Orographic Influences on the Mesoscale Structure of the 1998 Ice Storm

PAUL J. ROEBBER
Atmospheric Sciences Group, Department of Mathematical Sciences, University of Wisconsin at Milwaukee, Milwaukee, Wisconsin

JOHN R. GYAKUM
Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada

(Manuscript received 22 October 2001, in final form 31 May 2002)

ABSTRACT

The ice storm of 5–9 January 1998, affecting parts of the northeastern United States and the eastern Canadian provinces, was characterized by freezing rain amounts greater than 100 mm in some areas. The region of maximum precipitation occurred in a deformation zone between an anomalously cold surface anticyclone to the north and a surface trough axis extending from the Gulf of Mexico into the Great Lakes. Mesoscale processes were examined to understand their role in regulating the persistence, phase, and intensity of the event. The persistently cold near-surface air in the precipitating region was linked to orographic channeling of winds from the cold anticyclone to the north. The position of the surface-based freezing line was strongly tied to pressure-driven channeling of surface winds by orography during the first freezing rain episode (4–7 January), while this channeling contributed to the depth of the cold air north of the U.S. border and governed details of the position of the freezing line in the Lake Champlain valley during the second episode (7–10 January). For example, in the absence of the orographic channeling, model sensitivity simulations suggest that little or no freezing precipitation would have occurred at Burlington, Vermont (BTV), during the ice storm. A frontogenetical focus within the St. Lawrence, Ottawa, and Lake Champlain valleys was provided by orographic channeling of the cold air in combination with geostrophic southerlies in the warm air. The frontogenesis was an important contributor to the higher precipitation amounts during the ice storm, with model sensitivity estimates indicating that in the absence of the valleys, total freezing rain volumes would have been reduced by 12.1% and 16.5% in the first and second episodes, respectively. A discussion of the expected predictability of such events is provided.

1. Introduction

Gyakum and Roebber (2001; hereafter denoted GR1) analyzed the planetary-scale aspects of the freezing rain event of 5–9 January 1998, which devastated regions of northern New York and New England in the United States, and southern regions of the Canadian provinces of Quebec, Ontario, and New Brunswick. This meteorological event was unprecedented during the past two decades for this region of North America (DeGaetano 2000), both in terms of the quantity of the precipitation in that phase (e.g., 80–100 mm of freezing rain with accreted radial ice loads of 45 mm in Montreal, Quebec; Fig. 1a) and its persistence (5 days), and in terms of economic damage (approximately $4.4 billion, with $3 billion in Canada alone; NCDC 1999).

GR1 found that the ice storm consisted of two primary synoptic-scale cyclonic episodes (Figs. 1b and 1c), with the first (second) episode characterized by trajectories arriving in the precipitation zone that had been warmed and moistened by fluxes over the Gulf Stream current and the Gulf of Mexico (subtropical Atlantic Ocean). Based upon anomaly correlations of sea level pressure, 1000–925 hPa, and 1000–500-hPa thicknesses, GR1 found five analogs to this event in a 34-yr period. While four of these cases were characterized by extensive freezing rain, the best analog produced maximum precipitation amounts less than 50% of the 1998 event. GR1 found that a crucial distinguishing characteristic of the 1998 event was the prolonged contact of air parcels with the planetary boundary layer of the subtropical Atlantic Ocean, resulting in equivalent potential temperatures of approximately 330 K as the parcels traveled into the ice storm’s precipitation zone during the second episode.

The relative success in finding large-scale analogs to the ice storm, despite its unprecedented nature, suggests that the phase, intensity, and persistence of the event may also have been regulated by mesoscale processes. The focus of the accumulated precipitation within the valleys of the St. Lawrence and Ottawa Rivers during
Fig. 1. Precipitation analysis for: (a) the 6-day period ending at 1200 UTC 10 Jan 1998, with solid contour interval of 20 mm, beginning with 40 mm. Heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 1200 UTC 6 Jan 1998. Very heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 0000 UTC 9 Jan 1998. (b) The 3-day period ending at 1200 UTC 7 Jan 1998, with solid contour interval of 20 mm, beginning with 20 mm. Heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 0000 UTC 6 Jan 1998. Very heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 0000 UTC 7 Jan 1998. (c) The 3-day period ending at 1200 UTC 10 Jan 1998, with solid contour interval of 20 mm, beginning with 20 mm. Heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 1200 UTC 8 Jan 1998. Very heavy solid (dashed) contour is that of the 0°C isotherm at 850 hPa (surface) for 1200 UTC 9 Jan 1998. Locations of some sites mentioned in the text are indicated in (b).
the first episode [Figs. 1b, 2a; precipitation analysis based upon 433 cooperative, first-order synoptic and METAR (aviation routine weather report) stations] as compared to a rather broad axis of precipitation in excess of 50 mm, extending from south of Lake Ontario through northeastern New York state to northern New Hampshire, in the second episode (Fig. 1c), is suggestive that mesoscale details may be of some importance. We seek the answers to two questions: 1) what was the role of orography in maintaining the cold air supply and providing a frontogenetical focus? 2) what was the role of frontogenesis in modulating the intensity of the freezing precipitation during this event? These questions are motivated, in part, by related findings in the literature. For example, orographic channeling of cold air and topographically induced frontogenesis have been found to be contributers to both the phase and intensity of precipitation in events in the Pacific Northwest region (Ferber et al. 1993; Steenburgh et al. 1997; Doyle and Bond 2001).

The structure of the paper is as follows. Section 2 describes the datasets and methodologies used in this work. We document the synoptic and mesoscale aspects of the ice storm in section 3, including a discussion of the role of orography in regulating the surface winds during the event. In section 4, the influence of terrain
Table 1. Period model simulations and sensitivity tests. Start and end times are in UTC.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Start time</th>
<th>End time</th>
<th>Terrain resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>0000 UTC 4 Jan 1998</td>
<td>0000 UTC 6 Jan 1998</td>
<td>9 km</td>
</tr>
<tr>
<td>A2</td>
<td>0000 UTC 4 Jan 1998</td>
<td>0000 UTC 6 Jan 1998</td>
<td>111 km</td>
</tr>
<tr>
<td>B1</td>
<td>1200 UTC 5 Jan 1998</td>
<td>0000 UTC 7 Jan 1998</td>
<td>9 km</td>
</tr>
<tr>
<td>B2</td>
<td>1200 UTC 5 Jan 1998</td>
<td>0000 UTC 7 Jan 1998</td>
<td>111 km</td>
</tr>
<tr>
<td>C1</td>
<td>0000 UTC 7 Jan 1998</td>
<td>0000 UTC 9 Jan 1998</td>
<td>9 km</td>
</tr>
<tr>
<td>C2</td>
<td>0000 UTC 7 Jan 1998</td>
<td>0000 UTC 9 Jan 1998</td>
<td>111 km</td>
</tr>
<tr>
<td>D1</td>
<td>1200 UTC 8 Jan 1998</td>
<td>0000 UTC 10 Jan 1998</td>
<td>9 km</td>
</tr>
<tr>
<td>D2</td>
<td>1200 UTC 8 Jan 1998</td>
<td>0000 UTC 10 Jan 1998</td>
<td>111 km</td>
</tr>
</tbody>
</table>

on precipitation phase and its role in producing frontogenetic circulations that locally increase freezing rain is studied. A summary of the key findings is provided in section 5.

2. Observational and model data

a. Observations

Our study utilizes standard surface and rawinsonde observations and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) global reanalysis gridded data (Kalnay et al. 1996). These data were augmented with special rawinsondes from Ottawa, Ontario (YOW; Figs. 1b, 2a), taken at approximately 6-h intervals in conjunction with the Third Canadian Freezing Drizzle Experiment (CFDEIII; Isaac et al. 1998). We use daily precipitation data from Ontario, Quebec, and the Maritime Provinces of Canada, along with cooperative precipitation data from New York and the New England states. Remotely sensed data used include the National Weather Service Weather Surveillance Radar-1998.
b. Data assimilation

A diagnostic dataset, for the purpose of augmenting the available observations described above, was generated using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5). A doubly nested domain configuration, employing 18 (6) km grid spacing on the outer (inner) domain, was constructed. The outer domain extends from 23.5°N, 98.0°W (south of Brownsville, Texas) northeast to 52.2°N, 57.7°W in extreme northeastern Quebec. The inner domain ranges from western Pennsylvania northeast to the Gulf of St. Lawrence (Fig. 2a). Two-way interaction between the model domains was allowed, such that conditions in the inner domain feed back to the coarse domain, and vice versa, with matching at the nest boundary (see Zhang et al. 1986). Model initialization was cold start (i.e., no diabatic initialization) with the data supplied by the

Doppler (WSR-88D) radar data and the McGill wind profiler (Rogers et al. 1994).
NCEP reanalysis. These analyses, at 6 hourly intervals, also provided boundary conditions on the outermost grid domain throughout the course of the integrations. Vertical sigma levels are arranged such that the model output is available on a total of 23 levels (0.998, 0.991, 0.973, 0.945, 0.910, 0.870, 0.825, 0.775, 0.725, 0.675, 0.625, 0.575, 0.525, 0.475, 0.425, 0.375, 0.325, 0.275, 0.225, 0.175, 0.125, 0.075, 0.025), with a relatively large concentration at the lowest levels in order to resolve planetary boundary layer (PBL) structure. Precipitation processes were simulated in the outer domain using the Kain–Fritsch (1993) convective parameterization and an explicit moisture scheme that includes prognostic equations for cloud water, rainwater, ice, and supercooled water (Reisner et al. 1998). This latter scheme does not include graupel or riming processes, but does allow for time-dependent melting (freezing) of snow/ice (water) below (above) the freezing level. On the inner domain, only the explicit scheme was utilized. Radiative processes were computed using a cloud-radiation scheme in which diurnally varying shortwave and longwave radiative fluxes interact with explicit cloud and clear air, while the surface fluxes were used in the ground energy budget calculations (Dudhia 1989). The planetary boundary layer was modeled using the high-resolution Blackadar scheme (Zhang and Anthes 1982). Physiographic and land use patterns were back-interpolated from global topographic and land use da-
datasets (with resolution of approximately 19 km for the outer domain and 9 km for the inner domain; for the latter, see Fig. 2a) to the model grids.

To generate an analysis dataset that provides time continuity and dynamic coupling among the various model fields, a continuous four-dimensional data assimilation (FDDA) known as Newtonian relaxation is employed within MM5 (see Stauffer and Seaman 1990). The model solution was nudged toward three-dimensional gridded analyses of wind, temperature, and water vapor mixing ratio on the outer domain, while the inner domain was allowed to evolve freely within the constraints imposed by information flow from the outermost domain. This integration commences at 0000 UTC 3 January 1998 and ends at 0000 UTC 10 January 1998. Evaluation of this dataset in comparison to observations and analyses is provided in section 3b. All comparisons are conducted using data from the inner domain, which is hereafter referred to as the data assimilation run.

c. Terrain sensitivity simulations

As noted in section 2b, the data assimilation run was nudged towards the analysis in the large-scale domain. The data assimilation run cannot be used for sensitivity tests, since two-way interactions between the model domains are allowed and lateral boundary conditions for the inner domain are provided by the outer domain. Since the ice storm can be divided into two precipitation episodes, each of approximately 72 h in length, four
additional “control” simulations were made, along with four orographic sensitivity experiments, as summarized in Table 1 (note that while it would be possible to make two control simulations, each of 72-h duration, to cover the two precipitation periods, the periods were further subdivided to limit simulation drift). These eight simulations were made without nudging, for the reasons noted above.

Simulations A1, B1, C1, and D1 are the control runs, identical in every respect to the data assimilation run, except that no nudging is used and the simulation ranges are reduced as described. Simulations A1 and B1 extend for 48 and 36 h, respectively, and are designed to span the first ice storm precipitation episode, as shown in Fig. 1b. Similarly, C1 and D1 are designed to span the second ice storm precipitation episode, as shown in Fig. 1c. Simulations A2, B2, C2, and D2 are identical to A1, B1, C1, and D1, respectively, with the exception that the land use and topographic specifications are degraded from 9 km (finescale; Fig. 2a) to 111 km (coarse scale; Fig. 2b) resolution. This set of eight simulations will be analyzed below to directly assess the impact of orography on the details of the location, persistence, and intensity of the freezing precipitation during the period 4–10 January 1998.

A primary topographic feature of interest to this study, sharply defined in Fig. 2a and faintly evident in Fig. 2b, is the St. Lawrence River valley, running from Lake Ontario northeastward past Montreal (YUL; Figs. 1b, 2a) and Quebec City. South of Montreal is the Lake Champlain valley, which lies on the border of New York State and Vermont and leads to the Hudson River valley, which drains into the Atlantic near New York City. Another feature of importance to this study is the Ottawa River valley, which runs southeast to Ottawa and Montreal. These features are quite narrow and remain as only faint kinks in the elevation contours of Fig. 2b. Consequently, the impact of these features on the simulations can be addressed by direct comparison between the set of fine-scale terrain runs (A1, B1, C1 and D1) and the coarse-scale terrain runs (A2, B2, C2, and D2). This comparison is undertaken in section 4.

3. Analysis

a. Synoptic and mesoscale aspects

The large-scale setting for the ice storm was established by 4 January 1998, in which a deep Bermuda anticyclone transports moist air into New England and the eastern Great Lakes region (Figs. 3a,b). Precipitable water values of 20 mm, prevalent in the Ice Storm region, represent about 2.5 times that of the January climatological value. The warm, moist flow is shown by the surface observations (Fig. 4a) in which light to moderate rains are occurring in temperatures warmer than 5°C throughout the Lake Ontario region. The southwesterly surface winds, warm temperatures, and rain are evident at all four of the surface stations shown in the meteograms of Fig. 5a.

By 5 January 1998, a building, cold 1042-hPa anticyclone (Figs. 3c,d; Fig. 4b) became established over north-central Quebec, resulting in surface temperature falls of 15°C (30°C) across the upper St. Lawrence River valley (central Quebec). The winds, which had been 2.5–7.5 m s$^{-1}$ from the southwest, veered to the northeast in partial response to the channeling effect of the valley (see section 3c). In contrast, winds at Burlington, Vermont (BTV) switched from south-southwest to north-northwest during this time, again in an apparent reflection of the interaction of the cold outflow from the anticyclone with the channeling effects of topography (section 3c). By 0000 UTC 5 January at Ottawa, Ontario, and Montreal, Quebec, and a few hours later at Watertown, New York (ART), and BTV (Figs. 5a,b). This cold air was shallow in the valleys, however, as indicated by the height of the freezing level at YOW (~927 hPa) at 1200 UTC 5 January (Fig. 6) and the shallow layer of easterly winds (~200 to 400 m) at the McGill profiler site in downtown Montreal early on 5 January (Fig. 7a).

Later during 5 January, a weak approaching surface trough (Figs. 3d,f) led to a shift in the winds at BTV back to southerly (Figs. 4b,c; Fig. 5b) and the establishment of an axis of frontogenesis along the St. Lawrence River valley. Frozen or freezing precipitation commenced in the region at this time, although with continued southerly winds, the precipitation changed to rain at BTV by 1100 UTC 5 January (Fig. 5b). The surface winds continued from the northeast at YUL and YOW as the shallow pool of cold air remained entrenched in the valley (Figs. 6, 7a).

This pattern persisted until 1600 UTC 6 January, when the winds veered to southwest and then northwest at BTV (Fig. 5c), attended by falling temperatures as the weak surface trough passed (Figs. 3f,h; Figs. 4c,d). At this time, the continuous precipitation ended across the region, with totals ranging up to 50 mm (Fig. 1b). The localized regions of higher precipitation amounts in the St. Lawrence, Ottawa, and Champlain valleys are suggestive of the effects of frontogenesis arising through orographic interactions and will be further investigated below.

During 7 January, a sharp upper-level trough (delinedearedly by the analysis of potential temperature on the dynamic tropopause; Fig. 3h) was developing over the southwestern United States and a strong surface cyclone was forming over Texas and the Gulf of Mexico (Figs. 3g,h). Along and to the east of the surface trough axis, weaker tropospheric stability and stronger coupling with the dynamic tropopause was evident (as indicated by coupling index values of 10–
Fig. 5. Meteograms at (top) YOW, (top-middle) ART, (bottom-middle) YUL, and (bottom) BTW for: (a) 4, (b) 5, (c) 6, (d) 7, (e) 8, and (f) 9 Jan 1998. Plotted are surface temperature (°C), wind speed, and direction (same convention as in Fig. 4) and present weather. Because ART observations are not available for 9 Jan, Fort Drum, NY (GTB) is substituted in (f).

20 K, where the coupling index is defined as the potential temperature of the dynamic tropopause minus the maximum value of the equivalent potential temperature in the surface to 700-hPa layer; the use of the layer quantity is designed to account for the variable depth of the inversion layer; Fig. 3). At this time, the anticyclone was reinforced northeast of the ice storm region and the cold air deepened in the persistent northeasterly flow along the valley (Figs. 3g,h; Figs. 4d, 5d, and 6). Freezing precipitation renewed across the region with the approach of the cyclone (Figs. 4, 5). A plume of 20–30-mm precipitable water in association
with the transport of high equivalent potential temperature air from the subtropical Atlantic Ocean to the region (Figs. 3g,i,k; and GR1) favored heavy precipitation during this period, with totals from 7–10 January exceeding 100 mm in some areas and with extensive areas of 30–70 mm in the freezing rain zone (Fig. 1c). In contrast to the previous period, the winds persisted from the north at BTV and the freezing precipitation continued at BTV as well as the sites to the north during this time. By 1200 UTC 9 January, the surface-based inversion was intense, with the subfreezing air extending upwards to nearly 850 hPa, with a 10°C rise in temperature from 950 to 800 hPa (Fig. 6).

b. Data assimilation validation

As noted in section 2b, a diagnostic dataset, for the purpose of augmenting the available observations was generated using FDDA. This dataset will be used to assess details of orographic channeling of surface flows.
and frontogenesis in the ice storm region. In order to use these data for this purpose, it is necessary to establish that it adequately represents aspects of the ice storm that were captured by the observations. This verification is conducted in this section, using both qualitative and quantitative measures.

The MM5 data assimilation (Figs. 8, 9) captures the regional characteristics of the ice storm. The intensity and position of the cold anticyclone to the north of the precipitation region on 5 January is well represented, as is the ridge axis and northeasterly flow along the St. Lawrence River valley (Figs. 4b, 8a). A frontogenetic zone near ART appears in the data assimilation as a sharper band along the south shore of the St. Lawrence River, consistent with the higher resolution (compared to the analysis) of the assimilation data. By 0000 UTC 6 January, the observed southerlies in northern New England and northeasterlies along the St. Lawrence Riv-

The continued ridging and northeast flow along the St. Lawrence valley and progression of the cold air into the Lake Champlain region at 0000 UTC 7 January (Fig. 4d) is clearly represented in the data assimilation (Fig. 8c). The southward shift of the baroclinic zone accompanying the approach of the cyclone on 0000 UTC 8 January is also captured, as is the broad-scale southeasterly (northeasterly) flow in the warm (cold) air (Figs. 4e, 8d). The southward focus in the succeeding 24 h is also depicted, although the cold air as represented by the position of the 0°C isotherm is advanced too far equatorward (Figs. 4f, 8e). The evolution of the inversion structure, as indicated by the sounding section at YOW (Fig. 6), is captured by the data assimilation (Fig. 9). The representation of the increasing depth and intensity of the surface-based freezing layer and the overlying warm pool is robust.

Quantitative verifications of surface temperature, sea level pressure, and 500-hPa height from the data assimilation run were also conducted (Table 2). Every 12 h, for the period 0000 UTC 4 January to 0000 UTC 10 January 1998, all surface reports within the region depicted in Fig. 4 (comprising a total of 1013 observations) were compared with the data assimilation, for the same times and locations (the latter were interpolated from the 6-km grid to the station location using an inverse-distance Cressman scheme; Cressman 1959). For the same period, 500-hPa height data obtained from rawinsonde launches at eight sites within the ice storm region (Buffalo, New York; Maniwaki, Quebec; Ottawa, Ontario; Albany, New York; Chatham, Massachusetts; Gray, Maine; Caribou, Maine; Yarmouth, Nova Scotia) were compared with the data assimilation, for the same times and locations.

The results confirm that the FDDA successfully constrained error growth throughout the course of the 168-h simulation. Overall, although 500-hPa heights are biased low, root-mean-square (rms) error (18.5 m) is similar to the level of 12-h forecast error (Dimego et al. 1992). Under conditions in which large gradients in surface properties existed, the limited bias and low rms error in surface temperature and sea level pressure indicate a robust correspondence with observations.

Comparisons of precipitation produced by the data assimilation run with observed totals were also conducted (Table 3). Regional average precipitation from the data assimilation was comparable to observed totals, with some bias toward higher than observed amounts, most prominently in the second precipitation episode.
from 7–10 January. The spatial pattern of precipitation from the data assimilation captures key details of the observed pattern, such as amounts in excess of 40 mm north of Lake Ontario and in the Ottawa and St. Lawrence River valleys during the first episode (Fig. 1b; see Fig. 8b for a 3-hourly amount) and the axis of heavy precipitation (>50 mm) extending along a line from south of Lake Ontario through northeastern New York to northern New Hampshire during the second episode (Fig. 1c). The results as measured by threshold equitable threat scores (Rogers et al. 1995) are superior to those obtained from the operational Eta and MM5 for higher precipitation thresholds (e.g., Colle et al. 1999; Colle and Mass 2000), suggesting that some insight into the mechanisms that resulted in the magnitude as well as the distribution of precipitation can be obtained from these data.

c. Orographic channeling

Whiteman and Doran (1993) discuss a conceptual model of forcing mechanisms of valley winds. Under cold season, stably stratified conditions, one mechanism is of primary interest to the study of surface flows: pressure-driven channeling, in which the winds in the valley are driven by the component of the pressure gradient force aligned with the valley axis. Under these conditions, countercurrents (winds in the valley opposing the direction of the winds above the valley) are possible, since the winds in the free atmosphere will be approximately perpendicular to the pressure gradient force, and a component of that flow can align with the axis of the valley. Such flows have long been recognized as the primary determinant of the wind climatology in the St. Lawrence River valley (Powe 1968, 1969; Environment Canada 1987) and have been shown to be important to specific synoptic events in this region (Cohn et al. 1996). Such effects have been observed and successfully modeled in other broad valleys as well (e.g., Whiteman and Doran 1993; Gross and Wippermann 1987).

In order to understand the role of this process in producing the surface winds within the ice storm region, specific components of the momentum budget (frictional force, Coriolis force, and pressure gradient force) were computed as follows:

\[ F_x = -fv + g \frac{\partial z}{\partial x} \]  
\[ F_y = fu + g \frac{\partial z}{\partial y} \]

where the Lagrangian terms are neglected [as suggested by the small acceleration terms in Fig. 21 of Bell and Bosart (1988) and further emphasized by full momentum balance calculations for 0000 UTC 6 January 1998 (not shown)], and the frictional force \( F \) is determined as a residual. The calculations were made using the output from the data assimilation run for locations within the St. Lawrence (YUL, ART), Ottawa (YOW), and Champlain (BTV) valleys (Fig. 11). An a posteriori justification for using the data assimilation for this purpose is provided by the error statistics for sea level pressure and surface temperature cited in section 3b and the root-mean-square error of the along-valley component of the data assimilation winds at these four sites of \(-0.1 (3.2) \text{ m s}^{-1}\). Here, the term cold source will be used to refer to winds emanating from areas with colder surface temperatures than the affected downstream locations (denoted as an “easterly” flow at all sites in Fig. 11).

At all four locations, pressure-driven channeling was important, with the total wind being aligned with the pressure gradient force in 90% of the times examined (Fig. 11). This result is consistent with those of Bell and Bosart (1988), who found that pressure-driving was an important source of cold air during Appalachian cold air damming events. Similarly, Ferber et al. (1993) found that cold air channeled southward from an anti-

<table>
<thead>
<tr>
<th>Date</th>
<th>Surface temperature (°C)</th>
<th>Sea level pressure (hPa)</th>
<th>500-hPa height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0000 UTC 4 Jan 1998</td>
<td>-1.5 3.3</td>
<td>0.6 1.6</td>
<td>-16.4 23.6</td>
</tr>
<tr>
<td>1200 UTC 4 Jan 1998</td>
<td>0.7 3.8</td>
<td>-0.1 1.8</td>
<td>-7.2 9.1</td>
</tr>
<tr>
<td>0000 UTC 5 Jan 1998</td>
<td>2.0 3.6</td>
<td>-0.7 1.8</td>
<td>-15.1 16.6</td>
</tr>
<tr>
<td>1200 UTC 5 Jan 1998</td>
<td>1.6 3.0</td>
<td>-1.6 2.0</td>
<td>-15.6 18.3</td>
</tr>
<tr>
<td>0000 UTC 6 Jan 1998</td>
<td>0.0 2.3</td>
<td>-0.9 1.7</td>
<td>-20.1 21.9</td>
</tr>
<tr>
<td>1200 UTC 6 Jan 1998</td>
<td>-0.7 2.9</td>
<td>-0.8 1.8</td>
<td>-11.3 15.0</td>
</tr>
<tr>
<td>0000 UTC 7 Jan 1998</td>
<td>-0.8 3.1</td>
<td>-1.3 1.7</td>
<td>-7.7 11.1</td>
</tr>
<tr>
<td>1200 UTC 7 Jan 1998</td>
<td>-0.8 3.0</td>
<td>-0.4 1.8</td>
<td>-8.5 12.7</td>
</tr>
<tr>
<td>0000 UTC 8 Jan 1998</td>
<td>-1.4 2.8</td>
<td>-1.4 2.6</td>
<td>-11.3 22.8</td>
</tr>
<tr>
<td>1200 UTC 8 Jan 1998</td>
<td>-1.6 3.2</td>
<td>-0.5 2.4</td>
<td>-19.1 22.3</td>
</tr>
<tr>
<td>0000 UTC 9 Jan 1998</td>
<td>-2.0 4.7</td>
<td>-0.5 2.6</td>
<td>-25.6 26.1</td>
</tr>
<tr>
<td>1200 UTC 9 Jan 1998</td>
<td>-1.9 3.6</td>
<td>0.2 2.4</td>
<td>-10.2 14.1</td>
</tr>
<tr>
<td>0000 UTC 10 Jan 1998</td>
<td>-2.4 3.8</td>
<td>0.8 2.2</td>
<td>-3.0 15.8</td>
</tr>
<tr>
<td>0000 UTC 4–10 Jan 1998</td>
<td>-0.6 3.4</td>
<td>-0.5 2.0</td>
<td>-13.2 18.5</td>
</tr>
</tbody>
</table>
cyclone was important during heavy snow events in Seattle, Washington. Colle and Mass (1996) showed that pressure-driving was an important process in the production of the wind flow within the Strait of Juan de Fuca in northwest Washington State.

During the period 0000 UTC 4 January to 0000 UTC 10 January 1998, pressure-driven channeling was always in evidence at YUL and YOW, and included the production of countercurrents during the early portions of the storm (Figs. 6, 7). Both locations were dominated by cold flows (Fig. 11). At YUL, however, a component of the geostrophic flow was also along the valley for many times (e.g., 0000 UTC 9 January 1998) while at YOW, the geostrophic flow was usually cold source, with the pressure-driven channeling often opposed (e.g., 0000 UTC 8 January to 1200 UTC 9 January 1998). At BTV, the pressure-driving was usually cold source, whereas the geostrophic flow was often warm, leading to some intermittency in the source direction of the total wind. Finally, at ART, conditions were intermediate between YUL and BTV, with cold source pressure-driven channeling, but warm (cold) geostrophic flows during the first (second) stages of the ice storm. These results suggest that orographic effects may have been crucial.

Fig. 8. As in Fig. 4, but derived from the MM5 data assimilation run. Frontogenesis is shaded in range $20^\circ$–$370^\circ\text{C} (100 \text{ km})^{-1} (3 \text{ h})^{-1}$. Three-hourly model accumulated rainfall, valid at the time of the sectional, is shown by the heavy solid lines (0.2-, 1-, and 5-mm contours). Times are 0000 UTC (a) 5, (b) 6, (c) 7, (d) 8, and (e) 9 Jan 1998.
to the phase of the precipitation during the first episode at YUL, and during both episodes at BTV. At YOW, orographic effects may have been indirectly important by providing cold air upstream that could be advected into the region via the component of the geostrophic flow along the Ottawa River valley.

4. Terrain influences: Precipitation phase and frontogenesis

a. Finescale simulations and validation

Evidence was presented in section 3c that supports the idea that orographic effects were important to the phase of precipitation in the ice storm region. In addition, analysis of precipitation during this event, using information from both observations (section 3a) and the data assimilation run (section 3b), reveals a regional focus within a persistent frontogenetic zone. In this section, the specific contribution of the terrain to these aspects of the event is considered. This analysis is conducted primarily through comparison between the fine- and coarse-scale terrain simulations detailed in section 2c.

Prior to conducting such a comparison, however, it is necessary to document the validity of the finescale simulations as a baseline representation of the event. As for the data assimilation run, quantitative verifications of surface temperature, sea level pressure, and 500-hPa height were conducted. Every 12 h, all surface reports within the region depicted in Fig. 4 were compared with

Fig. 8. (Continued)
the finescale terrain simulations, for the same times and locations. For the same periods, 500-hPa height data obtained from rawinsonde launches at the eight sites within the ice storm region were also compared. The errors are comparable to those of the data assimilation at both the surface and 500 hPa (Table 4), with the exception of the last run (D1), where some slippage relative to the earlier simulations and the data assimilation is evident. Overall, however, these results indicate that the finescale terrain simulations provide an adequate representation of surface and upper-level features during the course of the ice storm.

A further verification is provided by comparing the mean depth, mean areal coverage, and volume of freezing rain within the ice storm region, obtained from the finescale terrain simulations, against estimates of those variables for the same time periods from observations (as represented by an objective analysis on a 0.5° latitude–longitude grid, using 433 observing locations; see Figs. 1b,c). Criteria were established to define freezing rain in a consistent way for both the model simulations and the observations. Freezing rain is defined at a model (analysis) grid point where precipitation is occurring if the lowest sigma-level (surface) temperature is less than or equal to 0°C and the 850-hPa temperature is greater than 0°C. These criteria were applied every 12 h, and provided that they were satisfied at each endpoint of that time, the precipitation for the 12-h interval was considered to be freezing (otherwise, the precipitation at the grid point was not counted). The comparison was made for two periods: 1200 UTC 4 January to 1200 UTC 7 January 1998 and 1200 UTC 7 January to 1200 UTC 10 January 1998. In order to construct this analysis from the model simulations, individual runs were concatenated. Specifically, the period 1200 UTC 4 January to 1200 UTC 7 January was constructed from the final
FIG. 10. Radar reflectivity for 0000 UTC 6 Jan 1998, composited from available NWS WSR-88D sites. Shading (see legend) denotes reflectivities of 5, 20, and 35 dBZ.

FIG. 11. Wind (filled arrow) and selected components of the momentum budget (open arrows, pressure gradient force, $P$; Coriolis force, orthogonal to the wind vector; friction force, obtained as the residual, $R$, for the period 0000 UTC 4–10 Jan 1998 for YUL, BTV, YOW, and ART. The scale is provided in the legend. All vectors are represented in a relative coordinate system with the axis of the valley transposed to an east-west orientation and the coldest air to the east (i.e., winds directed from Quebec City towards YUL, from YUL towards YOW, from YUL towards BTV, and from YUL towards ART are cold source and depicted as easterly). The angles used for these coordinate transformations are YUL 45°, BTV 90°, YOW 335°, and ART 50°.

The results of this analysis of the observations (finescale terrain simulations) are the following for 1200 UTC 4 January to 1200 UTC 7 January: mean depth of 29.6 (33.0) mm; mean area of 130 200 (75 100) km²; volume of $3856.5 \times 10^6 (2479.3 \times 10^6)$ m³. For 1200 UTC 7 January to 1200 UTC 10 January, the results from the observations (finescale terrain simulations) are: mean depth of 46.7 (53.5) mm; mean area of 139 840 (115 100) km²; volume of $6523.5 \times 10^6 (6675.9 \times 10^6)$ m³. The results indicate a model underestimate of freezing rain volume of 36% during the first episode, owing to considerably smaller areal coverage. This low bias suggests that comparisons between the fine- and coarse-scale terrain simulations will likely produce a conservative estimate of the impact of terrain on the production of freezing rain during this episode. During the second episode, the model performance is improved, showing a 2% overestimate of freezing rain volume.

b. Precipitation phase and frontogenesis

To assess the importance of terrain on the ice storm, the freezing rain analysis was repeated for the coarse-terrain simulations. For 1200 UTC 4 January to 1200 UTC 7 January, the coarse-scale terrain simulations resulted in 7.3% less mean depth, 5.2% less mean area

<table>
<thead>
<tr>
<th>Date</th>
<th>Mean (mm)</th>
<th>Rmse (mm)</th>
<th>Thresh</th>
<th>Bias</th>
<th>ETS</th>
</tr>
</thead>
<tbody>
<tr>
<td>4–7 Jan</td>
<td>21.2 (21.2)</td>
<td>12.2</td>
<td>15</td>
<td>1.09</td>
<td>0.33</td>
</tr>
<tr>
<td>7–10 Jan</td>
<td>63.2 (51.2)</td>
<td>25.6</td>
<td>50</td>
<td>1.28</td>
<td>0.31</td>
</tr>
<tr>
<td>4–10 Jan</td>
<td>84.4 (72.4)</td>
<td>28.5</td>
<td>50</td>
<td>1.10</td>
<td>0.26</td>
</tr>
</tbody>
</table>
and 12.1% less volume of freezing rain compared to the finescale terrain simulations. Similarly, for 1200 UTC 7 January to 1200 UTC 10 January, the coarse-scale terrain simulations resulted in 7.8% less mean depth, 9.5% less mean area and 16.5% less volume of freezing rain compared to the finescale terrain simulations. These results, although not conclusive in isolation, can be understood as follows. First, the increased areal coverage of freezing rain with terrain resolution occurs through the ability of the orography to anchor the position of the surface freezing line under conditions of synoptic-scale warm advection. Second, the increase in mean depth of freezing rain with terrain resolution is suggestive of a frontogenetic focus within a synoptic-scale pattern of precipitation. These ideas are supported by Doyle and Bond (2001), who found observational and model evidence of topographically induced frontogenesis and enhanced precipitation in association with cold outflow from the Strait of Juan de Fuca. Further, Steenburgh et al. (1997) presented composite and climatological evidence of the importance of cold gap flows to precipitation in the Washington Cascades. Additional evidence supporting these points is provided below.

At 0000 UTC 5 January, continuous precipitation occurred at ART and BTV, and began at both YOW and YUL within the next 6 h (Fig. 5). During 0000 UTC 5 January to 0000 UTC 6 January, the surface freezing line advanced northward markedly in response to the competing effects of geostrophic southerlies affecting the region (Figs. 6, 7). St. Lawrence River valley (Figs. 11, 13b), despite the absence of the orographic channeling, the sensitivity simulations suggest that little or no freezing precipitation would have occurred at BTV during the ice storm (Figs. 13b–e). These same effects persist and amplify through 9 January (Figs. 13d,e), except in the Lake Champlain valley, where the pressure-driven forcing is opposed by the geostrophic flow (Fig. 11). Owing to the depth of the cold air during this time (Figs. 6, 7), the position of the surface freezing line, even in the absence of orographic influences, would likely have been south of YOW and YUL (Fig. 13e). In contrast, pressure-driven channeling at BTV (combined with the geostrophic flow at ART), likely led to the maintenance of freezing precipitation south of the U.S.–Canada border during the second episode of the ice storm (Figs. 11, 13e). Hence, the interaction of orography with the prevailing synoptic conditions was critical to the phase of the precipitation in the ice storm region throughout the event (e.g., in the absence of the orographic channeling, the sensitivity simulations suggest that little or no freezing precipitation would have occurred at BTV during the ice storm (Figs. 13b–e)).

The role of orography in providing a frontogenetical focus and the frontogenesis in modulating the intensity of the precipitation during the first and second periods of the ice storm requires further investigation. During 4–7 January, locally higher precipitation amounts in the Ottawa and St. Lawrence River valleys were observed (Fig. 1b). This region was within the baroclinic zone established with the approach of a trough (Fig. 3) and frontogenesis was occurring within the ice storm precipitation zone (Figs. 4b,c). This frontogenesis is most apparent in the time sections at YUL and BTV on 5–6 January (Figs. 5b,c), which show the persistent northeast (southerly) flow across the baroclinic zone. The effects of this process are evident in the data assimilation and radar observations, which show a band of precipitation across the frontogenetic zone during this time (Figs. 8b, 10). Orography played a key role in providing a frontogenetical focus: northerly winds across the baroclinic zone result from the channeling effects of the St. Lawrence River valley (Figs. 11, 13b), despite the geostrophic southerlies affecting the region (Figs. 6, 7). The effect was to intensify the baroclinic zone and to promote a frontogenetical circulation (Fig. 12). As noted above, the finescale terrain leads to an increase in the

<table>
<thead>
<tr>
<th>Run (Date)</th>
<th>Surface temperature (°C)</th>
<th>Sea level pressure (hPa)</th>
<th>500-hPa height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ME Rmse</td>
<td>ME Rmse</td>
<td>ME Rmse</td>
</tr>
<tr>
<td>A1 (0000 UTC 4–6 Jan 1998)</td>
<td>–0.1 3.6</td>
<td>0.4 1.7</td>
<td>–12.8 16.3</td>
</tr>
<tr>
<td>B1 (1200 UTC 5 Jan to 0000 UTC 7 Jan 1998)</td>
<td>–1.5 3.5</td>
<td>–0.3 2.0</td>
<td>–12.8 16.3</td>
</tr>
<tr>
<td>C1 (0000 UTC 7–9 Jan 1998)</td>
<td>–1.1 3.3</td>
<td>0.1 2.5</td>
<td>–13.5 19.1</td>
</tr>
<tr>
<td>D1 (1200 UTC 8 to 0000 UTC 10 Jan 1998)</td>
<td>0.8 4.9</td>
<td>–0.1 2.7</td>
<td>–14.4 22.5</td>
</tr>
</tbody>
</table>

Table 4. Verification statistics for surface temperature (°C), sea level pressure (hPa) and 500-hPa height (m) from the finescale terrain simulations (A1, B1, C1, and D1).
volume of freezing precipitation, not only by increasing the areal coverage (through anchoring of the surface freezing line; Figs. 14a,b), but also by increasing the mean depth of the precipitation. Analysis of differences in 24-h accumulated precipitation associated with terrain effects, reveals increased precipitation (relative to that obtained with the coarse-scale orography) in the St. Lawrence River valley on 5–6 January (Fig. 14a) and the Ottawa River and Lake Champlain valleys on 5–7 January (Figs. 14a,b).

The second period (8–10 January) is characterized by a widespread region of precipitation totals in excess of 50 mm (Fig. 1c), suggesting a less prominent role for frontogenesis during this time. The baroclinic zone remains positioned across the ice storm region, but in this case, it is in association with a well-developed cyclone (Fig. 3). Surface sectionals make it clear that the primary axis of frontogenesis is shifted southward relative to the first period (Figs. 4e,f), in alignment with the axis of highest precipitation (Fig. 1c), but largely south of the freezing line. Time sections support this view, showing northerly winds sweeping across the region (Figs. 5e,f) as the cold air deepens at YOW (Fig. 6) and YUL (Fig. 7). However, localized frontogenesis and enhancement of precipitation are suggested by the data assimilation in the eastern townships of Quebec and along the St.
Fig. 13. MM5 model difference fields (finescale minus coarse-scale terrain run) for lowest sigma level (0.998), vector wind (arrows, 5 m s⁻¹ reference scale in lower left), and potential temperature [solid (dashed) positive (negative), 1°C interval]. Also shown in the finescale (9-km resolution) topography (shaded, with scale in m indicated on the left) and the position of the -1, 0, and +1°C lowest sigma-level temperature (heavy solid, finescale terrain; heavy dashed, coarse-scale terrain). Valid times are 0000 UTC: (a) 5 Jan (24-h simulations from A1 and A2), (b) 6 Jan (12-h simulations from B1 and B2), (c) 7 Jan (36-h simulations from B1 and B2), (d) 8 Jan (24-h simulations from C1 and C2), and (e) 9 Jan 1998 (12-h simulations from D1 and D2).

The ice storm of 5–9 January 1998 produced more than 50 mm of freezing rain in a widespread area within the northeastern United States and the eastern Canadian provinces with resultant damage in excess of $4 billion.
FIG. 13. (Continued)
In this case, the difficult problem of quantitative precipitation forecasting (location, timing, and amount) was complicated by issues of persistence (the event continued over 5 days) and precipitation phase. This storm was examined with a view towards understanding the role of mesoscale processes in regulating the persistence, phase, and intensity of the event. With respect to persistence, the operational numerical model guidance (Eta Model) was consistent in its depiction of the northward displacement with time of the freezing line from 5–10 January. In section 3c, it was established that the persistence of the surface-based cold air was tied to orographic channeling of the cold air. The persistence of the cold air within the St. Lawrence and Ottawa River valleys also supplied a frontogenetical focus and modulated the intensity of the precipitation within the ice storm region. As a result, while operational forecasts of this event correctly identified the risk of freezing rain, the magnitude of the event was not anticipated.

However, owing to the presence of large-scale factors that provide an environment in which such cases frequently develop (e.g., GR1) and a mesoscale focus that is strongly linked to in situ forcing from orography, we speculate that the intrinsic predictability of such events may be relatively high. Since the effectiveness of the orographic channeling is controlled by the synoptic-scale pressure field, an important sensitivity of this forecast is reflected in the position and magnitude of the cyclone–anticyclone couplet. For example, as the cold anticyclone progresses from a position north to south of the valley axis, the component of the pressure gradient force along the valley passes from zero to a local
maximum and back to zero again. Such alterations would have considerable impact on the persistence of the cold air, the associated frontogenetical zones, and the persistence, phase, and magnitude of the resultant precipitation. Hence, any intrinsic predictability at the mesoscale can only be accessed through sufficient prediction at the larger scales. These issues will be explored in detail in a subsequent paper.

The work reported herein revealed the following major points:

- The persistence of the cold air in the precipitating region was linked to orographic channeling of cold air from an anticyclone positioned to the north and east of the affected areas.
- The position of the surface-based freezing line was strongly tied to orographic channeling during the first freezing rain episode. Orographic channeling contributed to the depth of the cold air north of the U.S. border and governed details of the position of the freezing line in the Lake Champlain valley during the second episode. For example, in the absence of the orographic channeling, model sensitivity simulations suggest that little or no freezing precipitation would have occurred at Burlington, Vermont (BTV), during the ice storm.
- Frontogenetical focus within the St. Lawrence, Ottawa, and Lake Champlain valleys was provided by orographic channeling of the cold air in combination with geostrophic southerlies in the warm air. The frontogenesis was an important contributor to higher freezing precipitation amounts during the ice storm, with model sensitivity estimates indicating that in the absence of the valleys, total freezing rain volumes would have been reduced by 12.1% and 16.5% in the first and second episodes, respectively.

Acknowledgments. The authors would like to thank Mr. Ron McTaggart-Cowan of McGill University for providing the NCEP reanalysis dynamic tropopause and lower tropospheric equivalent potential temperature data, used to construct Fig. 3. Professor Frederic Fabry (McGill) assisted with the processing of the wind profiler data. Dr. George Isaac and Walter Strapp kindly provided the special Ottawa sounding data from the CFDE III. Mr. William Richards (Environment Canada) assisted with the processing of the Canadian daily precipitation data. Mr. Paul Sisson of the Burlington National Weather Service Forecast Office provided the Northeastern United States Cooperative Precipitation data. The authors acknowledge the assistance of the COMET staff, particularly Elizabeth Page and Tim Albert, for processing and furnishing the operational and MM5 grids in GEMPAK format, and for providing surface station, rawinsonde, and 8BD radar data. The paper benefited from the comments of the three anonymous reviewers.

REFERENCES


Powe, N. N., 1968: The influence of a broad valley on the surface winds within the valley. Canadian Department of Transport, Meteorology Bureau, Rep. TEC-668, 9 pp. [Available from the Canadian Government Publishing Centre, Supply and Services Canada; Ottawa, ON Canada K1A 0S9.]

——, 1969: The climate of Montreal. Canadian Department of Transport, Meteorology Bureau, Climatological Study 15, 51 pp. [Available from the Canadian Government Publishing Centre, Supply and Services Canada; Ottawa, ON, Canada K1A 0S9.]


