level; Fig. 9d). The remaining four regimes (B/LL/L$U$, B/LL/H$U$, B/HL/H$U$, and NB/LL/L$U$) feature moderate enhancement, with intermediate median values of InDisp, ER, and AE (Figs. 8c–f, 9b–d). These differences are statistically significant at a 95% level, although the distributions are broad, indicating substantial variability.

Across the eight regimes, nonbanded periods generally feature lower LPE rates, but a maximum that is displaced farther inland, with a larger ratio of upland to lowland LPE compared to banded periods. The absolute enhancement, however, is larger for banded periods than nonbanded periods, with the exception of
nonbanded periods with high LCAPE and high $\overline{U}$, which produce the greatest inland and orographic enhancement of all regimes. The effect of $\overline{U}$ across the eight regimes is generally to increase the mean LPE rate, InDisp, ER, and AE (Fig. 9). Thus, by all metrics, $\overline{U}$ has a strong effect on the inland and orographic enhancement.

5. OWLeS period

The analysis above highlights the influence of $\overline{U}$, LCAPE, and mode on the distribution of lake-effect precipitation downstream of Lake Ontario and over Tug Hill, but it is based on radar-estimated precipitation and there is substantial variability from period to period.
Therefore, in this section we examine a smaller subset of time, the OWLeS period, when disaggregated manual LPE observations from SC and NR are available. These events account for 58 of the 683 h examined above in section 4. The values of \( \bar{U} \) and LCAPE during this period were again divided into lower, middle, and upper quintiles, corresponding to low, moderate, and high values, respectively. Mode was divided into nonbanded, weakly banded, and banded. Because the sample size of the OWLeS period is small, not all of the eight multivariate regimes used in the previous section were observed, therefore, we opt for a univariate approach. Additionally, too few data points existed within the univariate regimes to construct conclusive box-and-whisker plots, therefore, the values of the variables presented for each regime are based on a mean LPE rate at each site.

The effect of \( \bar{U} \) upon the manual LPE at NR and SC during OWLeS is consistent with the radar-derived analysis for the 2008–17 period (Fig. 10). Both the ratio of disaggregated LPE rate at NR to SC (\( \text{ER}_{\text{NR} \to \text{SC}} \)) and the difference (\( \text{AE}_{\text{NR} \to \text{SC}} \)) increase with \( \bar{U} \). The

![Box-and-whisker plot of (a) LPE, (b) InDisp, (c) ER, and (d) AE for each KTYX scan along the transect corresponding to Fig. 8. The difference between the medians of two boxes is statistically significant at the 95% confidence level if the notched areas around the respective medians do not overlap. The colored boxes are added for easier differentiation of conditions, with red boxes indicating the “high” value of a variable and blue boxes indicating the “low” value.](image-url)
The effect of LCAPE is also consistent with the radar-derived analysis, as ER$_{SCNR}$ decreases with increasing LCAPE and the highest value of AE$_{NR-SC}$ occurs for moderate LCAPE, but the difference between low and moderate LCAPE periods is not statistically significant at the 95% confidence level (Fig. 11). For lake-effect mode, the ER$_{SCNR}$ decreases with increased banding, with the highest (lowest) value during nonbanded (banded) periods (Fig. 12). The values of AE$_{NR-SC}$ show little difference between modes.

It should be noted that ER$_{SCNR}$ and AE$_{NR-SC}$ are calculated using data from NR and SC, sites located at the middle and lower slopes of Tug Hill, respectively. In contrast, ER and AE are calculated using data at the shoreline and the maximum, wherever that may be located. Therefore, AE$_{NR-SC}$ and ER$_{SCNR}$ may not capture the full magnitude of the inland and orographic enhancement, but rather only the difference between the lower and middle slopes of Tug Hill. The LPE in each of these methods is also not analogous, as ER$_{SCNR}$ and AE$_{NR-SC}$ are calculated using disaggregated manual LPE observations, and ER and AE are calculated using radar-derived LPE. As can be seen in Figs. 10–12, the mean manual LPE rates at NR and SC (depicted with red dots) sometimes differ from the mean radar-derived LPE rates along the transect (depicted with the blue line). Nevertheless, the KTYX LPE rates are generally fairly close to the observations at NR and SC, and their values relative to one another generally agree with the shape of the KTYX LPE plot. Therefore, despite these caveats, the OWLeS manual LPE observations provide further evidence of the effects of $U$, LCAPE, and mode upon the inland and orographic enhancement.

6. Summary and conclusions

This study has examined the environmental variables affecting the inland and orographic enhancement of lake-effect LPE east of Lake Ontario and over the Tug Hill Plateau. Lake-effect events occurring along the transect were identified for cool season (November–March) periods from November 2008 to March 2017,
with the lake-effect mode along the transect also recorded. The effect of $U$, LCAPE, and lake-effect mode on inland and orographic enhancement was then evaluated for the long-term climatological period as well as the OWLeS period.

High (low) $U$ values are associated with more (less) LPE along the transect, an LPE maximum located farther inland (closer to the shoreline) and higher (lower) on Tug Hill, and high (low) inland and orographic enhancement. High (low) LCAPE values were associated with more (less) LPE along the transect, an LPE maximum located closer to the shoreline (farther inland) and lower (higher) on Tug Hill, and low (high) inland and orographic enhancement. Banded (nonbanded) periods were associated with more (less) LPE along the transect, an LPE maximum located closer to the shoreline (farther inland) and lower (higher) on Tug Hill, and low (high) inland and orographic enhancement. There is substantial variability in these relationships among individual radar scans, but over a sufficient length of time, they are robust and statistically significant at the 95% confidence level.

When considered in a multivariate sense, the effects of each variable can be more complex than those observed in a univariate approach. For periods with low $U$, increasing LCAPE results in more (less) LPE along the transect, an LPE maximum located closer to the shoreline (farther inland) and lower (higher) on Tug Hill, and low (high) inland and orographic enhancement. For periods with high $U$, increasing LCAPE generally has the opposite effect.

The effects $U$ could reflect a number of processes. Sensible and latent heat fluxes over the lake increase with $U$ (e.g., DeCosmo et al. 1996; Veron et al. 2008), and although partially offset by a decreased residence time of air parcels over the lake, this effect may explain the increase in LPE rates throughout the domain corresponding to increased $U$. The maximum in vertical velocity and hydrometeor concentration in lake-effect systems is generally located at the landfalling shoreline, and advection of this maximum can cause an LPE maximum in downstream areas, including Tug Hill (Alcott and Steenburgh 2013; Campbell and Steenburgh 2017). However, Campbell and Steenburgh (2017) show that enhancement over Tug Hill is an important

![Fig. 11. As in Fig. 10, but for LCAPE. There were 157, 149, and 157 scans in the low, moderate, and high LCAPE periods, respectively.](image-url)
contribution to this maximum. Ascent over a barrier requires sufficient flow velocity, and for low wind conditions, there may be flow blocking and stagnation that favors the confinement of LPE to the lowlands near the shoreline. High wind conditions would allow the band or cells to surmount Tug Hill and persist far into the lee. The increased upslope moisture flux that would potentially accompany increasing $U$ could also play a role in the enhancement of LPE over Tug Hill (e.g., Neiman et al. 2002), although this is impossible to explore further without performing a numerical simulation.

The effects of LCAPE are multifaceted. LCAPE represents the theoretical buoyancy of surface parcels over the lake, and higher values likely result in convection that is deeper and/or more intense, has higher LPE rates, and initiates farther upstream over the lake. On the other hand, LCAPE is also strongly correlated with the strength of the forcing for the land breeze. For periods with low $U$, there is a clear and dramatic shift of the LPE maximum toward the shoreline as LCAPE increases, with LPE greater near the shoreline than over Tug Hill. This effect reverses with high $U$, suggesting that the effects of LCAPE are overwhelmed by the strong prevailing flow. During high $U$ periods, increasing LCAPE instead produces higher LPE rates, displaces the LPE maximum farther inland, and increases AE. Increased forcing for convection plausibly explains the higher LPE rates, and it is possible that it allows the precipitating clouds to survive farther inland. The increase in AE is more difficult to explain, but the greater LPE rates throughout the transect could result in a greater magnitude of enhancement.

Lake-effect mode has effects generally consistent with those documented in Campbell et al. (2016) during OWLeS IOP2b. Consistent with their findings, the strong mesoscale ascent within bands produces intense LPE in both the lowlands and uplands, superseding the effects of ascent over Tug Hill, which has a greater effect during nonbanded periods. This reasoning also likely explains the effects upon LPE rates throughout the transect, and the inland displacement of the LPE maximum.

Our results indicate that $U$, LCAPE, and lake-effect mode affect the degree of inland and orographic

![Fig. 12. As in Fig. 10, but for mode. There were 397, 101, and 152 scans in the nonbanded, weakly banded, and banded periods, respectively.](image-url)
enhancement observed during lake-effect storms, but these are by no means the only important environmental variables, therefore, future work should attempt to identify additional variables affecting enhancement. Idealized numerical simulations would be useful in this task, and could also help to identify the underlying processes and causality associated with these variables. Furthermore, Tug Hill represents an isolated terrain feature with relatively little vertical relief above the surrounding plain (~500 m). The topography of other lake-effect regions, such as the west coast of Japan, presents greater relief, quasi-continuous barriers elongated normal to the flow, and is situated within a generally deeper boundary layer during lake-effect periods, making it an ideal location for further study.

Acknowledgments. This research was supported by National Science Foundation Grants AGS-1635654 and AGS-1262090. Comments and input from Justin Minder, Tyler West, and Tom Gowan greatly improved this manuscript. We thank NCEI, NCAR, MathWorks, Unidata, and the University of Utah Center for High Performance Computing for the provision of datasets and/or software. We thank Mike Dixon at NCAR for developing the Radx software package and providing invaluable help running it. The OWLeS data were gathered and made possible by a number of PIs and students working on the program. The instrumentation was also graciously hosted on the property of Jim, Cindy, John, and Cheryl Cheney and Diane and Gerhardt Brosch. Finally, we thank two anonymous reviewers, whose suggestions greatly improved the quality of this manuscript. Any opinions or findings do not necessarily represent those of the National Science Foundation or the University of Utah.

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