rearrangement perturbations, which is not possible if Eulerian perturbations are used.

## See also

**Dynamic Meteorology:** Balanced Flows; Potential Vorticity. **Hamiltonian Dynamics**. **Kinematics**. **Numerical Models:** Methods. **Tracers**. **Wave Mean-Flow Interaction**.

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# LAKE-EFFECT STORMS

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## Introduction

Each winter, lake-effect storms develop on the downwind shores of the North American Great Lakes, as arctic winds blow across the relatively warm water. The associated clouds and snow (or rain) showers tend to organize in narrow bands, usually only a few kilometers wide but sometimes over 200 km long. There may be one band, or there may be as many as 10 or 20, each separated from the next by only a few kilometers of clear sky. These bands may remain stationary over a region or they may oscillate in snakelike fashion. They may produce nothing more than one or two centimeters of snow, or they may dump over 120 cm of snow in a single storm. These lake-effect storms are primarily a product of relatively simple air mass modification by warm water, complicated lakeshore geometry, and the prevailing synoptic situation.

Lake-effect storms develop in other parts of the United States, Canada, and the world, but nowhere else do they occur as frequently or with such intensity as they do in the Great Lakes region. The reasons for the unique weather in the Great Lakes region can be traced to several geographic aspects. The fact that the Great Lakes are the largest single source of fresh water in the world (except for the polar ice caps), the fact that the Great Lakes are situated approximately halfway between the Equator and the North Pole, the fact that Great Lakes are located in the interior of a large continent, the fact that each of the lakes is approximately the size of a small inland sea, and the fact that there are several lakes – separated from each other by distances less than their own size, make for some very unique weather in the region. These characteristics suggest that the lakes rarely freeze over completely, even in the coldest of winters, and thus remain a nearly continuous and very large source of heat and moisture for the atmosphere. Lake-effect storms continue to be a forecast challenge despite improvements in numerical mesoscale models because of their meso-$\gamma$/meso-$\beta$ scale size.

## Climatology

Lake-effect snow accounts for 25–50% of the total annual snowfall in many lakeshore regions (**Figure 1**). The snowbelts (areas of heavier snow) that shoulder the southern and eastern shores of the Great Lakes reflect the direction of the prevailing north-westerly flow relative to the orientation of the lakes, the sharp contrast in surface friction between the relatively smooth lake surface and the rough land, and terrain

effects. The largest snowfall totals exist across the upper peninsula of Michigan, where north-westerly flow across Lake Superior is forced upward abruptly over steep terrain upon reaching the northern coast of the upper peninsula – especially around the Keewanaw Peninsula, and across the Tug Hill Plateau in western New York, where west-south-westerly flow across the lower lakes provides a long fetch and ample opportunity for the air to be moistened and destabilized. In both of these locations, long fetches and orography are key aspects. Terrain can enhance individual snowstorm totals by about 5 cm for every 100 m of rise. Additionally, portions of the lakeshore with enhanced concavity promote convergence zones that can further enhance snowfall totals. Heavier lake-effect amounts fall typically during cold winters, when the lake–air temperature differences are enhanced.

Lake-effect snow falls almost exclusively during the unstable *season* – that portion of the year when the lakes are climatologically warmer than the ambient air and thus provide heat and moisture to the lower atmosphere to destabilize it. Enhanced cloudiness and precipitation exist across much of the lake shore regions and far inland as well. The percentage of cloudy days peaks in November for many places of the Great Lakes region – owing in part to significant lake-enhanced cloudiness. Precipitation during the unstable season begins typically with episodes of nocturnal rain showers during cool nights in late August. As the mean air temperature drops through the fall months, lake-effect rain showers change to lake-effect snow showers. Much of the lake-effect snow falls typically between November and February, which constitutes the heart of the unstable season, when lake-air temperature differences tend to be greatest.
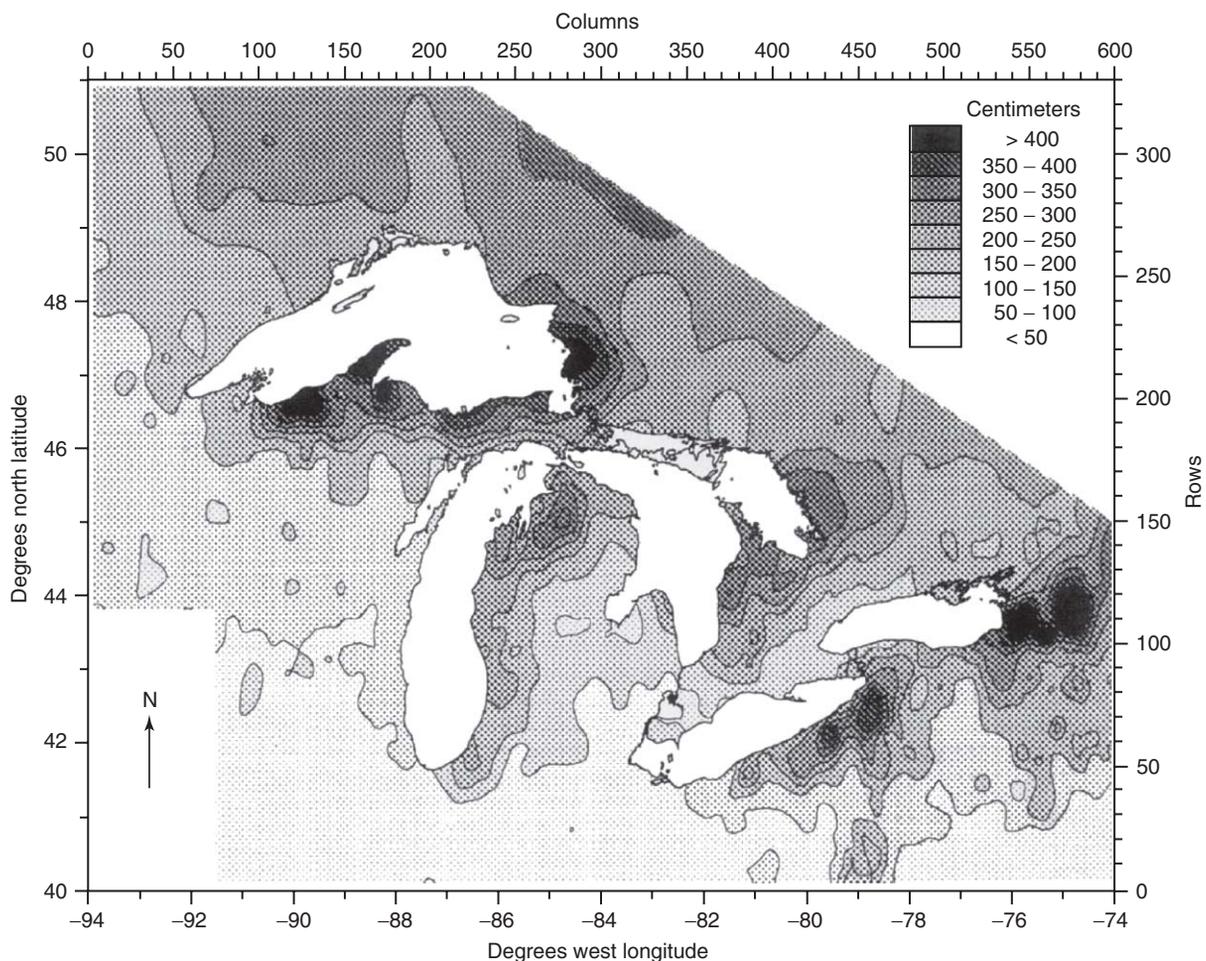


**Figure 1**   Average 1951–1980 Great Lakes seasonal snowfall total. (From Figure 2 in Norton DC and Bolsenga SJ (1993) Spatiotemporal trends in lake effect and continental snowfall in the Laurentian Great Lakes, 1951–1980. *Journal of Climate* 6: 1943–1956. Adapted with permission from the American Meteorological Society.)

Climatological lake–air temperature differences may be around 7–8°C, but may exceed 30°C during intense cold-air outbreaks. Coupled with winds sometimes in excess of $20\,m\,s^{-1}$, combined surface sensible and latent heat fluxes can typically exceed $1000\,W\,m^{-2}$ – comparable to that found in a category-1 hurricane (*see* **Hurricanes**).

Lake-effect clouds and snow can occur locally on ~30–40% of the days in winter under a variety of synoptic patterns – whenever there is an onshore fetch and the lake–air temperature difference allows the lowest layers of the air to destabilize. However, certain synoptic patterns are more favorable than others for allowing lake-effect snow to develop. A typical sequence of events begins with a synoptic-scale low moving across the Great Lakes region from south-west to north-east (**Figure 2**). Additionally, in late autumn especially, these lows are deepening as they cross the region because of baroclinic forcing and aggregate heating from all the lakes. Strong north-westerly winds on the back side of the low bring progressively colder polar or arctic air across the warm lakes. Subfreezing temperatures may reach as far south as the Gulf Coast and northern Florida, with $-20°C$ readings just north of the lakes. The strong winds and cold air generate strong surface fluxes over the lakes that moisten and destabilize the air, leading to snow showers along the downwind lakeshores of the Great Lakes. The deepening of the low and the destabilization both allow stronger winds from above to mix down to the surface and further increase the heat and moisture fluxes.

Depending on the wind speed, the orientation of the wind flow relative to the long lake axis, stability, moisture, and upper-level forcing, different types of lake-effect storms can develop. Basically, when the prevailing flow is more parallel to the short axis than to the long axis of a lake (i.e., there are strong short-axis winds), multiple wind parallel bands (Type II) develop (**Figure 3**, middle panel). These bands are typically 2–4 km wide and spaced 5–8 km apart. Snowfall is usually spread over a large area of the downwind lakeshore, and amounts are usually light (<4 mm liquid precipitation per day). When the prevailing short-axis winds are weak, midlake (Type I) or shore parallel (Type IV) bands can develop – even when the long-axis prevailing winds are strong (**Figure 3** upper panel). These bands can be 10–20 km wide and generate copious amounts of snow. If the short-axis wind is essentially not present then the band will be located near the middle of the lake. If the short axis prevailing wind is present but weak, then the band will be located closer to the downwind shore. Sometimes, especially over Lake Erie or Lake Ontario, a midlake band will develop at an obtuse angle to the long axis of the lake. These bands (Type III) actually develop as midlake bands in north-westerly flow over Lake Huron, farther upwind (**Figure 3**, lower panel). They may lose their visible cloud characteristics over southern Ontario, but maintain the convergence zone so that they redevelop once they reach the lower lakes. If the wind is very light against areas of enhanced concavity, then lake vortices (Type V) may develop (**Figure 3**, upper panel). Phenomena thought to play roles in this type of lake-effect storm are stretching and tilting of vorticity from low-level convergence and vertical wind shear; differential diabatic heating; and synoptic scale vorticity and temperature advection. **Table 1** summarizes some of the differences between various aspects of multiple- and single-band lake-effect snowstorms in western Michigan (lower peninsula).

The most common type of lake-effect event over the northern lakes (Superior, Huron, and Michigan) is the multiple-band variety, which occurs 50% of the time. The next most common is the shore-parallel or midlake variety, which occurs 25% of the time. Effects occurring during the remaining 25% include mesoscale vortices, hybrid combinations, and undeterminable forms. The most common type of lake-effect event over the southern lakes (Erie and Ontario) is the shore-parallel or midlake variety, because of the different orientation of these lakes (**Figure 4**). In general, across the region, the frequency of multiple bands decreases from west to east and the frequency of shore-parallel and midlake bands increases from west to east.

## Boundary Layer Dynamics

When cold air flows across the warm waters of the Great Lakes, strong sensible and latent heat fluxes warm and moisten the air closest to the surface first, causing the lowest levels of the atmosphere to destabilize. Strong turbulent motions mix upward the warmed and moistened air in convective fashion. Steam fog typically develops and steam devils may be visible near the surface, especially within a few tens of kilometers of fetch (**Figure 5**). A very unstable convective internal boundary layer (CIBL) forms near the surface and grows rapidly upward in the downwind direction (**Figure 6**). The upwind temperature, humidity, and flow characteristics of the air before it reaches the lake determine how the air will be modified. As the air crosses the downwind lakeshore, frictional convergence enhances ascent.

Convective updrafts can exceed $4–5\,m\,s^{-1}$ in narrow cores 100 m wide. After only a few kilometers of fetch the depth of the internal boundary layer may
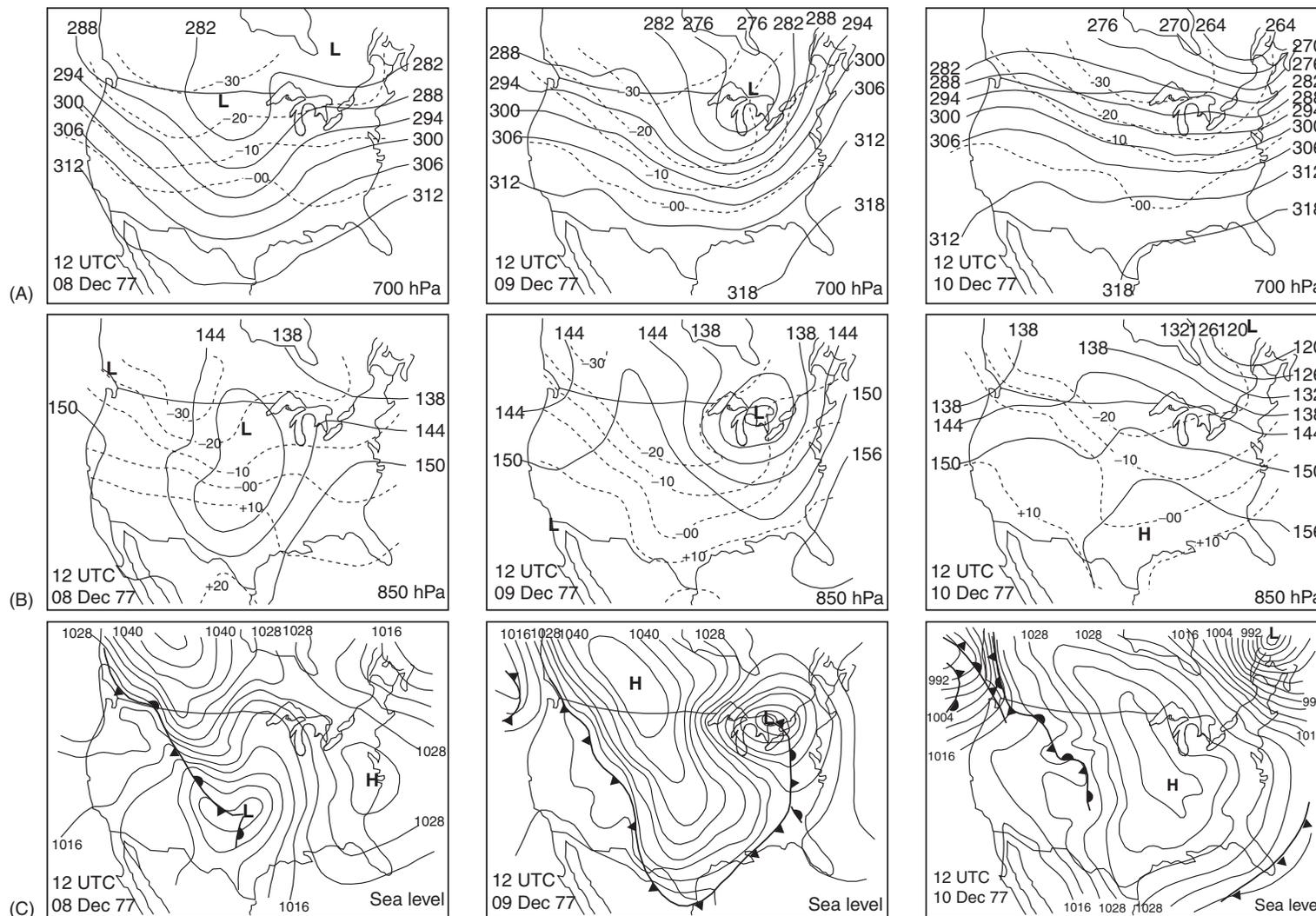
**Figure 2** Surface and upper air analyses depicting typical synoptic setting for lake-effect snow in the Great Lakes. (A) 700 hPa heights (solid, dm) and temperatures (dashed, °C); (B) 850 hPa heights (solid, dm) and temperatures (dashed, °C); (C) sea-level pressure (solid, hPa) and locations of highs, lows, and fronts for times shown. ((A) From Figure 6 and (B,C) from Figure 5 in Niziol TA, Snyder WR and Waldstreicher JS (1995) Winter weather forecasting throughout the eastern United States. Part IV: lake effect snow. *Weather and Forecasting* 10: 61–77. Adapted with permission from the American Meteorological Society.)
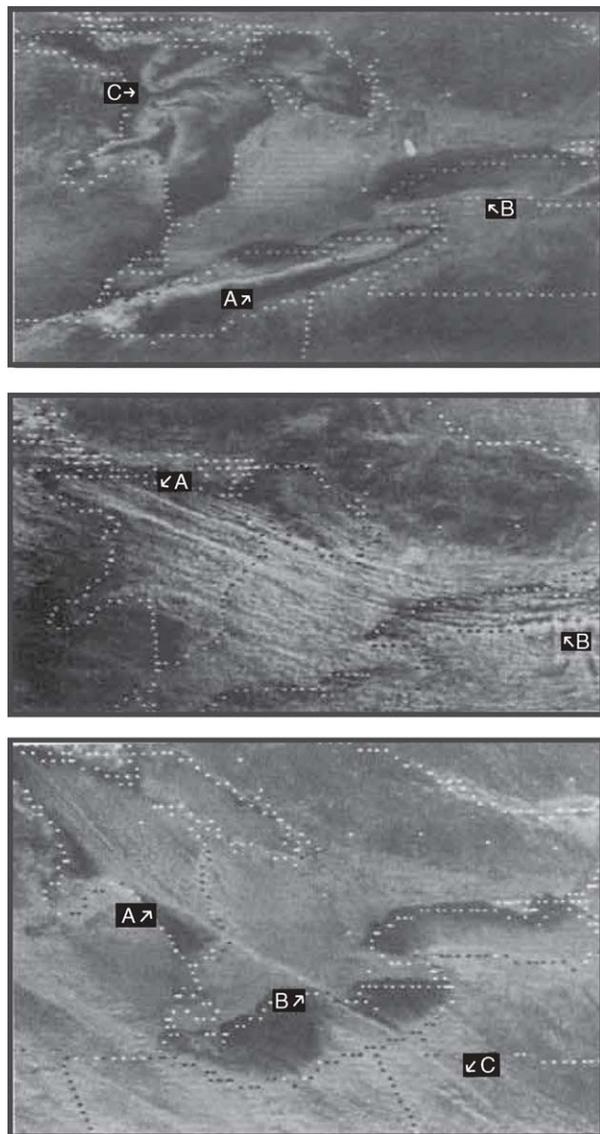
**Figure 3** Lake-effect snowband structures. Upper panel shows a lake vortex (Type V) over Lake Huron (C); midlake band (Type I) over Lake Erie (A); shore parallel band (Type IV) over Lake Ontario (B). Middle panel: multiple bands (Type II) over Lake Huron (A) and shore parallel band (Type IV) over Lake Ontario (B). Lower panel: midlake band (Type I) over Lake Huron (A); hybrid band (Type III) over Lake Erie (B and C). (From Figure 2 in Braham, RR Jr (1983) The midwest snow storm of 8–11 December 1977. *Monthly Weather Review* 111: 253–272. Adapted with permission from the American Meteorological Society.)

reach or exceed the depth of the planetary boundary layer. At some point the air within the thermal internal boundary layer becomes moist enough and deep enough for the lifting condensation level to be below the top of the boundary layer and cloud to form (Type I boundary layer). Subsequent latent heat release and radiative heat transfer become important and increase the rate of entrainment. The clouds develop initially as two-dimensional bands, but in time, with sufficient fetch and heating, these bands can develop into chains of three-dimensional cells and may eventually evolve into a regular array of three-dimensional mesoscale cellular convection. Maximum cloud drop concentration and liquid water content (e.g., $0.25\,\mathrm{g\,m^{-3}}$) occur near cloud-top and increase over the first two-thirds of fetch. Near the downwind lakeshore, snow particle production increases (e.g., $5\,\mathrm{L^{-1}}$) to reduce drop concentration and increase the height of cloud base.

The most common type of convection associated with cold air outbreaks over the Great Lakes is the longitudinal roll. This occurs in other parts of the world – wherever cold air crosses warm water. Over the Great Lakes, the rolls are usually oriented parallel to the wind at the base of the inversion and also to the low-level wind shear. Classical rolls are typically 2–5 km wide and 0.5–2 km deep, so the aspect ratio is roughly 2 to 5. Rolls over the Great Lakes are 10–20 km wide and 1–2 km deep, so the aspect ratio is about 10. Some roll circulations exhibit a multiscale configuration so that the cloud streets are best developed when the wavelengths of the convection and rolls are in phase. Such multiscale phasing can generate cloud streets with aspect ratios of 10–20.

Linear analytic studies indicate that three different types of instability mechanisms may be responsible for rolls. First, inflection point instability theory suggests that a dynamical instability can develop in a (rotating) Ekman layer that is neutrally stratified if the cross-geostrophic wind component exhibits an inflection point. The most unstable wind profiles lead to rolls with an aspect ratio of 3. Rolls are best developed when they are oriented about $14°$ to the left of the geostrophic wind (in neutral conditions). Second, parallel instability theory involves the curvature of wind speed profiles parallel to the roll axes and the boundary layer mean wind shear. This type of instability mechanism generates rolls with aspect ratios that are twice as large as those from inflection point instability and closer to those observed during lake-effect conditions. Third, convective instability (i.e., an unstable boundary layer) in the presence of wind shear also leads to the formation of rolls that are parallel to the mean wind shear but with aspect ratios smaller than those from inflection point instability (e.g., around 2).

Understanding roll development during lake-effect situations is further complicated by the fact that large-eddy numerical simulations suggest that only convective instability is capable of generating rolls. Gravity waves generated within the stable air above the convective boundary layer by wind shear near the

**Table 1** Typical values for multiple and shore-parallel band lake-effect storms in western Michigan (lower peninsula)

| Characteristics | Multiple | Single |
|---|---|---|
| *Mesoscale features* | | |
| | | |
| Location | Inland western shore | Along western shore |
| Band orientation | NW–SE bands | N–S band |
| Band length | 30 to 90 km | 90 to 180 km |
| Band width | 2 to 5 km | 10 to 20 km |
| Band (boundary layer) depth | 1 to 2 km | 2 to 3 km |
| Aspect ratio | 5 to 10 km | 10 to 20 |
| Lake temperature | 0 to 10°C | 0 to 2°C |
| 1000 air temperature | 0 to $-10$°C | $-10$ to $-20$°C |
| 850 air temperature | $-5$ to $-15$°C | $-10$ to $-30$°C |
| 700 air temperature | $-15$ to $-25$°C | $-20$ to $-40$°C |
| 1000 winds | W–NW above 10 m s$^{-1}$ | NW–N at 0 to 10 m s$^{-1}$ |
| 1000-850 shear dir/speed | 30°/0–10 m s$^{-1}$ | 10°/0–5 m s$^{-1}$ |
| 1000-700 shear dir/speed | 60°/0–20 m s$^{-1}$ | 20°/0–10 m s$^{-1}$ |
| | | |
| *Cloud/precip characteristics* | | |
| | | |
| Precipitation | Moderate snow ($<$ 10 cm) | Heavy snow ($>$ 10 cm) |
| Cell movement | With prevailing wind | With prevailing wind |
| Echo persistence | Short ($<$1 h) | Long ($>$1 h) |
| Cloud base | 1 km | $<$1 km |
| Cloud top | 2 km | 3 km |
| Updrafts | $<$1 m s$^{-1}$ | $<$2 m s$^{-1}$ |
| Liquid water content | 0.2 to 1 g m$^{-3}$ | 0.5 to 2 g m$^{-3}$ |
| Cloud droplet radius | 2 μm | 5 μm |
| Cloud droplet concentration | 100 to 300 cm$^{-3}$ | 300 to 500 cm$^{-3}$ |
| Natural ice nucleus | 1 to 10 L$^{-1}$ | 1 to 10 L$^{-1}$ |
| Ice-to-water ratio | $10^{-6}$ to $10^{-4}$ | — |
| Ice crystal size | 1 to 4 mm | — |
| Ice crystal concentration | 1 to 10 L$^{-1}$ | — |
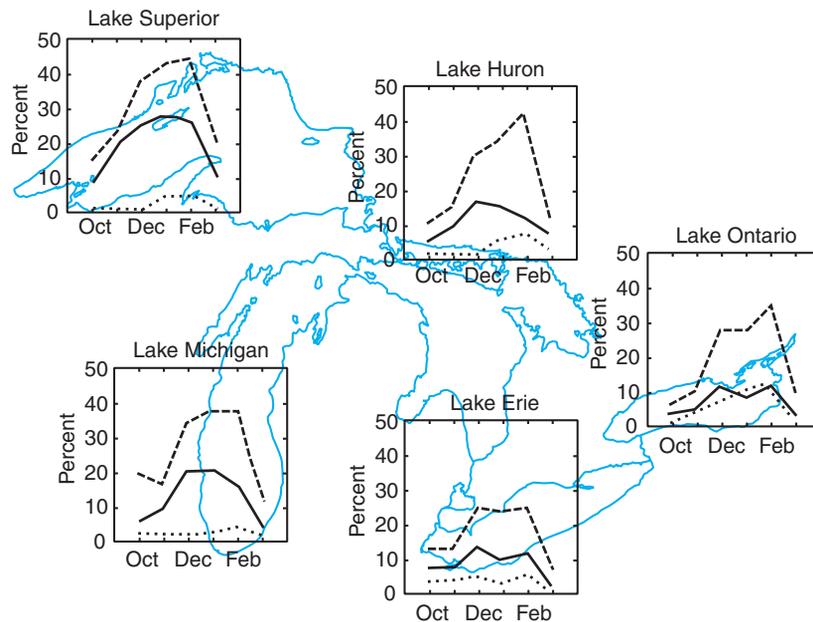


**Figure 4** Percentages of all days categorized as lake-effect (dashed), wind parallel bands (solid), and shore parallel bands (dotted). Results based on visible satellite imagery from 1988–93 (October–March). (From Figure 2 in Kristovich DAR, Steve III, RA (1995) A satellite study of cloud-band frequencies over the Great Lakes. *Journal of Applied Meteorology* 34: 2083–2090. Adapted with permission from the American Meteorological Society.)

**Figure 5** Example of steam fog and roll development during cold air outbreak over Lake Michigan. Photo taken at 14.25 UTC on 13 January 1998 during the Lake-Induced Convection Experiment. (Photo from David C. Rogers at NCAR taken over Lake Michigan during the Lake-ICE Experiment conducted during December 1997 to January 1998.)

boundary layer top have also been suggested as a possible mechanism. If the wind shear at the top of the boundary layer is roughly perpendicular to the boundary layer wind direction, then bands of convection parallel to the mean boundary layer wind can be induced by gravity waves. Latent heat release, cloud microphysical processes and low-level wind shear (e.g., below 200 m) may also influence the development of rolls. The partial agreement between observations and theoretical studies thus far suggests that a combination of mechanisms may be responsible for different aspects of roll development or during certain conditions.

## Sensitivity to Synoptic Conditions

The thermal modification of air over relatively cold land as it crosses a relatively warm lake results in a horizontal temperature gradient across the lakeshore,
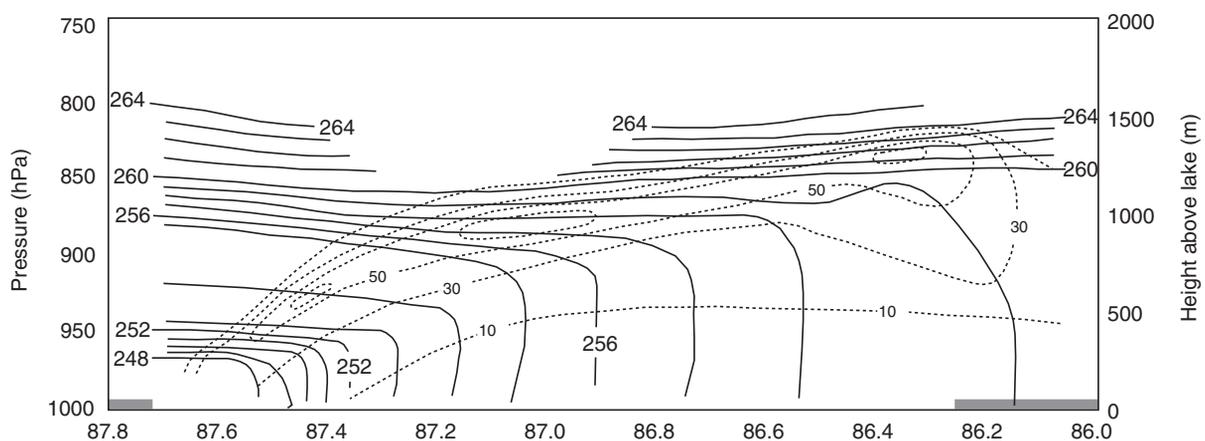


**Figure 6** Vertical cross-section of equivalent potential temperature (solid 1 K contour interval) and cloud frequency (dashed 10% contour interval) across southern Lake Michigan (from Sheboygan, WI, to Benton Harbor, MI) during a cold air outbreak (on 20 January 1984). Presence of cloud determined by cloud droplet concentrations greater than 10 cm$^{-3}$. Wisconsin and Michigan land surfaces are indicated by thick horizontal gray bars. (From Figure 5 in Chang SS, Braham RR Jr (1991) Observational study of a convective internal boundary layer over Lake Michigan. *Journal of the Atmospheric Sciences* 48: 2265–2279. Adapted with permission from the American Meteorological Society.)

which can generate a thermally direct solenoidal circulation similar to that of a sea or land breeze. In that sense, it can be seen that the lake–air temperature difference, speed and direction of the prevailing wind, and height and strength of the capping inversion are perhaps the most crucial parameters for determining not only how much lake-effect snow will fall, but how it will fall. For example, for conditions characterized by moderate lake–air temperature differences ($>10°$C), and strong winds ($>10\,\mathrm{m\,s^{-1}}$) blowing across the short axis of a lake, multiple snowbands will usually develop. If the winds across the short axis are lighter and/or the temperature difference is greater then some form of shore-parallel band will develop. If the wind is very light against areas of enhanced concavity then lake vortices can develop.

The impact of the wind speed on the lake-effect response characteristics can be analyzed two-dimensionally in terms of the Froude number $Fr = \pi U/NH$, where $U$ is the mean wind speed across the short axis of the lake, and $N$ and $H$ are the Brunt–Vaisala frequency and depth of the planetary boundary layer respectively. The values for $N$ and $H$ depend on the lake–air temperature difference and stability of the pre-lake-modified air. The Froude number may be interpreted as the ratio of the mean wind speed $U$ to the gravity wave speed $c_{\mathrm{g}} = NH/\pi$ for a boundary layer of depth $H$. Three regimes are important to consider. When $Fr < 1$ then the gravity wave speed exceeds the mean flow speed ($c_{\mathrm{g}} > U$). Opposing sea-breeze type circulations develop with respect to the short axis of the lake and the heaviest precipitation falls over the lake (**Figure 7A**). This regime corresponds to the midlake band type of event. When $Fr > 1$, the gravity wave speed is less than the mean flow speed ($c_{\mathrm{g}} < U$). The response is characterized by alternating regions of ascent and descent that propagate downwind from the leeward shore, and precipitation is diffuse and weak. This regime corresponds to the multiple-band type of event (**Figure 7B**). When $Fr = 1$, the gravity wave speed equals the mean flow speed ($c_{\mathrm{g}} = U$). A resonance condition develops where gravity waves generated at the downwind shore cannot propagate upwind. This regime corresponds to the shore-parallel band type of event that can generate significant precipitation at the downwind lakeshore (**Figure 7C**). Setting $Fr = 1$, and using typical values of $N = 10^{-2}\,\mathrm{s^{-1}}$ and $H = 2\,\mathrm{km}$, suggests that a value of $U \approx 6\,\mathrm{m\,s^{-1}}$ maximizes heavy snow along the downwind lakeshore, which is consistent with observations.

The impact of fetch on lake-effect storm development has been known since the early 1900s. Long fetches usually result in heavy snowfalls. Short fetches also may produce significant snowfalls if the pre-lake-modified air is relatively unstable, if the lake shore geometry enhances radial convergence, or if the nearby orography enhances lifting. These observed impacts of fetch have been confirmed only recently using analytic and numerical models. For example, it has been shown analytically for Lake Michigan that three convergence centers develop near the eastern shore when a westerly wind prevails, two cells or snowbands develop when a north-westerly wind prevails, and one midlake band develops when a northerly wind prevails.

The vertical structure of the environmental wind also affects lake-effect storms. For example, when the prevailing wind is parallel to the long axis of Lake Erie, moderate directional shear (e.g., between $30°$ and $60°$) from the surface to $700\,\mathrm{hPa}$ causes weakly to-moderately precipitating multiple snowbands rather than a single intensely precipitating snowband to occur. Stronger shear (e.g., greater than $60°$) over Lake Erie causes the breakdown of precipitating snowbands altogether – allowing only instead the development of a non-precipitating stratocumulus deck. While wind shear can understandably inhibit the organization of rolls or wider bands, it is not clear whether some of the observed effects from shear are simply a manifestation of a shallow boundary layer.

The height and strength of the (capping) inversion are significant limiting factors to cloud depth and therefore to precipitation. Typically, the boundary layer must have a depth greater than $1\,\mathrm{km}$ in order for lake-effect snow to develop. The most convectively active lake-effect storms have inversion heights exceeding $3\,\mathrm{km}$. Sometimes the capping inversion is entirely absent. During such cases, thunder and lightning typically accompany copious snowfall rates e.g., these exceeding $10\,\mathrm{cm\,h^{-1}}$.

The vertical temperature and moisture distributions within the boundary layer also play a role. For example, it has long been known that a minimum temperature difference of $13°$C between the lake surface and the upstream airflow at $850\,\mathrm{hPa}$ is required for lake-effect storms to develop. This temperature difference criterion means that the lapse rate should be unstable with respect to unsaturated ascent. A dry boundary layer is less conducive for lake-effect snow than a moist one, although a long fetch can compensate for very dry boundary layers. The impact of moisture is greater for low-stability profiles than for high-stability ones.

The presence of large-scale forcing can also influence lake-effect storm development. Typically, the coldest air passes over the lakes as high pressure at the surface moves eastward across them, accompanied by negative vorticity advection at upper levels and cold advection near the surface. Thus, the impacts of synoptic-scale forcing typically act to suppress
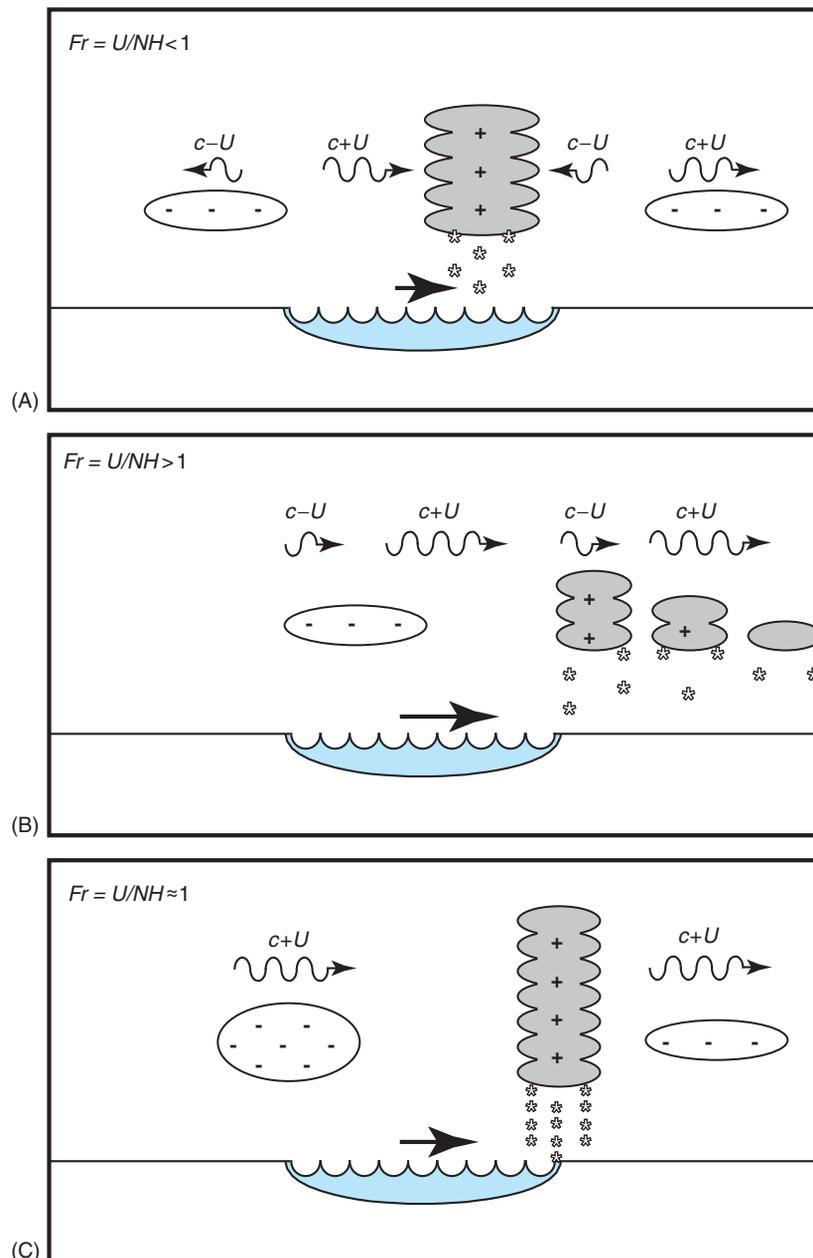
**Figure 7**    Gravity wave interpretation of lake-effect morphology dependence on wind speed. (A) Weak windspeeds create subcritical ($Fr < 1$) regime, which allows gravity waves to propagate upwind and downwind and a midlake band and moderate snow to develop over the lake; (B) strong windspeeds create supercritical ($Fr > 1$) regime, which allows gravity waves to propagate only downwind and multiple bands and light snow to develop beyond the downwind lakeshore; (C) moderate windspeeds create near-critical ($Fr \approx 1$) regime, which allows gravity waves to travel downwind but traps gravity waves trying to propagate upwind – resulting in an intense shore-parallel band and heavy snow to develop along the downwind lakeshore. Heavy arrows indicate windspeed and wavy arrows indicate gravity wave propagation. Plus ($+$) signs and shaded columns indicate ascent; minus (–) signs and open ovals indicate descent. Asterisks indicate snowfall.

lake-effect storm development (**Figure 2**). There are however instances when cold air, positive vorticity advection, and even warm advection exist simultaneously over the region. Such situations usually come in the form of Alberta Clippers (short waves) that develop in cold air masses and move south-eastward across the region. The synoptic forcing coupled with the cold air that is already established over the region can combine to generate intense snowfall.

Sensible and latent heating from all the Great Lakes (e.g., the lake aggregate) can also influence lake-effect storms over individual lakes. Basically, if warming

(and moistening) occurs over all the Great Lakes for at least a day then surface pressures and stability can drop over a broad region and cause a perturbation aggregate-scale, low-level cyclonic circulation to develop. The position, the size, and the warmth and moisture from this aggregate circulation can modify lake-effect precipitation throughout the region. Specifically, when the synoptic-scale flow is north-westerly, aggregate effects can augment snowfall along the north-western shores of lower Michigan, and reduce snowfall along the south-western shores (**Figure 8**). Shore-parallel bands located offshore can migrate eastward (e.g., onshore) or evolve into multiple bands. These aggregate affects over Lake Michigan include enhanced westerly flow, increased heat and moisture, and lower stability. The lake aggregate can also influence lake-effect precipitation in the lower lakes region. For example, as the lake-aggregate-induced plume of heat and moisture extends south-eastward, surface winds across Lake Ontario (north of the aggregate induced plume) can become more northerly.

In contrast, surface winds across Lake Erie (south of the aggregate-induced plume) can become more westerly. The aggregate-altered winds can cause a longer fetch across Lake Erie and a shorter fetch across Lake Ontario, and can shift the regions of lake-effect convective bands, so that less (intense) lake-effect precipitation can fall along the lakeshores downwind (east) of Lake Ontario and more lake-effect precipitation can fall along the eastern shores of Lake Erie (**Figure 8**).

## Forecasting

The mesoscale nature of lake-effect storms, their intensity, and the short development times continue to challenge forecasters. Highly variable snow-to-liquid ratios (10:1 to 50:1) and terrain effects, especially near Lakes Erie and Ontario, can enhance the inherently large spatial variability of lake-effect snow and hence the forecast challenge. While the problems
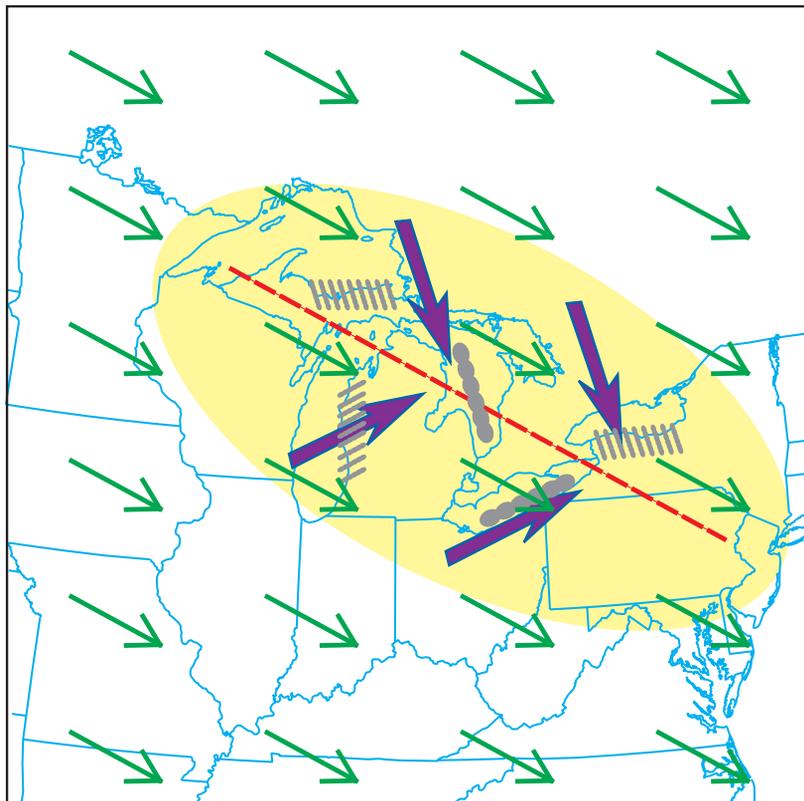


**Figure 8**    Illustration of lake aggregate effect on prevailing winds and lake-effect snowstorms. On the south side of the developing warm plume (shaded oval), north-west winds respond in sea-breeze fashion to become south-west winds with increased fetch and heavy snow across Lake Erie. Lake effect snows across portions of western Michigan (lower peninsula) may or may not change characteristics. On the north side, north-west winds respond to become north-north-westerly winds with reduced fetch and light snow across Lakes Superior and Ontario and increased fetch heavy snow across Lake Huron.

of forecasting when lake-effect snow is going to occur have essentially been solved, the equally significant problems of exactly where lake-effect snow will occur, what form(s) it will take, how intense it will be, and how long it will last remain outstanding forecast issues. A combination of high-resolution numerical weather prediction models, statistical methods, Doppler radar, and forecaster savvy are the basic forecast tools.

Numerical models have come a long way since the use of the limited fine mesh (LFM) model. The horizontal grid spacing (180 km) and the exclusion of the lakes in terms of their heat, moisture, and momentum characteristics in that model precluded any explicit model development of lake-effect precipitation. Regardless, operational forecasters relied on this model because of its ability to forecast the large-scale conditions to which lake-effect snowstorm development is very sensitive. The LFM model, in conjunction with forecaster decision trees based on key large-scale parameters, and experience, allowed forecasters to at least be able to issue general forecasts of when lake-effect snow was going to occur.

Two operational models currently being used include the nested grid model (NGM), with 48 km horizontal grid spacing, and the Eta model. Several different versions of the Eta model are run at several different resolutions and times including one run at 12 km horizontal grid spacing four times daily. The increased resolution in the Eta model has been especially helpful for identifying areas where lakeshore enhanced snowbands may develop. Recently,



Lake-effect guidance: Ontario
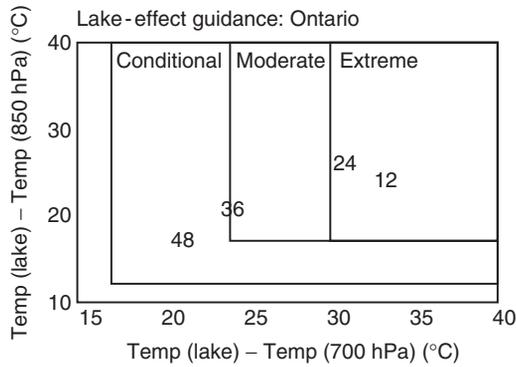
12 h extreme instability...030 degree shear from 700 hPa to SFC...and 090-mile fetch at 850 hPa.

24 h moderate instability...020 degree shear from 700 hPa to SFC...and 070-mile fetch at 850 hPa.

36 h moderate instability...000 degree shear from 700 hPa to SFC...and 090-mile fetch at 850 hPa.

48 h moderate instability...090 degree shear from 700 hPa to SFC...and 150-mile fetch at 850 hPa.

Output from NGM, 12Z 18 Feb 93: Forecast parameters

| Variable | 12 h | 24 h | 36 h | 48 h |
|---|---|---|---|---|
| Wind direction (degrees) | | | | |
| 700 hPa | 260 | 310 | 290 | 270 |
| 850 hPa | 290 | 300 | 290 | 240 |
| B.L. | 290 | 290 | 290 | 180 |
| Change in wind direction with height | | | | |
| 850/700 | 030 | 010 | 000 | 030 |
| B.L./700 | 030 | 020 | 000 | 090 |
| Fetch (miles) | | | | |
| 850 hPa | 090 | 070 | 090 | 150 |
| Temp (°C) | | | | |
| 700 hPa | −30 | −28 | −21 | −18 |
| 850 hPa | −21 | −23 | −17 | −14 |
| B.L. | −10 | −14 | −10 | −09 |
| Lake | 02 | 02 | 02 | 02 |
| TS/T3 layer inversion intensity | | | | |
| TS-T3 | −08 | −08 | −03 | −05 |
| Vertical velocity (microbar/s) | | | | |
| 700 hPa | +02 | −02 | −02 | +00 |

**Figure 9** Lake-effect snow guidance product for 1200 UTC on 18 February 1993 generated for Lake Ontario at WSFO Buffalo. TS and T3 correspond to the surface and 900 hPa temperatures respectively and BL corresponds to the boundary layer. (From Figure 10 in Niziol TA, Snyder WR and Waldstreicher JS (1995) Winter weather forecasting throughout the eastern United States. Part IV: Lake effect snow. *Weather and Forecasting* 10: 61–77. Adapted with permission from the American Meteorological Society.)

several forecast offices have experimented with running locally high-resolution mesoscale models. The Forecast Office in Buffalo, New York, has been running a 10 km version of the PSU/NCAR model MM5 since 1996. The Forecast Office in Detroit, Michigan, has been running a 6 km version of the Eta model since 1998. Both of these offices have reported the ability to provide more specific and more accurate forecasts.

Despite significant and continuing improvements in numerical weather prediction, lake-effect snow continues to challenge the abilities of even the most sophisticated numerical models, because of several inadequacies. These inadequacies include horizontal resolution that is still too coarse for resolving the 2–4 km wide bands, convective schemes tuned originally for deep (tropical) convection that are inappropriate to simulate intense shallow precipitating convection, and boundary layer schemes that are too simplistic to develop the low-level temperature, moisture, and cloud-microphysical structures that exist within lake-effect snow environments near the surface.

To address some of these inadequacies, forecasters currently use various statistical methods. These methods were used almost exclusively prior to the existence of numerical models. As early as the middle of last century, various investigators had outlined conditions necessary for prolonged lake-effect storms to occur at the eastern end of Lake Erie. Afterwards, more sophisticated statistical models, based on multiple discriminant analysis, the perfect prog (PP) method, model output statistics (MOS), and classification and regression trees (CART), were developed for many of the lake-effect snow belts. Currently, the use of numerical model output in terms of larger-scale features, coupled with highly tuned, sophisticated statistical models, has proven a very effective forecast method (**Figure 9**).

Remaining challenges for numerical lake-effect snow forecasting include resolving the lakeshore geometry and nearby terrain, simulating more accurately the evolution of the boundary layer – including heat, momentum, and moisture fluxes – and simulating more accurately the convective precipitation. A fourth challenge is initializing more accurately the lake surface temperatures, which are specified currently using AVHRR satellite data that represent a multiday average and may have gaps because of persistent cloudiness. Finally, the simulation of subsequent changes in lake surface temperatures, which may also improve forecast accuracy, has yet to be included.

## See also

**Air–Sea Interaction:** Storm Surges. **Boundary Layers:** Convective Boundary Layer; Modeling and Parameterization; Overview. **Climate:** Overview. **Convective Storms:** Convective Initiation; Overview. **Hurricanes**. **Mesoscale Meteorology:** Mesoscale Convective Systems; Models. **Numerical Models:** Methods. **Synoptic Meteorology:** Forecasting. **Weather Prediction:** Regional Prediction Models.

## Further Reading

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