Preliminary Numerical Simulations of the 31 January 1991 Lake-Effect Snowstorm

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ABSTRACT

A three-dimensional numerical model is used to investigate the relative importance of factors that influenced the development of the major lake-effect snowstorm of 31 January 1991 in Central New York. Sensitivity experiments indicate that the major controls on the snowfall rate are lake temperature and the moisture content of the air upwind of Lake Ontario. As expected, the location and movement of the snowband depend mainly on the geostrophic wind direction.

INTRODUCTION

When cold air moves over the warmer Great Lakes in late Fall and winter, bands of heavy snow often form in the region downwind of the lakes. The bands are narrow (5 to 25 km wide usually) and elongated (50 to 300 km long typically). The snowbands can be especially troublesome because visibility may be reduced to zero in the heavy fall of low-density snowflakes. The goal of this research is to learn more about the mechanisms that control the development and life-cycle of lake-effect snow bands by using numerical simulation.

MODEL DESCRIPTION

An earlier version of the model was described by Ballentine (1980). In the present version, height z above sea level is used as the vertical coordinate and flat terrain is assumed. While the authors recognize that terrain effects are important in some lake-effect snowstorms, the main goal of this research is to determine the sensitivity of snowband development to changes in meteorological conditions and lake surface temperature. Furthermore, the 31 January 1991 snowband was located over the nearly flat terrain between Lake Ontario and the western Mohawk Valley for most of its duration. The model domain covers the Lake Ontario region with a 10 km-by-10 km grid extending 440 km in the east-west direction and 300 km in the north-south direction. Lake Erie and smaller bodies of water, such as the Finger Lakes, are omitted.

A crude form of synoptic-scale forcing is included in the model through the horizontal pressure gradient force. The pressure (Exner function) gradient consists of a large-scale, slowly-varying part proportional to the geostrophic wind, and a more rapidly-varying part computed from the potential temperature predicted by the model. In order to allow the model snowband to respond to temporal changes in the large-scale flow, the geostrophic part of the pressure gradient is specified at each time step by interpolating linearly in time the geostrophic winds estimated from the 0000 UTC and 1200 UTC Buffalo soundings. In most of the major lake-effect snowstorm cases that we have studied so far, the assumption of a linear change in geostrophic wind is justified during the time period after cold front passage.

Lateral boundary conditions for humidity, temperature, and wind need to be provided only on the inflow boundaries since the upstream method is used to solve the governing equations. Boundary tendencies for temperature are computed from the 12-hour changes observed on mandatory levels at the nearest upper-air stations. Specific humidity and boundary layer height are held fixed on the upwind boundaries, and the wind components vary according to the geostrophic tendencies specified from the Buffalo soundings.

JANUARY 31, 1991 SNOWSTORM

A low pressure system deepened explosively as it moved from southeastern Pennsylvania at 0000 UTC 31 January 1991 to the Canadian Maritimes 24 hours later. The storm produced light snow over eastern New York, but no precipitation in central New York. However, the deepening low helped maintain a strong, moist west-northwesterly airflow over Lake Ontario.

Fig. 1 shows that the atmosphere was cold and nearly saturated up to about 660 hPa.

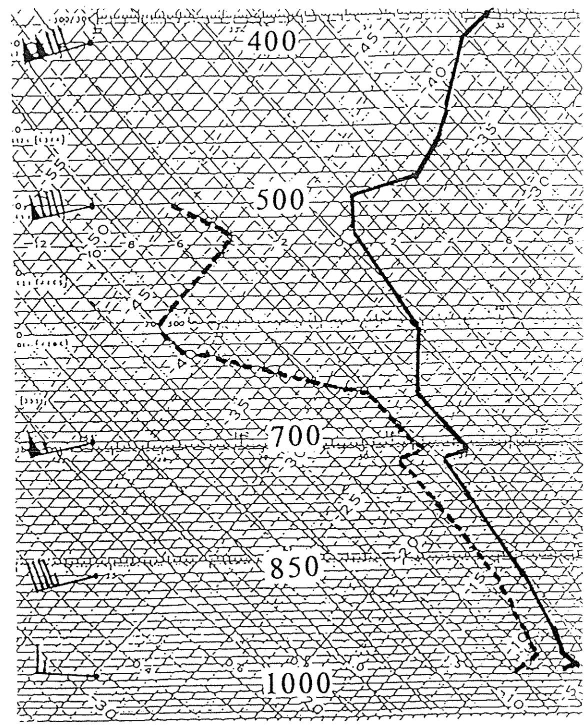

Fig. 1. Buffalo, NY sounding for 12 UTC 31 January 1991. Temperature (solid); dew point (dashed). For wind., flag equals 25 m/s; full barb equals 5 m/s.

Assuming a lake temperature of 1 C and 850 hPa temperature of -14 C, the temperature difference between the lake surface and the air at 850 hPa exceeded the 13 C criterion usually regarded as a necessary condition for heavy lake-effect snowfall in the Great Lakes region (Niziol, 1987).

The snowband formed over northern Oswego County. As the southern edge of the band moved over Nine Mile Point, the wind speed (15-minute average) at the

60-meter tower increased from 3 knots at 1045 UTC to 30 knots at 1145 UTC. The wind at 60 meters continued to blow from about 280 degrees at speeds averaging about 30 knots for most of the morning. The rapid increase in wind speed suggests that there was a strong pressure gradient near the southern edge of the band.

Heavy snow was accompanied by thunder and lightning when the band reached SUNY Oswego at 1100 UTC. Trained student observers reported zero visibility in heavy snow during the 5 hours ending at 2000 UTC. The snow continued until 2130 UTC after which the band moved southward toward Syracuse. Satellite photographs indicate that the band was about 20 km wide and about 300 km long extending eastward from the south-central portion of the lake to the Mohawk Valley. The cloud temperature at the top of the band was about -31 C, based on infrared satellite imagery, which would put the top of the band at about 590 hPa, or about 4100 m. The band was observed by the National Weather Service radar at Binghamton, NY about 130 km to the south.

The banded structure of the precipitation is evident in Fig.2. The snowfall and water equivalent precipitation at locations marked by a one or two letter abbreviation on the map are as follows:

Station	Abbrev.	Snow	Precip.
Bennett's Br.	BB	40.6 cm	11.4 mm
Boonville	B	68.6	35.6
Camden	C	56.9	18.0
Highmarket	H	50.0	14.5
Newport	N	33.0	20.8
Oswego East	OE	39.9	20.3
Rochester	RR	10.1	2.5
Rome	R	29.0	29.0
Syracuse	S	19.6	4.1

Snowfall-to-liquid water ratios varied considerably reflecting the difficulty and lack of consistency

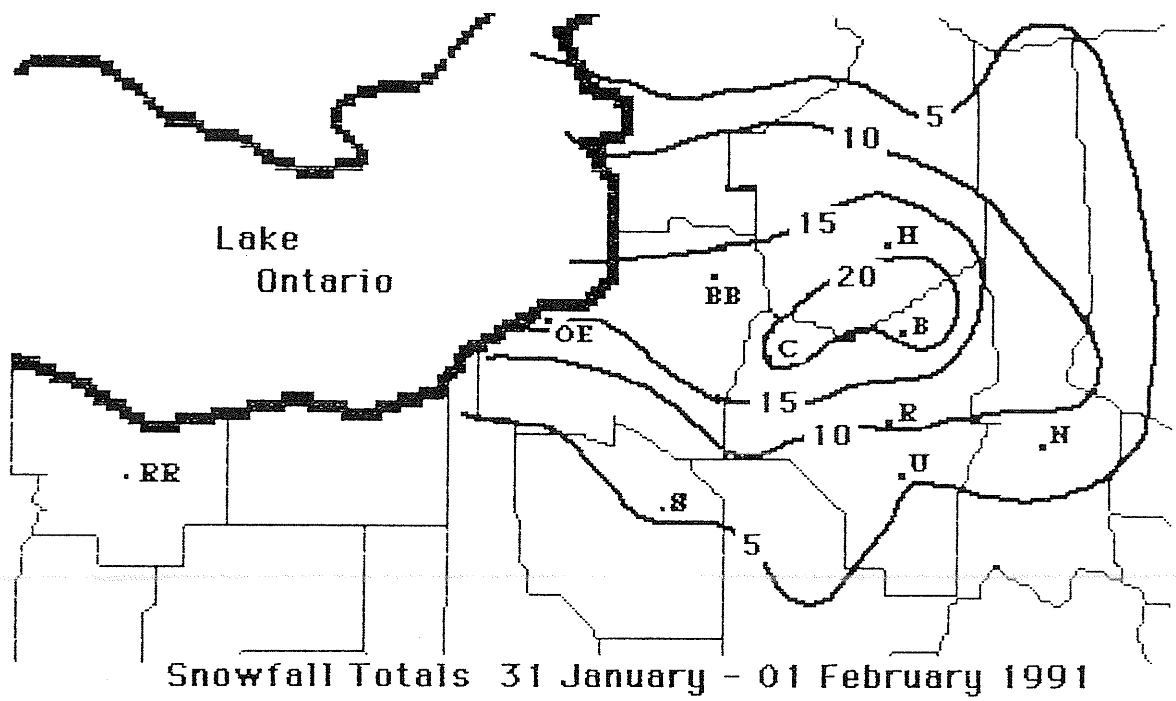

Fig. 2. Observed snowfall in central New York for 31 January 1991. Courtesy Northeast Regional Climate Center, Cornell University.

in measuring lake-effect snowfall. The area of observed precipitation is wider than the band itself because the band moved southward during the day.

INITIAL CONDITIONS

Initial conditions for the model are based on soundings at Buffalo, NY. Since the snowband formed before 1200 UTC 31 January, the initial time for all runs is assumed to be 0900 UTC. The 0900 Buffalo 'sounding' is computed by linear interpolation of the 0000 and 1200 UTC soundings. Humidity, wind, and temperature are interpolated to the 15 prediction levels of the model. For simplicity, the initial data is assumed to be horizontally homogeneous. Although this is convenient, it violates the thermal wind law since vertical wind shear is present in the Buffalo sounding.

One sensitivity test is included to determine the effect of allowing baroclinity in the initial data consistent with the thermal wind law. The thermal wind equation is

$$\partial \vec{V}_g/\partial z = (g\hat{k} \times \nabla\theta + f\vec{V}_g \partial\theta/\partial z)/f\theta$$

where $\vec{V}g$ is the geostrophic wind, g is gravity, f is the Coriolis parameter, and θ is potential temperature. Using representative horizontal temperature gradients based on the 1200 UTC surface and upper-air charts, and using the 1200 UTC Buffalo sounding to estimate the vertical gradient of potential temperature, it may be shown that the second term on the right is more than one order of magnitude smaller than the first term and may be omitted. Solving for the gradient of potential temperature gives:

$$\partial\theta/\partial x = \frac{f\theta(v_{gu} - v_{gl})}{g(z_u - z_l)}$$

$$\partial\theta/\partial y = -\frac{f\theta(u_{gu} - u_{gl})}{g(z_u - z_l)}$$

In the sensitivity test involving the effect of initial baroclinity, the temperature distribution is computed at each model level using bilinear interpolation in the x and y directions. The vertical gradients of geostrophic wind are computed using the geostrophic wind (estimated from the Buffalo sounding) at the two closest heights surrounding each model level. Since the imposed baroclinity lowers temperature at some grid points (mainly to the north and west of Buffalo), the initial water vapor content must be adjusted to avoid supersaturation. The simplest solution is to assume that the relative humidity is horizontally homogeneous at each level.

CONTROL EXPERIMENT

The model inputs for the control experiment (summarized in Table 1) are chosen to reflect conditions observed just before the snowband formed. The initial (0900 UTC 31 January) and final (0000 UTC 01 February) geostrophic winds are derived from the Buffalo soundings corrected in an attempt to remove the effects of friction.

Table 1. Input data for control experiment

Level	Height (m)	Theta deg K	RH %	Initial Wind (kts) Direction	Speed	Final Wind (kts) Direction	Speed
15	8400	324.1	2	250	131	280	115
14	7200	311.4	3	250	137	285	104
13	6000	298.0	3	255	106	295	92
12	4800	284.7	47	255	80	301	78
11	3700	281.7	28	260	69	310	65
10	2800	277.5	81	260	54	310	56
9	2100	274.5	84	260	48	304	44
8	1560	273.2	87	260	48	296	37
7	1130	272.2	88	268	49	288	40
6	790	271.1	89	279	46	283	40
5	520	270.3	89	287	40	280	39
4	310	269.6	89	295	30	280	36
3	160	269.2	88	297	27	280	33
2	60	269.1	80	299	26	280	31
1	10	268.5	89	300	25	280	30

Surface Temperature:	Lake = 1 C	Land = -5 C
Surface Roughness:	Lake = .001 m	Land = .1 m
Mixed Layer Height:	Lake = 2500 m	Land = 1500 m

The first trace of precipitation from the model snowband falls about 2 hours after the start of the simulation. The snow begins simultaneously all along a 150 km long line stretching from the south-central portion of the lake to just east of Oswego. Fig. 3 shows the predicted vertical velocity at 1800 UTC during a period of 'whiteout' conditions at Oswego. Maximum upward motion is 68.6 cm/sec at a point about 40 km east of Nine Mile Point. A narrow core of strongest upward motion (exceeding 40 cm/sec) is located east of the lake from Nine Mile Point to near Boonville, NY. The model band is about 20 km wide and more than 300 km long giving it a shape similar to that shown in the satellite imagery. By 0000 UTC on the 1st, the model snowband (as shown in Fig. 4) has weakened considerably and moved southward. The precipitation predicted in the 15 hours ending 0000 UTC on the 1st is shown in Fig. 5. The core of heaviest snowfall (more than 10 mm water equivalent) extends from near Nine Mile Point eastward to Boonville with the greatest amount (20.1 mm) predicted at a point 30 km east of Nine Mile Point. Model snowfall ended in this region as the model snowband drifted southward after 2000 UTC in response to veering of the geostrophic wind. Nearly all of the precipitation south of the main core fell after 1800 UTC as the model snowband was weakening.

The model underestimates the amount of precipitation especially at locations far inland from the lake. One reason may be that the horizontal grid size (10 km) is only a little smaller than the width of the snowband. Finer resolution would yield stronger vertical motions and a greater condensation rate. Neither cumulus

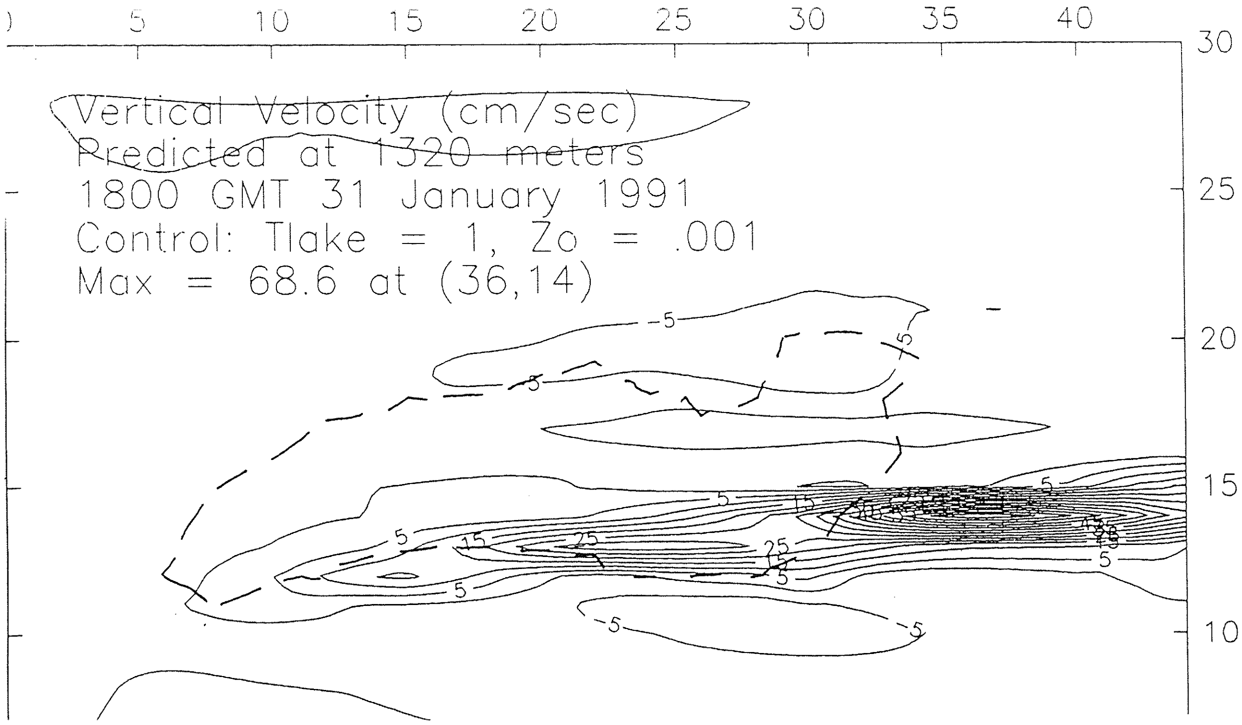

Fig. 3. Predicted vertical velocity (cm/s) at 1800 UTC 31 January 1991 for control experiment. Contour interval equals 5 cm/s.

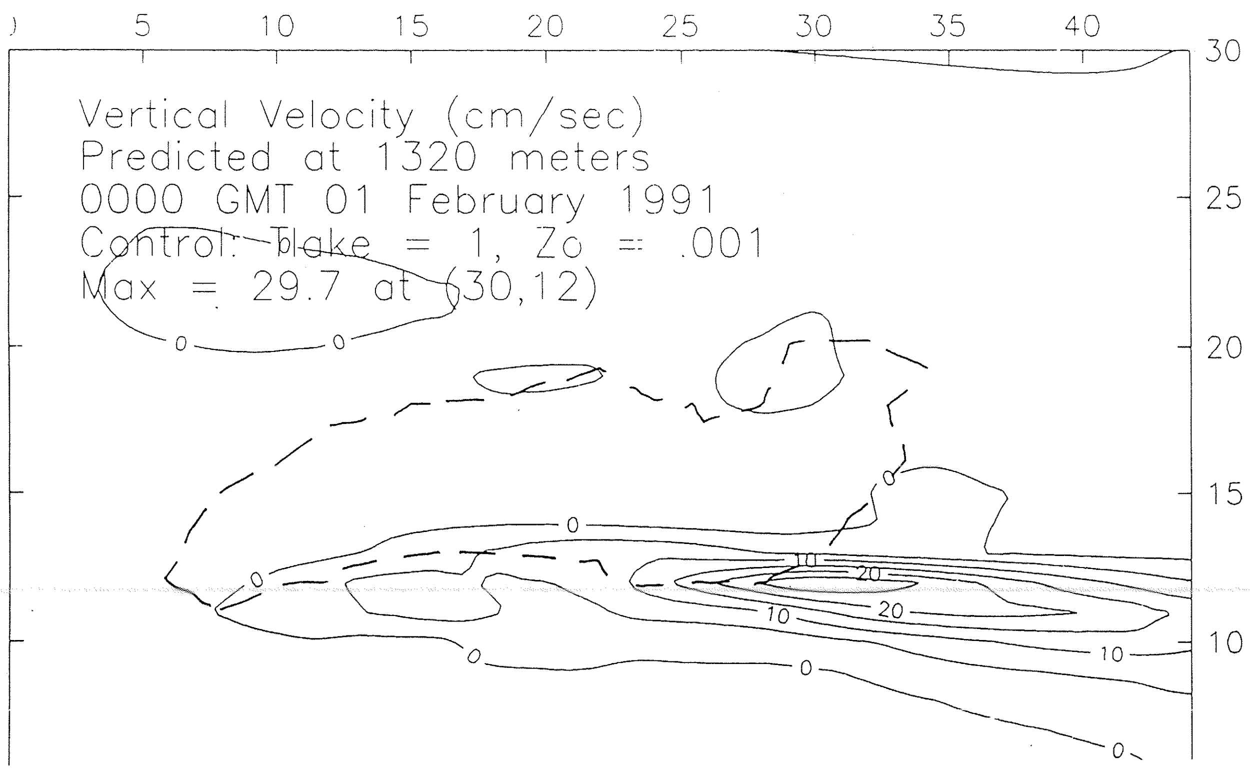

Fig. 4. Same as Fig. 3 except 0000 UTC 01 February 1991.

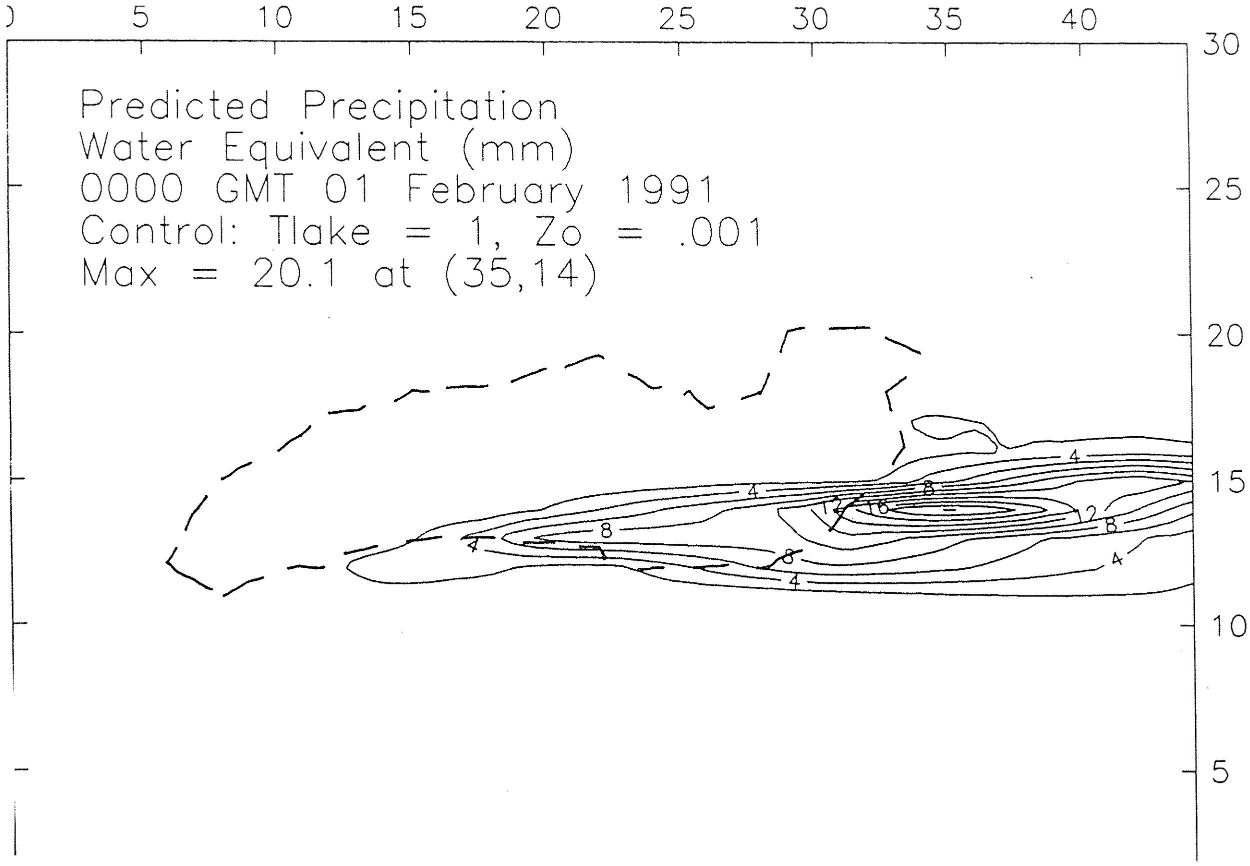

Fig. 5. Predicted precipitation (mm) for the 15 hours ending 0000 UTC 01 February 1991 for control experiment. Contour interval 2 mm.

transports nor convective precipitation are included in the present version of the model even though convective processes probably play an important role in many lake-effect snowstorms. Furthermore, no allowance is made for the inland drift of snowflakes as they fall which would alter the distribution of snowfall in favor of more inland locations (Lavoie, 1987).

SENSITIVITY EXPERIMENTS

The sensitivity experiments are listed in Table 2. Each run is identical to the control except for the single change described in Table 2.

In the warm lake experiment, the distribution of precipitation (not shown) is similar to that in the control run, but the maximum amount of precipitation is nearly two-and-one-half times greater. Apparently, an increase of only 2 degrees C in lake temperature can have a very large effect on the precipitation rate.

In the cold land experiment, the difference in temperature between the lake and the land is increased from 6 to 12 degrees C. The model snowband forms about 10 km farther south and produces about 30% more precipitation. While there is no obvious explanation for the increase in precipitation, it might be related to the 30% increase in surface pressure gradient between the lake and the

Table 2. List of sensitivity experiments

No.	Name	Description
1	Warm Lake	Lake temperature increased from 1 C to 3 C
2	Cold Land	Land temperature decreased from -5 C to -11 C
3	Warm Air	Air temperature increased by 5 C
4	Cold Air	Air temperature decreased by 5 C
5	Dry Air	Relative humidity decreased by 30 per cent
6	Reduced Wind	Prevailing wind speed reduced by 50 per cent
7	Baroclinic	Initial temperature computed via thermal wind

land noted in this experiment due to the colder air over land.

In the warm air experiment, the initial potential temperature is increased by 5 C at every grid point. As a result, Tlake-T850 mb decreases from 15 C to 10 C. The model produces much less upward motion than in the control, and the maximum precipitation is only 3.9 mm (one fifth as much as in the control). The snowband forms about 10 km south of the control position. Even though the initial (warmer) atmosphere contains more water vapor, the air is just too stable to allow the formation of a significant snowband circulation.

In the cold air experiment, the initial potential temperature is decreased by 5 C at every grid point. As a result, Tlake-T850 mb increases from 15 C to 20 C. The model vertical velocities are much stronger (maximum 126.6 cm/sec) than in the control, but the maximum precipitation is only about 25 per cent greater. This is probably because there is less water vapor in the initial (colder) atmosphere in this case. The snowband forms about 30 km farther north than the control position, and the maximum precipitation is displaced farther inland.

In the dry air experiment, the initial relative humidity is decreased by 30% at every grid point. The location and movement of the band are almost identical to the control. The maximum vertical velocity (56.4 cm/sec) is only 20% less than in the control, but the maximum precipitation (5.9 mm) is less than one-third of the control amount. Apparently, the water vapor content of the air mass moving over the lake is a major control on the precipitation rate.

In the reduced wind speed experiment, the initial and final geostrophic winds are decreased by 50 per cent at every grid point. The predicted precipitation (Fig. 6) is concentrated at grid points closer to the lake. The model snowband weakens sooner in this experiment, although the maximum snowfall, predicted near Oswego, is slightly greater than in the control run.

In the baroclinic experiment, horizontal temperature gradients derived from the thermal wind law, are used to construct the initial temperature field. Since no attempt was made to eliminate superadiabatic layers, some unrealistic temperature profiles were introduced in the initial data. Although vertical diffusion in the model was able to remove the superadiabatic layers within the planetary boundary layer, this simulation was much noisier than any of the others. The maximum vertical velocity (99.0 cm/sec) is about 20% greater and the maximum precipitation (31.2 mm) is about 50% greater than in the control experiment.

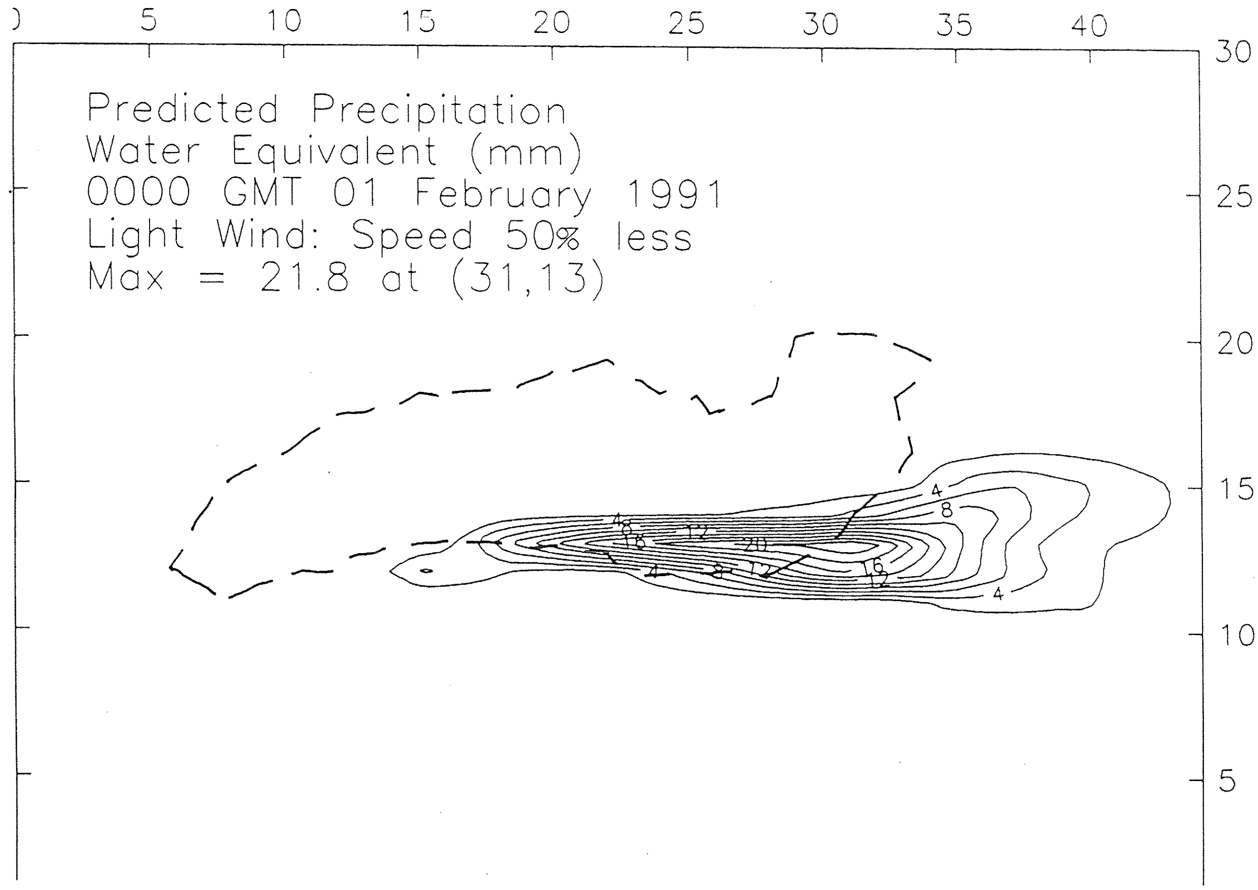

Fig. 6. Same as Fig. 5 except for reduced wind experiment.

SUMMARY AND CONCLUSIONS

The numerical model was successful in simulating the location, movement, and change in intensity of the 31 January 1991 lake-effect snowband. The predicted precipitation for the control experiment was somewhat less than observed, although there is some uncertainty with regard to the measurement of lake-effect snowfall.

The sensitivity experiments show that the lake temperature, moisture content of the air mass, and the air/lake temperature difference are the primary controls on precipitation rate provided that the prevailing wind direction is favorable. There may be some advantage in initializing the model using the thermal wind.

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